



The effects of resolving the Canadian Arctic Archipelago in a finite element sea ice model

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ABSTRACT

Though narrow straits may have a strong influence on the large-scale sea ice mass balance, they are often crudely represented in coarse resolution sea ice models. Unstructured meshes, with their natural ability to fit boundaries and locally increase the mesh resolution, propose an alternative framework to capture the complex oceanic areas formed by coasts and islands. In this paper, we develop a finite element sea ice model to investigate the sensitivity of the Arctic sea ice cover features to the resolution of the narrow straits constituting the Canadian Arctic Archipelago. The model is a two-level dynamic-thermodynamic sea ice model, including a viscous-plastic rheology. It is run over 1979–2005, forced by daily NCEP/NCAR reanalysis data. Confronting qualitatively numerical experiments with observations shows a good agreement with satellite and buoys measurements. Due to its simple representation of the oceanic interactions, the model overestimates the sea ice extent during winter in the southernmost parts of the Arctic, while the Baffin Bay and Kara Sea remain ice-covered during summer. In order to isolate the benefits from resolving the Canadian Arctic Archipelago, a numerical experiment is performed where we artificially close the archipelago. Focusing on the large-scale sea ice thickness pattern, no significant change is found in our model, except in the close surroundings of the archipelago. However, the local and short-term influences of the ice exchanges are nonnegligible. In particular, we show that the ice volume associated to the Canadian Arctic Archipelago represents 10% of the Northern Hemisphere sea ice volume and that the annual mean ice export towards Baffin Bay amounts to $125 \text{ km}^3 \text{ yr}^{-1}$, which may play an important role on the convective overturning in the Labrador Sea.

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1. Introduction

Sea ice is a key component of the high latitudes climate (e.g., Serreze et al., 2007). In regions where the incoming solar radiation at the top of the atmosphere is higher during local summer than elsewhere on the Earth (e.g., Wallace and Hobbs, 1977), most of the downwelling shortwave radiation is reflected due to the high snow and ice albedos (e.g., Perovich et al., 2002). Moreover, the salt rejected into the underlying water during sea ice formation takes an important part in high-latitude oceanic convection and thus in the global thermohaline circulation (e.g., Aagaard and Carmack, 1989). Sea ice further constitutes a natural barrier between the atmosphere and ocean, insulating the relatively warm ocean waters from the cold atmosphere by drastically limiting mutual heat, momentum and water mass exchanges. All these interactions

illustrate the complex role played by sea ice in the climate system. In particular, the ice-albedo feedback (Ebert and Curry, 1993) is considered to be largely responsible for the high sea ice sensitivity to climate change, as highlighted by simulations with climate general circulation models (CGCMs) on the one hand (e.g., Manabe and Stouffer, 1980), and by the recent Arctic sea ice minima of extent on the other hand (Stroeve et al., 2005). Nevertheless, recent comparisons (e.g., Holland and Bitz, 2003; Arzel et al., 2006; Zhang and Walsh, 2006; Lefebvre and Goosse, 2008) between CGCM simulations plead for the need to improve our understanding of sea ice and its representation in climate models.

Sea ice is drifting at the surface of the ocean, mainly driven by the wind and the oceanic currents. The large-scale Arctic ice circulation is characterized by two main features. In the western part of the Arctic, the ice describes a large anticyclonic gyre named the Beaufort Gyre where the ice can be trapped for up to 10 years (Hibler, 1980), with typical ice velocities of 2 cm s^{-1} . In the east, the ice is transported by the Transpolar Drift from the north of Siberia to the North Atlantic through Fram Strait. This strait is approximately 435 km wide and constitutes the main exit for the Arctic sea ice, the ice volume export reaching about $2200 \text{ km}^3 \text{ yr}^{-1}$

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(Kwok et al., 2004). Narrow straits, on their part, may have a strong impact on the geophysical flows in general (e.g. water mass exchanges between different oceanic basins, complex flow in localized areas, etc.) and, in the polar regions, they can be responsible for a somewhat chaotic dynamical behavior of sea ice. Due to the high ice concentration and low ice temperature, the formation of ice bridges in channels like Nares Strait (Samelson et al., 2006) may prevent any ice flow for several months. Later, the ice is suddenly released when the plugs collapse. Moreover, the ice characteristics in these channels have a strong interannual variability (Melling, 2002; Kwok, 2006).

Of particular concern is the complex area formed by the numerous islands and coastlines in the Canadian Arctic Archipelago (CAA, see Fig. 1). The straits constitute an important pathway for the cold waters from the Arctic Ocean to the Atlantic. Even if many recent studies have explored the freshwater cycle of the Arctic (e.g., Stigebrandt, 2000; Serreze et al., 2006), large uncertainties remain, especially concerning the sea ice contribution. Assembling different sources, we get the following picture: estimates of the ice freshwater outflow through the southern part of the CAA based on observations in Lancaster Sound (Prinsenbergh and Hamilton, 2005) roughly amount to $160 \text{ km}^3 \text{ yr}^{-1}$ (Serreze et al., 2006), while, in the north, the Nares Strait sea ice flux (Kwok et al., 2004) accounts for $\sim 100 \text{ km}^3 \text{ yr}^{-1}$ of freshwater (assuming a sea ice salinity of 4 psu, for a reference seawater salinity of 34.8 psu, and a density of 900 kg m^{-3} , see Serreze et al., 2006). In comparison, the oceanic freshwater flux through the CAA is evaluated by Prinsenbergh and Hamilton (2005) at $3200 \text{ km}^3 \text{ yr}^{-1}$. The rest of the northern gates of the CAA contributes to a net inflow into the Arctic (Kwok, 2006), suggesting, as stated in the latter study, that most of the ice exported into Baffin Bay (except from Nares Strait) is produced in the CAA rather than exported from the Arctic. Because sea ice is landfast during a large part of the year in the northern CAA (Melling, 2002), the different straits act as a “buffer” on sea

ice between the Arctic Ocean and Labrador Sea, preventing Arctic ice outflow in the north and providing ice generation in the south (Sou, 2007).

The role of the freshwater flux through the CAA is thought to be important regarding the freshwater budget of the region (Stigebrandt, 2000; Dickson et al., 2007), and may play an important role in the Labrador Sea convective overturning (Goosse et al., 1997). Less known is the influence of the ice contribution to this freshwater export, on the one hand, and the impact of the ice flux on the water stratification in Baffin Bay, on the other hand. Together with the North Water Polynya formation (northern Baffin Bay, see Barber et al., 2001), the latter impact might constitute a part of the explanation why most of the models miss the decay of the summer sea ice cover in Baffin Bay (e.g., Timmermann et al., 2005; Johnson et al., 2007; Vancoppenolle et al., submitted for publication). Finally, data in this harsh region of the world are scarce (e.g., Melling, 2002), especially concerning the ice thickness, ranging between 2 and 6 m in the CAA (Bourke and Garrett, 1987; Melling, 2002; Kwok, 2005), underlining the need for numerical experiments. Furthermore, this area is currently being of strong economical importance and political interest in the context of the opening of the Northwest Passage.

Finite element methods have been proposed at the onset of sea ice modeling. In the framework of the AIDJEX project (Arctic Ice Dynamics Joint Experiment), Mukherji (1973) was the first to make use of a finite element method code for its aptitude to simulate the crack propagation in sea ice. A few years later, (Becker, 1976) introduced the guiding lines of the method and asserted: “Because of their generality and widespread use, finite element techniques seem a worthwhile alternative to the difference scheme. [...] The ease with which finite element techniques can be used to model complicated shapes with arbitrary variation in the mesh spacing is likely to be the motivating factor leading to any such use”. In the early-80s, Sodhi and Hibler (1980) used in a pioneer study an

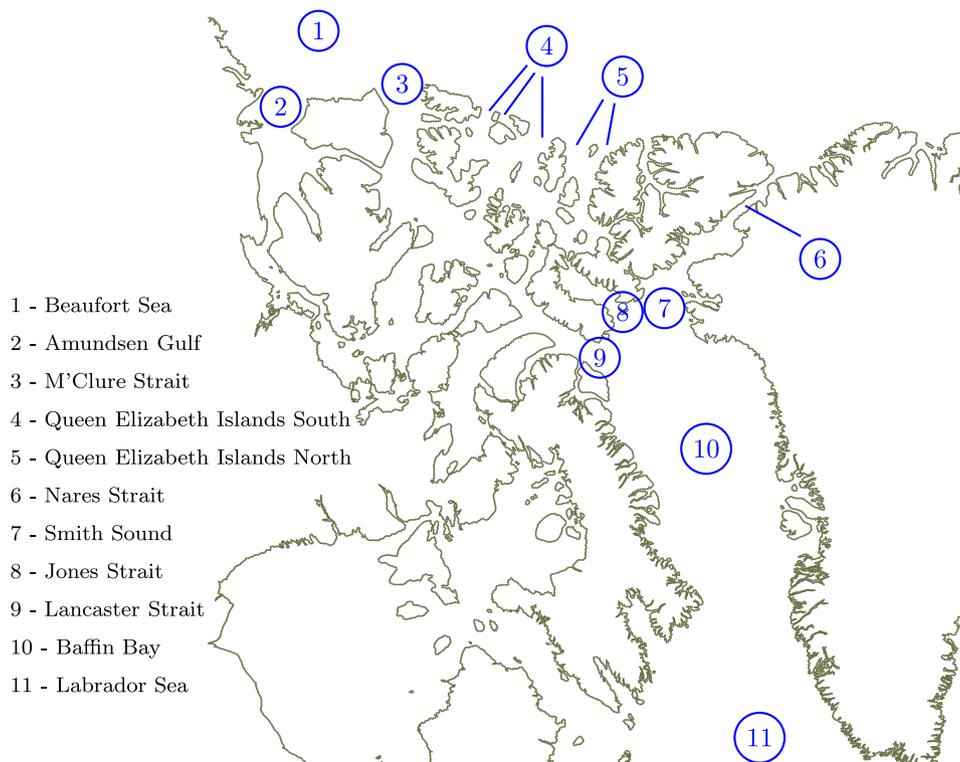


Fig. 1. The Canadian Arctic Archipelago is composed of numerous islands and coastlines. All the geographical places of interest for this study are clearly labeled on the map. Most of them correspond to the gates for the computation of ice area/volume fluxes.

unstructured grid along with a finite element method in order to resolve the ice drift in the complex region of Strait of Belle Isle. Another work investigating the finite element method in sea ice modeling was held by Thomson et al. (1988). They performed a comprehensive comparison between different constitutive laws for Eulerian and Lagrangian descriptions in order to model the short-term ice motion in Beaufort Sea.

Unstructured grids have among others the advantage to avoid the pole singularity and related stability problems (Williamson, 1979) and to represent in a more accurate fashion coastlines (e.g., Adcroft and Marshall, 1998; Legrand et al., 2006). Recently, many efforts have been brought to investigate the contribution of unstructured meshes in ocean modeling (e.g., Walters, 1992; Lynch et al., 1996; Le Roux et al., 2000; Danilov et al., 2004; Pietrzak et al., 2005; White et al., 2008a) and in sea ice modeling (Schulkes et al., 1998; Yakovlev, 2003; Wang and Ikeda, 2004; Sulsky et al., 2007). Very few modeling research has been performed on the CAA. Two studies are based on unstructured grid regional models: Kliem and Greenberg (2003) constructed gridded fields of potential temperature and salinity, used in a second step to simulate the mean circulation in the area, while Sou (2007) studied the impact of future climate scenarios on the CAA. The present paper in turn presents a large-scale sea ice model operational for climate studies which, to the authors' knowledge, is the first to investigate the effects of resolving the CAA on the ice cover features.

The model presented is aimed to be coupled to the finite element ocean model SLIM¹ (White et al., 2008b). As both models are still under development, and because a large part of the variability of the ice circulation can be explained by the winds (Thorndike and Colony, 1982; Serreze et al., 1989), the study presented here is conducted with an uncoupled sea ice model with no ocean dynamics. Furthermore, the ocean stress is not significant in most of the northern CAA due to the weak ocean currents (Stronach et al., 1987). Though tides are strong in other parts of the CAA (e.g., Hell Gate, Nares strait, Gulf of Boothia), they are not represented in this model. Since most of the results are considered on a climatological time scale, the effects of tides should not alter significantly the results. The sea ice model includes a viscous-plastic rheology and the Semtner (1976) zero-layer thermodynamic scheme. It is evident that some caution must be taken while analyzing the model results because of the simple thermodynamics used and the absence of ocean currents. The model complexity is nevertheless found sufficient for the purpose of this paper is to focus (i) on the feasibility and advantages of unstructured grids in large-scale sea ice modeling, and (ii) on the overall influence of the CAA on the Arctic ice cover and freshwater budget.

This paper is organized as follows. Section 2 describes the sea ice model, while Section 3 deals with its numerical resolution and focuses on its finite element implementation. In Section 4, the model results are first confronted to the observations in order to validate the model. The influence of the CAA resolution on the results is then investigated via sensitivity experiments. Concluding remarks are finally given in Section 5.

2. Model description

Following Hibler (1979), sea ice is assumed to behave dynamically as a viscous-plastic fluid with nonlinear viscosities. The model includes two thickness categories (see Fig. 2): thick ice with an actual thickness h_r and thin ice corresponding to a thickness lower than a cutoff constant h_0 . In this so-called two-level model, we introduce two variables to describe sea ice on a given oceanic area: the mean ice thickness h and the ice concentration A , which is de-

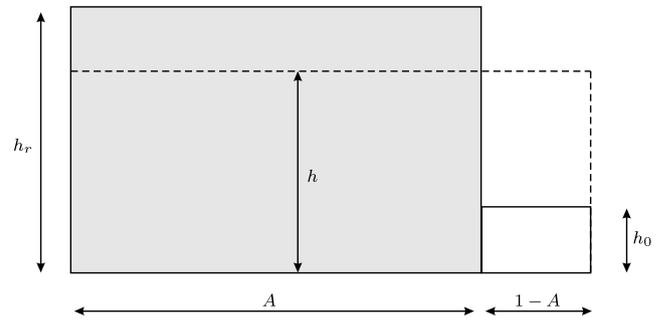


Fig. 2. Two-level model: neglecting h_0 leads to a simplified relation between mean (h) and actual (h_r) ice thicknesses: $h = Ah_r$.

fin as the proportion of thick ice recovering the oceanic area. The relation between those variables is illustrated in Fig. 2.

According to the aforementioned definitions, the equation of momentum conservation reads:

$$\rho h \frac{D\mathbf{u}}{Dt} = -\rho h f \mathbf{k} \times \mathbf{u} + A(\boldsymbol{\tau}_a + \boldsymbol{\tau}_w) - \rho h g \nabla \zeta + \nabla \cdot \boldsymbol{\sigma}, \quad (1)$$

where ρh is the ice mass per unit area, \mathbf{u} is the ice velocity, $\frac{D}{Dt} = \frac{\partial}{\partial t} + \mathbf{u} \cdot \nabla$ is the material derivative, f is the Coriolis parameter (taken to be constant for the high latitudes), \mathbf{k} is a unit vector normal to the surface, $\boldsymbol{\tau}_a$ and $\boldsymbol{\tau}_w$ are the air and ocean surface stresses on the ice weighted by the ice concentration, respectively (Connolley et al., 2004), g is the acceleration due to gravity, ζ is the sea surface dynamic height and $\boldsymbol{\sigma}$ is the ice internal stress tensor. Both the air and water stresses are parameterized by quadratic drag laws, the water stress depending on the ice velocity:

$$\boldsymbol{\tau}_w = \rho_w C_w |\mathbf{U}_w - \mathbf{u}| [(\mathbf{U}_w - \mathbf{u}) \cos \theta_w + \mathbf{k} \times (\mathbf{U}_w - \mathbf{u}) \sin \theta_w], \quad (2)$$

where C_w is the water drag coefficient, ρ_w the water density and θ_w the water turning angle.

The force due to the ocean tilt is commonly computed by establishing the geostrophic balance with the ocean current:

$$-\rho h g \nabla \zeta = \rho h f \mathbf{k} \times \mathbf{U}_w. \quad (3)$$

In our model, the ocean velocity \mathbf{U}_w is uniformly equal to zero, hence reducing the water surface stress to a simple drag of the ocean on the ice. Additionally, the contribution of the ocean tilt is neglected. We shall nevertheless take these terms into account in the rest of this paper for the sake of generality.

The viscous-plastic rheology assumes that sea ice is behaving as a nonlinear compressible viscous fluid with the general form:

$$\boldsymbol{\sigma} = 2\eta(\dot{\epsilon}, P)\dot{\epsilon} + (\zeta(\dot{\epsilon}, P) - \eta(\dot{\epsilon}, P))\text{tr}(\dot{\epsilon})\mathbf{I} - \frac{P}{2}\mathbf{I}, \quad (4)$$

where $\dot{\epsilon} = \frac{1}{2}(\nabla \mathbf{u} + \nabla \mathbf{u}^T)$ is the strain rate tensor, \mathbf{I} is the identity tensor, η and ζ are the shear and bulk viscosities, respectively, and P is the ice strength. For typical strain rates, ice is assumed to behave plastically, while for small strain rates, ice is supposed to flow very slowly in a viscous way. In the one-dimensional case, a fixed strain rate $\dot{\epsilon}_0$ defines the boundary between the plastic and the viscous regimes. In two dimensions, this threshold concept is generalized through a yield curve in the principal stress plane. If the stress state lies inside this curve, ice behaves viscously; otherwise, ice is constrained to plasticity and the state lies on the curve. Hibler's (1979) original elliptical yield curve has been widely used, and, combined with the normal flow rule, it yields the following viscosities:

$$\zeta = \min\left(\frac{P}{2\gamma}, 2.5 \times 10^8 P\right) \quad \text{and} \quad \eta = \zeta/e^2, \quad (5)$$

¹ Second-generation Louvain-la-Neuve Ice-ocean Model, <http://www.climate.be/SLIM>.

where e is the ellipse eccentricity (typically 2), P determines the ellipse size and γ is given by:

$$\gamma = \left[(\dot{\epsilon}_{xx}^2 + \dot{\epsilon}_{yy}^2)(1 + 1/e^2) + 4\dot{\epsilon}_{xy}^2/e^2 + 2\dot{\epsilon}_{xx}\dot{\epsilon}_{yy}(1 - 1/e^2) \right]^{1/2} \quad (6)$$

and can be interpreted as a measure of the strain rate determining whether the material undergoes plastic deformation or not. In the principal stress plane, all the stress states characterized by the same γ value lie on an ellipse centered on $(-1/2, -1/2)$ whose size normalized by P is proportional to γ (Fig. 3). Finally, we need an expression of the pressure term to close the system. Following Hibler (1979), P is computed as:

$$P = p^* h e^{-C(1-A)} \quad (7)$$

with two empirical parameters p^* and C .

Mass conservation is ensured by an advection equation for the mean ice thickness, while another continuity equation is required for the ice concentration:

$$\frac{\partial h}{\partial t} + \nabla \cdot (\mathbf{u}h) = S_h, \quad (8)$$

$$\frac{\partial A}{\partial t} + \nabla \cdot (\mathbf{u}A) = S_A, \quad (9)$$

where S_h and S_A are source terms accounting for sea ice ablation or accretion. These terms are computed thanks to three heat budgets: two at the top and the bottom of the thick ice and one over the thin ice/open water area. This explains why the actual ice thickness h_r will be used as variable when computing the thermodynamical processes. In addition, the ice concentration is constrained ($A \leq 1$) to roughly account for the mechanical redistribution of sea ice. Inspired by Thomson et al. (1988), the redistributed values of the actual ice thickness h'_r and concentration A' are given by:

$$h'_r = h_r + h_r[A - 1]^+, \quad (10)$$

$$A' = A - [A - 1]^+, \quad (11)$$

where $[f]^+$ is the positive part of f , i.e. $\max(f, 0)$. This formulation clearly shows a source term for h_r and a sink term for A : the idea here is to prevent the ice concentration to exceed unity while the actual ice thickness is allowed to increase.

The computation of heat diffusion within the thick ice is based on Semtner's (1976) zero-layer model. This model neglects the storage of sensible and latent heat, resulting in a linear temperature profile in the ice. Following Fichfet et al. (1998), two boundary conditions are needed, expressed as a heat budget at the surface (giving the pos-

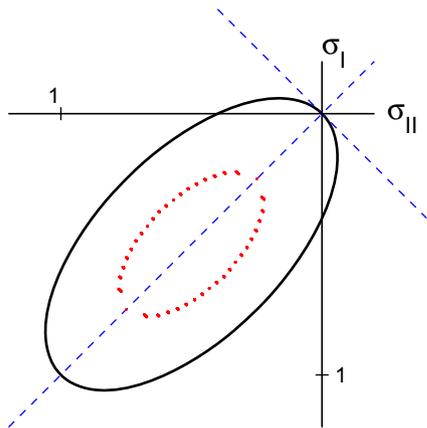


Fig. 3. Principal stress diagram of the constitutive relationship. Inside the ellipse, the material is considered as viscous. On the ellipse itself, the material undergoes plastic deformation, keeping a constant stress even if the deformation rate is higher than the limit defined by the geometrical locus ($\gamma = 2 \times 10^{-9} \text{ s}^{-1}$). Material points yielding a $\gamma = 10^{-9} \text{ s}^{-1}$ are also plotted in red. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this paper.)

sible ablation rate S_{hr}^{su} of the actual ice thickness) and at the bottom of the ice (S_{hr}^{b}). At the upper surface, an equilibrium surface temperature T_s is computed from the heat balance:

$$(1 - \alpha)F_r + F_L + F_s + F_l + F_c = 0, \quad (12)$$

where α is the surface albedo, which in our model is prescribed for each month and takes the presence of snow into account. F_A is the net atmospheric flux to the upper ice surface and F_C is the conductive flux through ice. Fluxes directed to the ice surface are taken to be positive. In order to solve the heat balance equation, the model includes a parameterization of the solar radiation F_r (Zillmann, 1972), the net longwave radiation F_L (Berliand and Berliand, 1952) and turbulent fluxes of sensible F_s and latent F_l heat (classical bulk formulae, see Goosse, 1997). If the predicted temperature T_s is above the melting point, its value is reduced to that point, yielding an excess of energy counterbalanced by latent heat absorption:

$$S_{hr}^{\text{su}} = -\frac{F_A + F_C}{L_i}, \quad (13)$$

with L_i the volumetric latent heat of fusion of ice.

At the base of the ice, the bottom temperature is kept at the freezing point of seawater (T_f). The melting or growth rate of the ice depends only on the conductive heat flux and the oceanic heat flux F_b :

$$S_{hr}^{\text{b}} = \frac{F_c - F_b}{L_i}. \quad (14)$$

We use a simple slab ocean model and assume that the ocean has a constant mixed layer depth $h_{\text{mixl}} = 30$ m. This ocean layer is characterized by a unique temperature (T_w) that is equal to the freezing point of seawater T_f if there is ice on the area element. When the area element is ice-free, the ocean temperature T_w is computed thanks to the following prognostic equation:

$$\rho_w C_{pw} h_{\text{mixl}} \frac{\partial T_w}{\partial t} = B_1. \quad (15)$$

B_1 is the heat budget of the oceanic area:

$$B_1 = (1 - \alpha_w)F_r + F_L + F_s + F_l + F_b, \quad (16)$$

where α_w stands for the ocean surface albedo. The oceanic flux consists of a restoring term towards a climatological temperature of the mixed layer (T_{mixl}) taken from the Polar science center Hydrographic Climatology (PHC 3.0, updated from: Steele et al., 2001):

$$F_b = \rho_w C_{pw} h_{\text{mixl}} \gamma_t (T_{\text{mixl}} - T_w), \quad (17)$$

where C_{pw} is the specific heat of seawater and γ_t a relaxation constant ($6 \times 10^{-8} \text{ s}^{-1}$).

In order to account for the lateral growth and decay of sea ice, the lead is partially filled with ice of thickness h_0 if the heat budget B_1 of the lead area is negative (Fichfet et al., 1998):

$$S_A^{\text{acc}} = -\frac{(1 - A)B_1}{L_i h_0}. \quad (18)$$

In the case of open water, the ocean temperature has first to be decreased (Eq. (15)) to the freezing point T_f before a new ice layer can be formed. When the heat balance is positive, the lateral decay is computed as:

$$S_A^{\text{ab}} = -\frac{A}{2h_r} \left[-(S_{hr}^{\text{su}} + S_{hr}^{\text{b}}) \right]^+. \quad (19)$$

The term S_h is hence made of four contributions:

$$S_h = A(S_{hr}^{\text{su}} + S_{hr}^{\text{b}}) + S_A^{\text{acc}} h_0 + S_A^{\text{abl}} (h_r + dt(S_{hr}^{\text{su}} + S_{hr}^{\text{b}})). \quad (20)$$

Finally, the atmospheric forcing data sets used to run the model are daily NCEP/NCAR reanalysis data for the air temperature and the

wind velocity, and monthly climatologies for the relative humidity (Trenberth et al., 1989) and the total cloudiness (Berliand and Strokina, 1980).

3. Numerical method

In this section, the equations and the initial and boundary conditions of the sea ice model are presented. The finite element spatial discretization of this boundary value problem is derived and the system of equations obtained is integrated with respect to time.

The sea ice model consists in finding the ice velocity $\mathbf{u}(t,x,y)$, the mean ice thickness $h(t,x,y)$ and the ice concentration $A(t,x,y)$ satisfying the following partial differential equations:

$$\rho h \frac{\partial \mathbf{u}}{\partial t} = \nabla \cdot \boldsymbol{\sigma}(\mathbf{u}, h, A) + \mathbf{S}_u(\mathbf{u}, h, A), \quad (21)$$

$$\frac{\partial h}{\partial t} + \nabla \cdot (\mathbf{u}h) = S_h(h, A), \quad (22)$$

$$\frac{\partial A}{\partial t} + \nabla \cdot (\mathbf{u}A) = S_A(h, A), \quad (23)$$

on the two-dimensional domain Ω englobing the whole Arctic sea ice north of the parallel 50° North. In the momentum equation (21), the term \mathbf{S}_u takes into account the wind and ocean forcings, the Coriolis force and the force due to the tilt of the ocean, and the advection has been neglected (Fichefet et al., 1998) in the momentum balance (Eq. (1)). All the calculations are performed from a given initial situation where all fields are specified. The kind of boundary conditions is twofold: on the one hand, a no-slip condition is prescribed along the coastline and islands. On the other hand, along the open sea boundary, a stress-free boundary condi-

tion is applied, while the inward ice thickness and concentration fluxes are prescribed.

The finite element method consists of three major steps.

- We divide the domain Ω into a triangulation of non overlapping elements Ω_e as shown in Fig. 4. Unstructured grids enable a flexible distribution of the degrees of freedom in the domain according to the areas of interest. As far as sea ice is concerned, the mesh resolution can be coarse where the spatial variability is mostly on large scale and it should be fine where smaller scales are relevant. Hence, in this study focusing on the influence of the CAA on the Arctic ice cover, large elements are used in Central Arctic and in areas of low latitudes, while the mesh is refined along coastlines, especially in the CAA.
- We replace the three unknown fields by piecewise polynomial approximations defined as follows:

$$\mathbf{u}(t, x, y) \approx \mathbf{u}^h(t, x, y) = \sum_{j=1}^p \mathbf{U}_j(t) \phi_j(x, y), \quad (24)$$

$$h(t, x, y) \approx h^h(t, x, y) = \sum_{j=1}^m H_j(t) \psi_j(x, y), \quad (25)$$

$$A(t, x, y) \approx A^h(t, x, y) = \sum_{j=1}^m A_j(t) \psi_j(x, y), \quad (26)$$

where the shape functions ϕ and ψ are defined a priori, and $\mathbf{U}_j(t)$, $H_j(t)$ and $A_j(t)$ are the unknown degrees of freedom. Generally, they correspond to the nodal values in a usual continuous finite element discretization. The local support of those polynomial shape functions is the key ingredient of the finite element method allowing us to derive a discrete algebraic system enjoying nice sparsity properties.

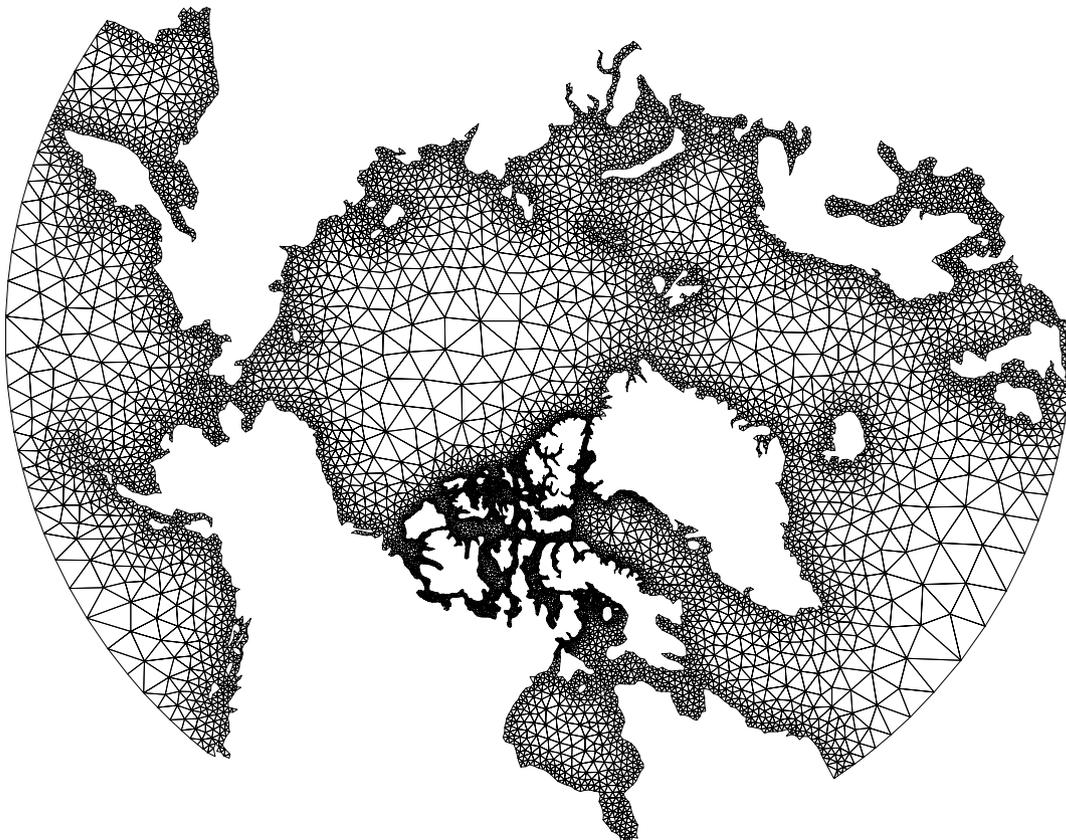


Fig. 4. Finite element mesh of the Arctic. The domain extends north of the parallel 50° North. The mesh contains 17,053 triangles and 10,505 nodes and its resolution varies from 10 to 400 km.

- Finally, we define a discrete Galerkin formulation to obtain numerical values for the unknown degrees of freedom. Such a procedure imposes that the weighted residuals of the partial differential equations have to vanish. Typically, we integrate over the whole domain the partial differential equations weighted by the shape functions. However, even if those residuals vanish, it does not ensure that the differential equations will be satisfied everywhere by \mathbf{u}^h , h^h and A^h . Moreover, some integrals by parts are performed in order to reduce the degree of differentiability required by the shape functions. One clear advantage of the finite element method for sea ice modeling is its facility to handle Neumann or so-called natural boundary conditions (Thomson et al., 1988). Taking advantage of the homogeneous stress boundary condition, the spatial discretization reads: Find \mathbf{u}^h , h^h and A^h such that:

$$\begin{aligned} \left\langle \phi_j \rho h^h \frac{\partial \mathbf{u}^h}{\partial t} \right\rangle &= -\langle \nabla \phi_j \cdot \boldsymbol{\sigma}^h \rangle + \langle \phi_j A^h \mathbf{S}_u(\mathbf{u}^h, h^h, A^h) \rangle \\ j &= 1, \dots, p \\ \left\langle \psi_j \frac{\partial h^h}{\partial t} \right\rangle &= \langle \nabla \psi_j \cdot (\mathbf{u}^h h^h) \rangle - \ll \psi_j h^h (\mathbf{u}^h \cdot \mathbf{n}) \gg + \langle \psi_j S_h(h^h, A^h) \rangle \\ j &= 1, \dots, m \\ \left\langle \psi_j \frac{\partial A^h}{\partial t} \right\rangle &= \langle \nabla \psi_j \cdot (\mathbf{u}^h A^h) \rangle - \ll \psi_j A^h (\mathbf{u}^h \cdot \mathbf{n}) \gg + \langle \psi_j S_A(h^h, A^h) \rangle \\ j &= 1, \dots, m \end{aligned}$$

where \mathbf{n} denotes the outward normal, the notation $\langle \cdot \rangle$ holds for $\int_{\Omega} \cdot d\Omega$ and $\ll \cdot \gg$ for $\int_{\partial\Omega} \cdot d\Gamma$.

This spatial discretization leads to a nonlinear system at each time step whose unknowns are the discrete degrees of freedom. A nonconforming linear approximation (Hua and Thomasset, 1984) is chosen for the velocity field, while the mean ice thickness and concentration are approximated by constant shape functions (Fig. 5).

To illustrate this spatial discretization, the degrees of freedom for velocities are located on the edges of the triangles, while the degrees of freedom for scalar properties A and h are located in the center of the element and represent a value that is constant over the element - similar to an Arakawa C grid. The two-level model proposed by Hibler (1979) suggests indeed a constant mean ice thickness and concentration in each cell, even if not excluding a higher order representation of both variables. Advantages of the linear nonconforming elements include an easier computation of the fluxes through the edges and the orthogonality of the shape functions. In the set of discrete equations, it is important to emphasize that the integral over the domain has to be splitted into the sum of the integrals over each element because the derivative no longer exists on the edge. This is so because nonconforming shape functions are discontinuous across elements edges. In the same way, the boundary integral is now the sum of the integrals along all internal and external edges. For more details, we refer the read-

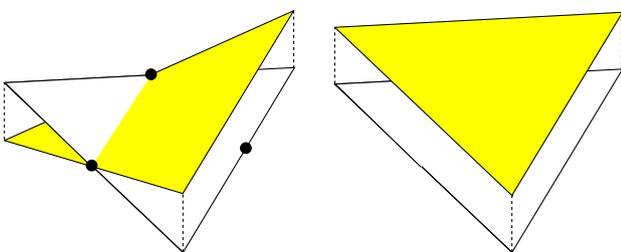


Fig. 5. Linear nonconforming (left) and constant (right) shape functions.

er to Hanert et al. (2004). Finally, we perform a hybrid time stepping scheme where Coriolis force and external forcings due to the ocean are implicitly integrated. To take into account nonlinear terms, a standard linearization based on a Newton–Raphson scheme is introduced, and the resulting sparse matrix is solved with a “skyline” linear solver. Typically, for a simple Euler scheme, the discrete equations read:

$$\begin{aligned} \left\langle \phi_j \rho h_n^h (\mathbf{u}_{n+1}^h - \mathbf{u}_n^h) \right\rangle &= \Delta t (-\langle \nabla \phi_j \cdot \boldsymbol{\sigma}_n^h \rangle + \langle \phi_j \mathbf{S}_u^h(\mathbf{u}_{n+1}^h, h_n^h, A_n^h) \rangle), \\ \left\langle \psi_j (h_{n+1}^h - h_n^h) \right\rangle &= \Delta t (-\ll \psi_j h_n^h (\mathbf{u}_{n+1}^h \cdot \mathbf{n}) \gg + \langle \psi_j S_h(h_n^h, A_n^h) \rangle), \\ \left\langle \psi_j (A_{n+1}^h - A_n^h) \right\rangle &= \Delta t (-\ll \psi_j A_n^h (\mathbf{u}_{n+1}^h \cdot \mathbf{n}) \gg + \langle \psi_j S_A(h_n^h, A_n^h) \rangle), \end{aligned}$$

where the subscript denotes the time evaluation.

Advection is computed via a first-order, finite volume upwind-weighted scheme. Though numerically diffusive, it is monotonic and conservative, two essential ingredients to ensure the positive-definiteness of the transported variables h and A .

4. Results and discussion

The model has been run over the period 1979–2005. We first present results from the control simulation, essentially focusing on the patterns of ice concentration, mean ice thickness and ice drift. A numerical experiment is further presented where we investigate the influence of the ice flow through the narrow straits of the CAA on the Arctic ice thickness distribution and freshwater budget.

4.1. Control run

The model is spun up for only 6 years with annual mean forcings and then integrated with daily wind and air temperature forcings between 1979 and 2005. The mesh used for this simulation has a varying resolution between 10 km in the CAA and 400 km (Fig. 4). Since the resolution in Central Arctic is low, some small-scale and short-term processes might be neglected. However, on the one hand, the results should not be significantly altered, especially as most attention is being paid to the climatological results of the model. On the other hand, the aim of this study being to focus on the influence of the CAA on the Arctic ice cover, this resolution was found sufficient in Central Arctic.

We first compare the annual mean sea ice drifts simulated during the whole 1979–2005 period to the available data (Fig. 6). The daily observed ice motion vectors for the 1979–2005 period are derived from data of Advanced Very High Resolution Radiometer (AVHRR), Scanning Multichannel Microwave Radiometer (SMMR), Special Sensor Microwave/Imager (SSM/I), and International Arctic Buoy Programme (IABP) (<http://nsidc.org>, Fowler, 2003). The simulated ice drift vectors are only drawn on grid points where the annual mean ice concentration is higher than 15%, a value that will be considered here as delimiting the ice edge.

Both the Beaufort Gyre and the Transpolar Drift Stream are recognizable in the observations and the simulation, attesting that the model captures well the essentials of the ice circulation in the Arctic. However, the simulated Beaufort Gyre appears too weak and the model is clearly missing the ice flow in the region of Bering Strait: on a climatological time scale, observations suggest that the ice is almost static there, while the model exhibits a clear ice outflow from the Arctic. This difference is attributed to the effects of the oceanic surface currents flowing from the North Pacific to the Arctic Basin through Bering Strait which are missing in our model and tend to counterbalance the wind stress. The simulated ice dynamics indicates also a higher ice drift along the east coast of Greenland, while a parabolic-like velocity profile across Fram Strait (e.g., Kwok et al., 2004) is expected. It is worth noting that the model does not include any parameterization of landfast sea

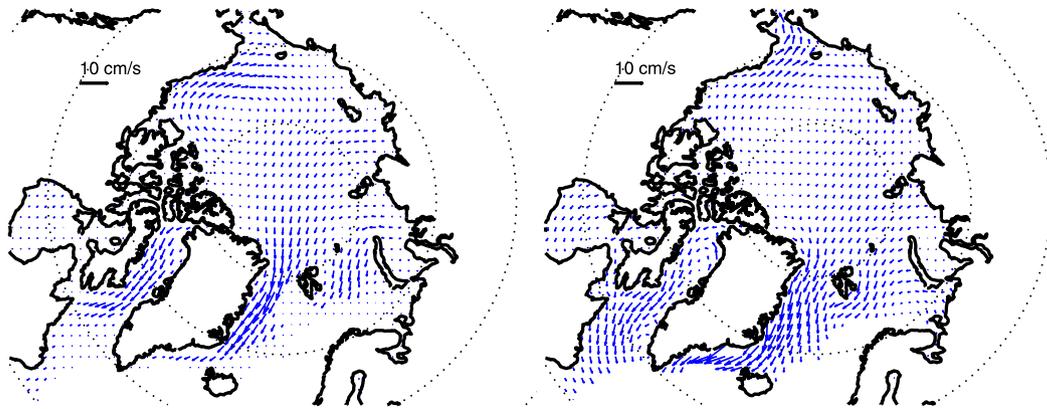


Fig. 6. Mean annual climatological ice drifts (1979–2005). Left: observations; right: the velocity field as computed by our model. Both fields have been interpolated on a regular grid for visual convenience.

ice. We suppose that the fine mesh resolution along the coastline allows the influence of the no-slip boundary condition to vanish rapidly, resulting in a steep variation of the velocity profile. We also note in the simulation the too large ice tongue south of Fram Strait.

Fig. 7 compares the mean ice concentrations from our simulation to those derived from SMMR and SSM/I data sets (<http://nsidc.org>, Comiso, 1999) and interpolated on the finite element mesh. As can be seen from the observations, in winter, the ice is very concentrated in the Arctic Basin (>90%). The simulated ice pack repre-

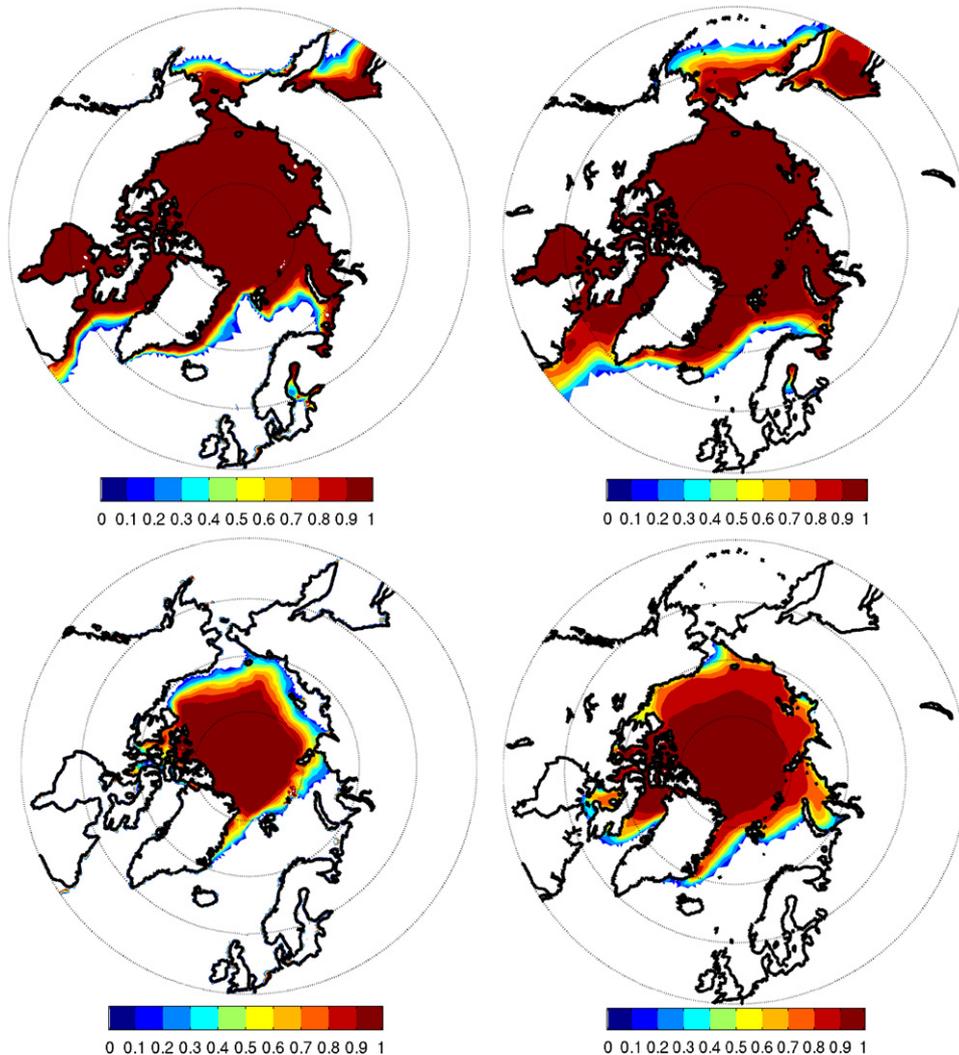


Fig. 7. Mean ice concentrations in March (top) and September (bottom). For both months, we compare the data sets (left, 1979–2005) derived from SMMR and SSM/I and interpolated on the finite element grid with the simulated ice concentrations (right, 1979–2005).

sents this feature, but the ice is too extensive in the southernmost part of the domain. The missing oceanic heat advection is clearly responsible for this overestimation. In the North Atlantic, a huge amount of heat is brought into the Nordic Seas by the North Atlantic Drift Current that originates from the Gulf Stream. This feature is not present in our simulation and mostly explains the too large presence of sea ice in the Norwegian and Labrador Seas. In summer, only the central Arctic Basin remains ice-covered. The simulated ice concentration better matches the observations, even if the model still overestimates the ice cover, especially in Baffin Bay and the Kara Sea.

A question regards whether such a simple model is able to capture the interannual variability of the sea ice cover. For this purpose, Fig. 8 shows the time series of the monthly mean anomalies of the ice extent as simulated by the model and in observations for the 1979–2005 period. Even if the mean climatological ice extent is overestimated in our model, the interannual variability is in reasonable agreement with the observations. Surprisingly, the simulated ice extent is clearly underestimated during the first three years of the simulation. This may be attributed to two factors: the necessary time adaptation after the model spin-up and the too high NCEP/NCAR reanalyzed winter temperatures in the Northern Siberian Coast in 1980 and 1981 (Vancoppenolle et al., submitted for publication). Moreover, the extent minima are not accurately represented and their phase is somewhat delayed.

Finally, there is a general agreement between the simulated ice thicknesses and observations (Fig. 9). In the model Central Arctic, ice thickness lies between 3 and 4 m. Thickness progressively increases from Siberia to the CAA “providing natural barriers to the

movement of the ice” (Bourke and Garrett, 1987), where thickness reaches up to 6 m. The essential of the ice ridging occurs indeed in front of the CAA. Inside the CAA, some spots attain about 7 m (Fig. 11). The influence of resolving the narrow straits constituting the CAA will be investigated in a separate subsection.

4.2. Sensitivity experiment

In order to explore the sensitivity of the sea ice cover to the resolution of the narrow straits of the CAA, two numerical experiments are conducted. The first one (OP) corresponds to the control simulation with the CAA open. In the second experiment (CL), the straits connecting the Arctic Basin to the CAA are closed. For this purpose, the edges of the mesh crossing those straits are converted to land boundaries, thereby disconnecting the Arctic Basin from the CAA. This finally reduces to imposing a zero velocity for the velocity degrees of freedom situated on the edges in red² (see Fig. 10). This procedure is meant to mimic the situation encountered in coarse resolution sea ice models that are not able to represent the exchanges of sea ice through the CAA.

Fig. 11 illustrates the climatological ice thicknesses in March for OP and CL minus OP. As mentioned earlier, the mean ice thickness reaches about 6 m along the CAA. Inside the CAA, thickness progressively decreases southward in direction of Baffin Bay and the Labrador Sea. What strikes first when analyzing the difference between CL and OP is the high similarity between results of both experiments. Outside a restricted area surrounding the CAA entrance, there is no clear difference between OP and CL. Henceforth, on a climatological time scale, this experiment tends to attest that closing the CAA does not significantly alter the general ice

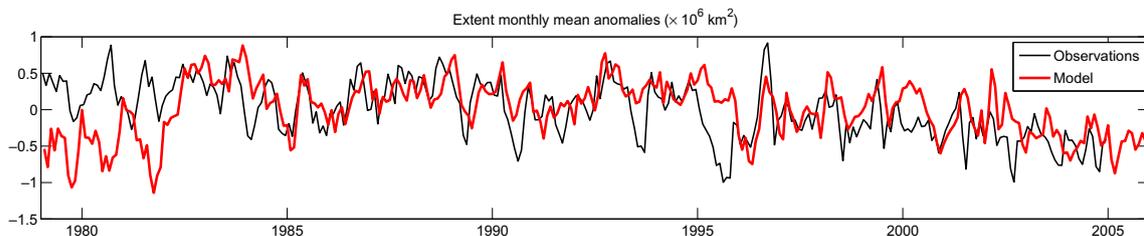


Fig. 8. Time series of monthly mean anomalies of sea ice extent for the Arctic as simulated by the model (red line) and in observations (black line) derived from satellite microwave recordings (Comiso, 1999). The anomalies were obtained by taking the monthly ice extent for each individual month and by subtracting the mean ice extent for that month over the 1979–2005 period. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this paper.)

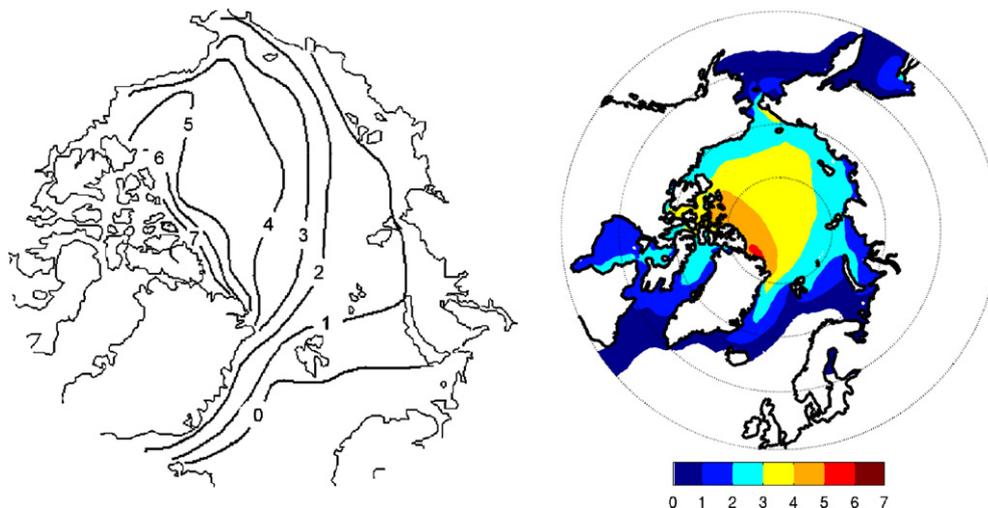


Fig. 9. Global sea ice thickness pattern (in m) as computed by the model (March 1979–2005, right) and as observed (January–April 1960–1982, left, Bourke and Garrett, 1987).

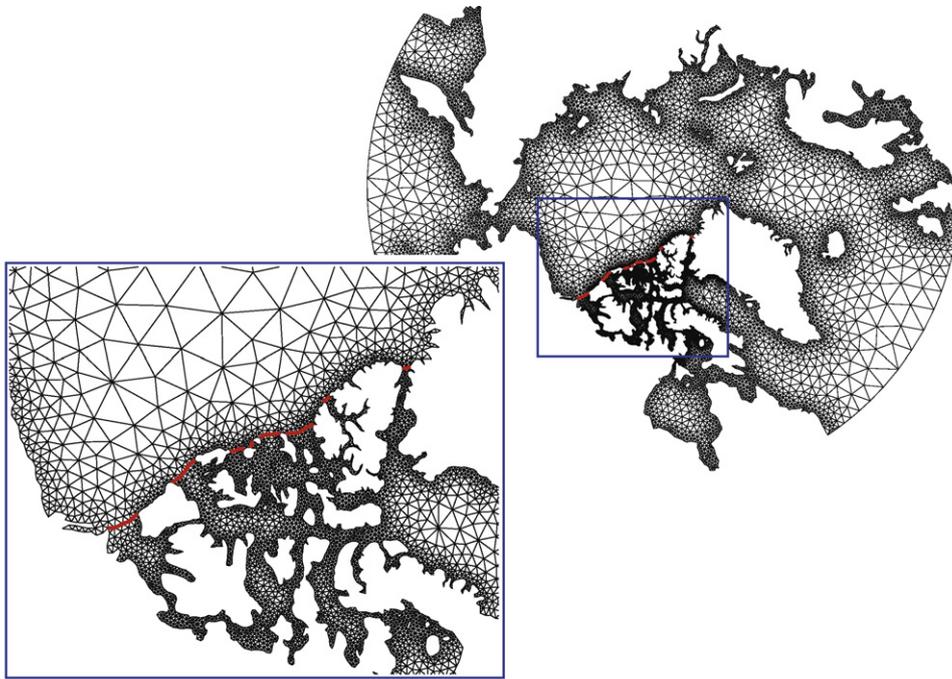


Fig. 10. Zoom on the CAA. The edges in red are considered as land in the CL experiment, thereby closing artificially the Arctic Basin in the Canadian area. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this paper.)

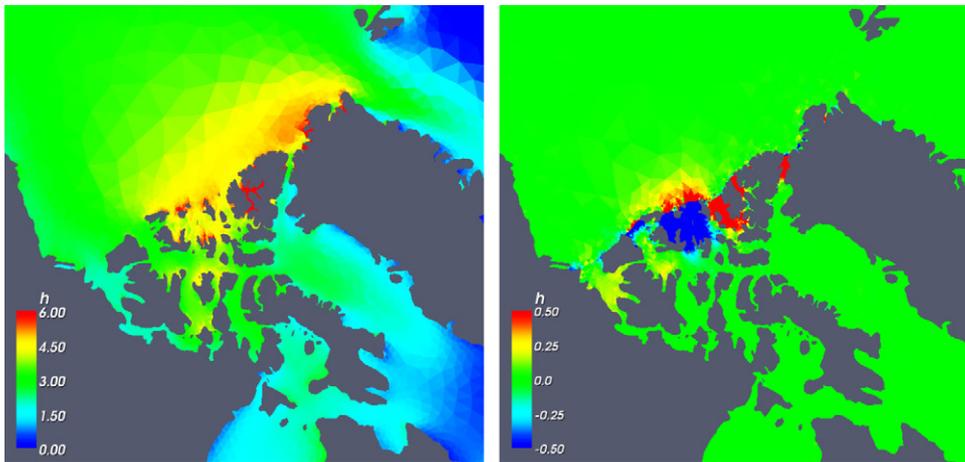


Fig. 11. March ice thicknesses in the CAA for the 1979–2005 period. On the left: control simulation OP, the scale ranges from 0 to 6 m; on the right: difference between CL and OP, the scale ranges from -0.5 m to 0.5 m.

thickness pattern. Nevertheless, the local thickness pattern shows a clear sensitivity to the resolution of the narrow straits north of Canada. In particular, the ice thickness difference between OP and CL is up to 2 m (not shown) in the southern gates of the Queen Elizabeth Islands (QEI-South, see Fig. 1), because of the stopped ice flux from the Arctic to the CAA.

On a daily basis, the comparison between OP and CL reveals stronger differences. The mean ice thickness patterns are compared for late October 2005 in Fig. 12. During that period, the wind is blowing offshore the CAA. Two points are to be outlined here. First, along the Queen Elizabeth Islands, the ice divergence caused by the wind is much more pronounced in CL because the closed straits prevent any ice thickness outflow from the CAA. The influence of the domain boundary and the related boundary condition diffuses inside the domain. Moreover, even a very thick ice cover, which is supposed to have high ice strength, is very sensitive

to the wind direction. In a few days, the mean ice thickness along the CAA has decreased by more than 2 m. It is worth mentioning that, in Hibler's (1979) model, no tensile strength is applied to the ice and that no landfast parameterization is included. The ice accumulated against the CAA is hence more sensitive to the wind direction in the CL experiment. Second, for the same reason, closing Amundsen Gulf (southern strait in Fig. 10) in CL leads to an ice accumulation inside the CAA, while in Amundsen Gulf, it generates a large open water area, with all the physical processes involved. The ice fluxes between both regions are thus important to accurately represent the ice flow in northern Canada.

Exchanges of sea ice between the CAA and the Arctic are extensively described and quantified by Kwok (2006). This study based on RADARSAT ice motion provides a 5-year complete record of ice area fluxes in the northern CAA. The mean annual observed ice fluxes for the 1998–2002 period (Table 1) indicate a northward

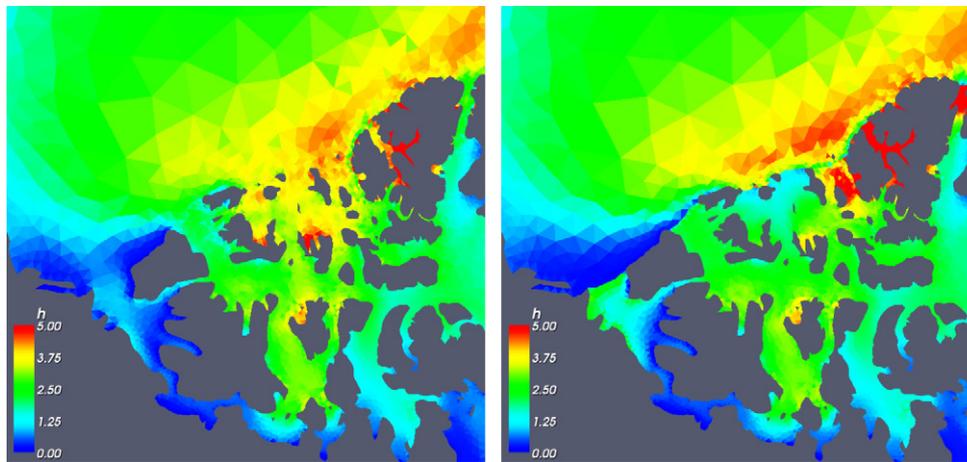


Fig. 12. Ice thicknesses (from 0 to 5 m) in the CAA simulated by the model for late October 2005. On the left: control simulation OP; on the right: CL. Note the formation of open water in the Beaufort Sea for simulation CL and the difference in thickness between OP and CL along the Queen Elizabeth Islands.

Table 1

Mean annual ice area ($\times 10^3 \text{ km}^2 \text{ yr}^{-1}$) and volume ($\text{km}^3 \text{ yr}^{-1}$) fluxes for the 1998–2002 period through Amundsen Gulf (AG), M'Clure Strait (MS) and Queen Elizabeth Islands (QEI) in observations (Kwok, 2006) and as computed by Sou's (2007) model and by our model

	Kwok (2006)			Sou (2007)			Model		
	Area	Volume	Thickness	Area	Volume	Thickness	Area	Volume	Thickness
AG	–85	–85	1.0	–130	–97	1.1	–70	–87	1.5
MS	–20	–80	4.0	2	54	2.0	–19	–32	2.6
QEI	8	27	3.4	11	72	3.6	33	131	4.2
Net	–97	–138		–117	29		–56	12	

Positive sign indicates Arctic outflow. The ice volume estimated by Kwok (2006) is based on a representational ice thickness (m).

ice movement in Amundsen Gulf and M'Clure Strait and an Arctic outflow in the Queen Elizabeth Islands. Overall, ice area is exported from the CAA to the Arctic, but the seasonal and the interannual variability of these ice fluxes are very high. Ice volume fluxes are estimated on the basis of the ice area fluxes and representational ice thicknesses.

Sou's (2007) study uses a finite difference regional ice-ocean model forced by NCEP/NCAR reanalyzes data with two ice thickness categories and viscous-plastic rheology. The model is run on a nested grid with a 0.2° horizontal resolution which covers the CAA region. The ice area fluxes computed by this model (Table 1) compare reasonably well with the observations, in particular in the QEI region. On the low side, the ice outflow in Amundsen Gulf is too large and the direction of the flux in M'Clure Strait is opposite to what is observed. Sou (2007) has reservations about the accuracy of the use of representational ice thicknesses to estimate the “observed” ice volume fluxes in regions of high thickness or ice direction variability. For example, in the case of an oscillating ice flux through a gate whose ice thickness is higher at one of its sides, the net ice area flux would be close to zero while the ice volume flux will not. The weighting played by the non uniform ice thickness appears to be the leading factor.

Our model shows a good agreement with the observed ice area fluxes, with typically the same direction and order of magnitude. The model does particularly well in M'Clure Strait. Discrepancies of the model include too large an ice flux through the Queen Elizabeth Islands, especially in winter when the model is missing the landfast sea ice. The net exchange in ice area between the Arctic and the CAA is hence smaller than in observations. Due to the method used, the agreement between modeled ice volume flux to the “observed” one depends highly on the variability of the region (Sou, 2007). In both models for example, the estimated ice volume

flux based on the representational ice thickness overestimates the modeled ice volume flux in M'Clure Strait (+54% in our simulation). However, in the QEI, the estimated ice volume flux ($139 \text{ km}^3 \text{ yr}^{-1}$) is very close to the one computed by our model. Both models exhibit the same discrepancy between the net ice area (Arctic inflow, consistently with observations) and volume (Arctic outflow) fluxes. The weighting of the ice area fluxes by the non uniform ice thickness in the CAA seems to explain in both models this discrepancy.

As expected, North Nares Strait monthly ice area fluxes are weak on average ($<200 \text{ km}^2$, not shown here), but with large northward exports for a couple of months, yielding a net Arctic inflow ($-3 \times 10^3 \text{ km}^2 \text{ yr}^{-1}$, 1998–2002). This discrepancy with the observations is also present in Sou's (2007) study and is mainly attributed to the coarse wind forcing missing the effects of the topography on the local wind pattern (Sou, 2007). The absence of oceanic currents and the lack of mesh resolution in North Nares Strait can also be adduced. In the next subsection, we investigate the influence of those ice fluxes on the freshwater budget of the Arctic.

4.3. Importance of the CAA for the mass balance of Arctic sea ice

The resolution of the CAA still constitutes in the large majority of sea ice models a missing piece of the Jigsaw puzzle. In order to evaluate the importance of the CAA for the mass balance of Arctic sea ice, we perform two additional analyses in the control run OP. First, the sea ice volume in the CAA is computed by summing the contributions of all the elements situated inside a restricted area defined by a triangle (see Fig. 13) supposed to englobe the whole CAA. According to our model, the mean CAA ice mass represents a rough 10% of the total simulated sea ice volume, which is far from being negligible.

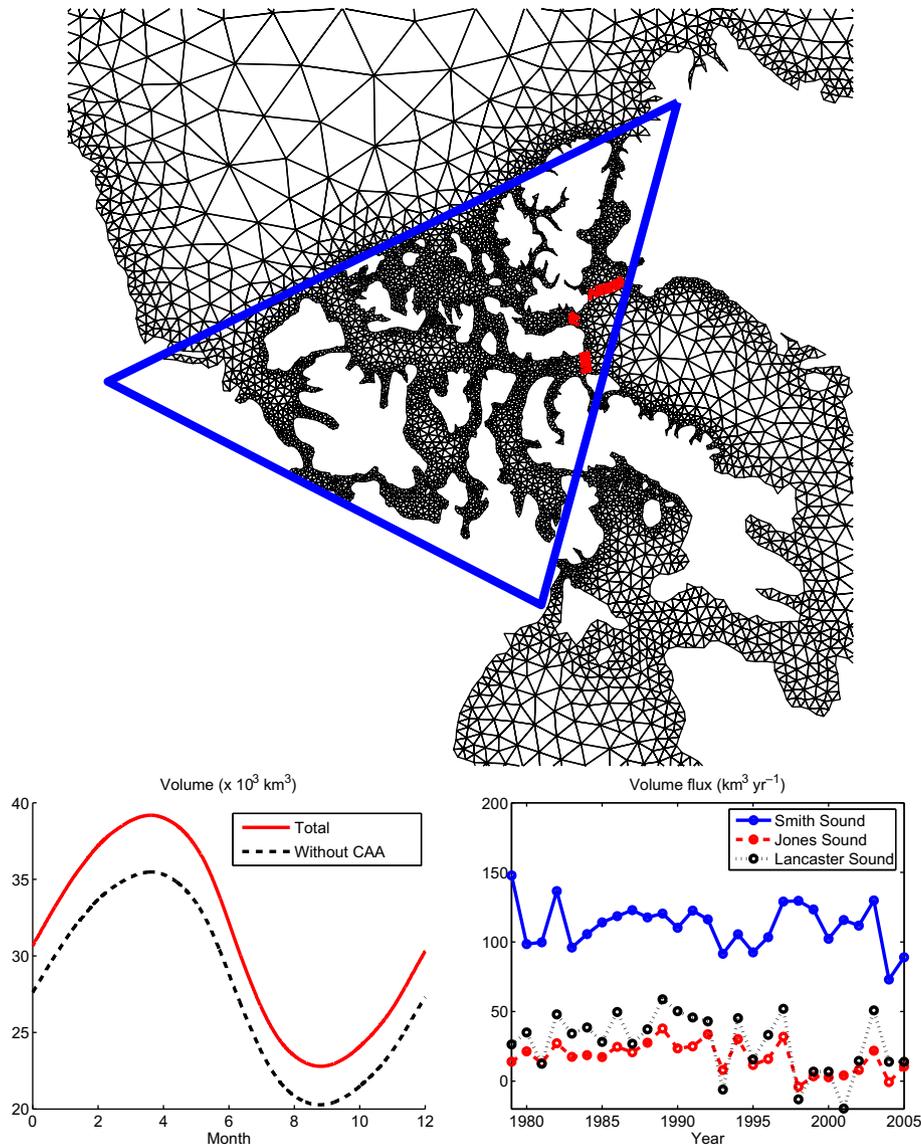


Fig. 13. Top: close-up view of the mesh in the CAA. Bottom left: seasonal cycles of the Arctic sea ice volume with (solid) and without (dashed) the contribution of the CAA. Bottom right: annual ice volume fluxes (positive if directed towards Baffin Bay) through the three gates represented in the upper figure (bottom-up: Lancaster Sound, Jones Sound and Smith Sound). All the elements included in the triangle in the upper figure are considered being part of the CAA.

Second, the ice volume fluxes towards Baffin Bay are evaluated at three gates described in Fig. 13 and situated in Lancaster Sound, Jones Sound and Smith Sound. The fluxes experience a strong inter-annual variability and, except for three years in Lancaster Sound, contribute to a positive ice flux to Baffin Bay. As expected, Smith Sound yields the major contribution with roughly 70% of the mean total sea ice export, while Lancaster and Jones Sounds contribute to 20% and 10%, respectively. Converted into mean annual freshwater fluxes, we find a contribution of $\sim 20 \text{ km}^3 \text{ yr}^{-1}$ for Lancaster Sound, $\sim 15 \text{ km}^3 \text{ yr}^{-1}$ for Jones Sound and $\sim 90 \text{ km}^3 \text{ yr}^{-1}$ for Smith Sound. The mean total sea ice freshwater export towards Baffin Bay is hence of $125 \text{ km}^3 \text{ yr}^{-1}$, which is close to the $160 \text{ km}^3 \text{ yr}^{-1}$ estimated by Serreze et al. (2006), on the basis of Prinsenberg and Hamilton's (2005) study, and computed by Sou (2007) on a long period (1950–2004). As mentioned by Sou (2007), the ice volume transport is probably underestimated in models in narrow straits like Jones and Lancaster Sound due to the coarse wind forcing and the low mesh resolution. In order to evaluate the relative importance of the ice export through the CAA, this value is to be

compared to the mean freshwater export through Fram Strait. According to recent observations, our estimate represents only 7% of the freshwater equivalent of the ice export through Fram Strait ($1800 \text{ km}^3 \text{ yr}^{-1}$, estimation from Kwok et al. (2004) converted into freshwater equivalent) and 5% of Fram Strait's liquid freshwater outflow ($2400 \text{ km}^3 \text{ yr}^{-1}$, Serreze et al., 2006). Though relatively small, the impact of this ice export on the water stratification in Baffin Bay and on the convective overturning downstream in Labrador Sea has to be investigated.

5. Conclusions

In this paper, we have presented a finite element, unstructured grid sea ice model. The model includes a viscous-plastic rheology along with a complete parameterization of the atmospheric fluxes and is driven by daily NCEP/NCAR reanalysis data. The climatological sea ice drifts, thicknesses and concentrations computed by the model compare qualitatively well with the observations. The finite element method presented in this paper reveals effectiveness for

sea ice modeling and enables the use of unstructured grids with all their potential and flexibility.

A clear advantage of unstructured grids is to enable the resolution of the narrow straits of the Canadian Arctic Archipelago (CAA). A numerical experiment has hence been performed to investigate the influence of the ice flow through the CAA. Focusing on the large-scale sea ice thickness pattern, within our general framework and hypotheses, we have shown that the inclusion of those straits is not essential, the impact being merely local. This tends to validate the idea of the role of the CAA as a “buffer” on sea ice between the Arctic Basin and Baffin Bay. However, we have shown that the local and short-term influences of the ice exchanges are nonnegligible. In particular, depending whether the straits are open or closed in the numerical experiment, the domain boundary and the associated boundary condition influence directly the numerical solution in the proximity of those straits. Moreover, on average, the sea ice volume in the CAA represents a nonnegligible 10% of the total sea ice volume in our model. Finally, the annual sea ice volume flux towards Baffin Bay has been evaluated in order to assess its importance in the freshwater balance of the Arctic and yields a mean freshwater outflow of $125 \text{ km}^3 \text{ yr}^{-1}$, which is in good agreement with the $160 \text{ km}^3 \text{ yr}^{-1}$ suggested by Serreze et al. (2006), on the basis of Prinsenberg and Hamilton's (2005) study. Though relatively small compared to the ice contribution of the freshwater outflow through Fram Strait (7%), this ice export, together with the North Water Polynya formation, might be important regarding the water stratification in Baffin Bay and the decay of the summer sea ice cover in Baffin Bay. All these diagnoses show how important the CAA is on the mass balance of the Arctic sea ice and plead for the need to better understand the complex processes and interactions being held in this region.

The results presented in this study must be considered with some care though since no ocean dynamics nor oceanic feedbacks are included in this uncoupled version of the model. The deficiencies of the model, including the Bering outflow, are to be attributed to these factors. In particular, the absence of oceanic heat advection is clearly responsible for the overestimation of the simulated sea ice extent. An important issue addressed by Sou (2007) is to know whether or not the coarse resolution atmospheric forcings typically used in the models is sufficiently accurate to enable realistic simulations in the CAA. As stated by Wang et al. (2003), high-resolution atmospheric forcing fields might help to solve this issue. Another open question concerns the mesh resolution needed for resolving the CAA versus the continuum assumption inherent of the viscous-plastic rheology.

Some additional investigation based on a coupled sea ice-ocean model should be carried out to assess the actual role of sea ice passing through the CAA on the global ocean circulation, by way of the interactions between the associated freshwater flux and the water mass properties of the adjacent seas. Recently, this issue has been readdressed by Kwok (2005) concerning the impact of the ice export through Nares Strait and more generally from the CAA on the convective overturning downstream in the Labrador Sea.

Future work on the sea ice model will include a better representation of the sea ice physics [halo-thermodynamic model (Vancoppenolle et al., 2007), ice-thickness distribution (Thorndike et al., 1975), etc.]. An intermediate step towards the development of a coupled ice-ocean model might consist of including climatological modeled currents (e.g., Maslowski et al., 2004) in order to improve the ice circulation in areas where these currents play an important role (Bering Strait, Nares Strait, etc.). A coupled unstructured grid ice-ocean model might enable to assess the role of the ice fluxes through the CAA, explain the formation of static arches or propose lacking data (such as thickness order of magnitude for estimations of mass budgets). As outlined by Dickson et al. (2007), numerical models have indeed a role to play to make up

for the inherent difficulties and financial limitations to collect data, and in particular ocean fluxes. Finally, the Hibler (1979) sea ice rheology imposes a zero tensile strength. We have shown that, with our model, we are not able to represent static landfast sea ice in the North Canadian Archipelago. This issue might be resolved by including some parameterization of the landfast sea ice in the model. However, detailed bathymetry and well-resolved coastal currents are needed for an appropriate simulation of landfast sea ice (Wang et al., 2003). Hence, it appears that, instead of adding new features and parameterizations to the actual model, significant improvements might be gained by investigating a new rheology integrating recent laboratory research work on the mechanical properties of the ice and observations based on buoys tracking (e.g., Schreyer et al., 2006; Coon et al., 2007; Weiss et al., 2007).

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