Modeling the dynamics and drift of icebergs in the Barents Sea

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Preface

This synthesis and collection of papers is submitted for the degree of Philosophiae Doctor in oceanography at the Geophysical Institute, University of Bergen. This work was carried out at the Nansen Environmental and Remote Sensing Center, in the Mohn Sverdrup department. An iceberg forecasting system for the Barents Sea is developed and evaluated. The system is used to improve our understanding and potential predictability of iceberg drift in the area. The thesis consists of an introduction, where the motivations, objectives and results are presented, and three papers that build on each other, listed below:

**Paper I**  

**Paper II**  

**Paper III**  
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Introduction

The World Meteorological Organisation sea ice nomenclature defines an iceberg as a massive piece of ice of varying shape, protruding more than 5 m above sea-level, which may be afloat or aground. It has broken away from a glacier or an ice shelf\(^1\), which is a process known as iceberg calving. This process involves the propagation of fractures which can be triggered by ice stresses, tides, waves, bottom melting and ablation. The calving rate is controlled by the glacier speed, the geometric changes at the terminus region and submarine melting (Rignot et al., 2010; Straneo et al., 2010; Van der Veen, 2002).

The shape and size of the icebergs is determined by formation or deterioration processes. They are subdivided into different categories (see Figure 1 and Table 1).

The drift of icebergs is influenced mainly by ocean currents, winds, the Coriolis force, waves, sea ice (concentration and drift) and bathymetry (by grounding). The acceleration of water entrained in the turbulent wake of the iceberg also has a small influence (Sodhi and El-Tahan, 1980; Smith, 1993). Atmospheric and oceanic forces act on the areas above and below the water line, respectively. Form drag forces act on the vertical plane and surface drag forces act on the horizontal plane of the iceberg. In general, form drag forces dominate surface drag forces. The surface drag forces become the dominant forces if the horizontal surface drag area exceeds about 250 times the sail area of the iceberg (Smith and Banke, 1983). This may occur for ice islands or huge tabular icebergs. Sea ice, if sufficiently packed with strong internal ice stresses, may lock the iceberg and the iceberg then drifts only with the sea ice (Lichey and Hellmer, 2001). Generally, during its drift, the iceberg undergoes strong modification through ablation and melting, depending on the season and the region. The wave erosion, calving of overhanging slabs, lateral melting, basal melting and melting due to solar radiation are the most important iceberg deterioration processes (Kubat et al., 2007).

Iceberg dynamics and thermodynamics are of great interest because they potentially represent

\(^1\)Floating glacier of large dimensions extending beyond the coastline.

<table>
<thead>
<tr>
<th>Descriptive name</th>
<th>Freeboard height (m)</th>
<th>Length (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Growler</td>
<td>&lt; 5</td>
<td>&lt; 5</td>
</tr>
<tr>
<td>Bergy Bit</td>
<td>1 - 5</td>
<td>5 - 15</td>
</tr>
<tr>
<td>Small Berg</td>
<td>5 - 15</td>
<td>15 - 60</td>
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<tr>
<td>Medium Berg</td>
<td>16 - 45</td>
<td>61 - 120</td>
</tr>
<tr>
<td>Large Berg</td>
<td>46 - 75</td>
<td>121 - 200</td>
</tr>
<tr>
<td>Very Large Berg</td>
<td>&gt; 75</td>
<td>&gt; 200</td>
</tr>
</tbody>
</table>

Table 1: Iceberg size classification used by the International Ice Patrol.
severe marine hazards and impact the ecosystem (Teixidó et al., 2007), the geology (Freiwald et al., 1999) and the hydrography of the world’s oceans (Heinrich, 1988; Jansen et al., 2007; Wiersma and Jongma, 2009). They can be dangerous through a collision or via scouring\(^2\) of the ocean floor. The latter process damages deep sea life (Gerdes et al., 2003), offshore installations and reshapes the sea bed (Barnes and Lien, 1988). Icebergs are also a source for marine life: inland ice collects materials during its advance towards the coast. Hence, while melting in the ocean, icebergs provide nutrients for the ocean’s euphotic zone (Lichy and Hellmer, 2001). Additionally, they leave sediments and stones on the sea floor that can be used as paleo-climate indicators (Alvarez-Solas et al., 2010; Heinrich, 1988). Finally, they are a source of fresh water for the world’s oceans that may impact the ocean circulation (Jansen et al., 2007). When icebergs calve from land into the sea, they furthermore have an impact on the global sea level. West Antarctica and Greenland have experienced recent mass loss acceleration. Since 2003, their contribution to the global annual sea level rise (3.3 ±0.4 mm) has nearly doubled and

\(^2\)Plough marks made by icebergs keels as they pass over sea beds.
represents almost 30% of the total rise (Nicholls and Cazenave, 2010).

**Icebergs in the world’s oceans**

**Southern Ocean**

The major source of icebergs from the Southern Ocean are the ice shelves that make up to 44% of the Antarctic coastline (Barnes and Lien, 1988). Icebergs from Antarctica are often relatively big tabular icebergs (see Figure 1) with a surface area of several tens of squared kilometres and several hundred meters thick. They represent about 2000 Gt of glacial ice released each year in the Southern Ocean. It is about four times bigger than the fresh water flux due to basal melting beneath the ice shelves (Jacobs et al., 1992; Jansen et al., 2007). Hence, icebergs may influence the ocean circulation by affecting deep water formation. During their melting, they stabilize the weakly stratified water column. Fresh waters released from deep iceberg bases enhance heat transfer and up-well nutrient-rich waters from below the pycnocline to the surface (Jenkins, 1999; Lichey and Hellmer, 2001). In the late seventies, the potential freshwater source from Antarctic icebergs interested scientists, engineers and entrepreneurs. Two international conferences were held to investigate the feasibility of transporting Antarctic icebergs towards arid regions such as Australia, California or Saudi Arabia. This controversial idea did not materialize but fortunately initiated iceberg modeling studies.

Antarctic icebergs also have a non-negligible impact on waves. A wave forecast model needs to include iceberg distributions to limit the errors in the Southern Ocean. Indeed, Ardhuin et al. (2010) show that iceberg distribution, after wind and sea ice, is an essential component for wave forecast.

Icebergs drifting northward are hazardous for ships off the Cape Horn. They may also be a threat for research and tourist ships operating in Antarctic waters. Their implication in the Southern Ocean circulation and the danger that they represent motivated the development of detection techniques that aim to become systematic.

Antarctic iceberg observations are available through ship or aircraft reconnaissance, GPS buoys (Schodlok et al., 2006) and satellite observations. The National Ice Center (NIC; Washington D.C., USA) systematically tracks icebergs longer than 18.5 km. More recently, Silva and Bigg (2005) presented a method to identify and track icebergs that are only 200 m long using the Synthetic Aperture Radar (SAR). Finally, (Tournadre et al., 2008) demonstrated the capability of radar altimetry to detect icebergs less than 1 km in size. The recent launch of Cryosat 2 by the European Space Agency is expected to strengthen this latter approach.

**North Atlantic and Arctic oceans**

Icebergs are common in Arctic waters, along the Baffin Bay, the Labrador Coast, and on the Grand Banks of Newfoundland. Icebergs found in the North Atlantic mainly originate from the western and eastern Greenland glaciers. Icebergs found in the “central” Arctic such as in the Beaufort Sea come from Arctic ice shelves. The biggest Arctic ice shelves are located on

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3Iceberg Utilization for Fresh Water Production, Weather Modification and Other Application, 1977; Conference on use of icebergs, 1980.
Ellesmere Island. Finally icebergs found on the eurasian side of the Arctic come from glaciers located on islands surrounding the Barents Sea. Their initial size varies from growler size (see Table 1) to icebergs 1 km long and over 200 m high. Ice islands can be found in the Beaufort Sea Gyre. Note that on 5 August 2010, a huge ice island, with an area of 270 km\(^2\), detached from Petermann floating ice shelf, north east of Greenland. It is the biggest calving event in the Arctic in nearly 50 years. The Greenland ice sheet is the most important source of icebergs in the northern hemisphere. Its estimated iceberg calving rate is about 200-400 Gt per year (Reeh, 1994; Bigg, Grant R., 1999; Rignot and Kanagaratnam, 2006).

Icebergs near the Atlantic shipping lanes are of greatest concern. International cooperative actions for safety started after the RMS TITANIC collided with an iceberg south of the Grand Banks on 15 April 1912, causing the death of more than 1500 people. Since then, the International Ice Patrol (IIP, http://www.uscg-iip.org/cms/) was established. Each year, the IIP reports the position of icebergs and estimates their probable courses. Today, satellite images from RADARSAT-2, reconnaissance flights and ships, and an iceberg forecast model are used to define the southern limit of the iceberg area and to track individual icebergs over the Grand Banks. The iceberg forecast model (Kubat et al., 2005) is developed by the Canadian Ice Service (CIS).

Another area where human activities are expected to increase is the Barents Sea. As this is the focus area of the thesis, its case is presented in more details in the next section.

**Scope of the thesis**

In recent years, the high north and in particular the Barents Sea has become an important strategic area for Norway. Its policy is intended to protect the environment, assure safety and promote the developments of sustainable exploitation and management of ocean resources, such as the oil and gas industry and fisheries. One of the world’s largest natural gas fields, Shtokman Gaz Condensate Field (SGCF), lies south east of the Barents Sea (Figure 2). Pipeline gas production might start in 2016 and will intensify ship traffic. The potential collision with an iceberg represents the highest risk for floating platforms and ground installations, not to mention the additional difficulties associated with the cold environment, such as the presence of sea ice and intense ice frost. For these reasons, there is a need for an adequate iceberg monitoring and forecasting system for the region.

**The Barents Sea**

Barents Sea icebergs evolve in a region with irregular topography and complicated ocean, sea ice and wind conditions that are challenging to forecast. In the following, an overview of the local iceberg characteristics precedes a description of the climatic context on seasonal and interannual time scales.

The main source of icebergs in the Barents Sea is the Svalbard Archipelago and especially the Austfonna ice cap (Dowdeswell et al., 2008). The Franz Josef Land glaciers and in particular the Renown Glacier on Wilczek Land (Kubyshkin et al., 2006) are among the secondary sources. A smaller contribution of icebergs migrates from glaciers of the northern tip of Novaya Zemlya (Figure 2). Observational campaigns under the Ice Data Acquisition Program (Spring,
Figure 2: Locations of the different sources of icebergs (filled blue circles) and the main ocean currents. Light and dark blue contour lines are the isobaths at 100 and 200 m respectively. The location of the Shtockman Gaz Condensate Field is identified with the red symbol and abbreviation.

1994, IDAP) from 1988 to 1993 reported that their average and maximum size above the sea surface was: 91 m length $\times$ 64 m width $\times$ 15 m height and 320 m $\times$ 279 m $\times$ 40 m, respectively. Although a great proportion are grounded or melt close to their calving area, icebergs were found as far south as 67.2°N during the summer of 1929 (Abramov and Tunik, 1996) and more recently close to the SGCF, south east of the Barents Sea in May 2003 (Zubakin et al., 2004).

The Barents Sea is a shallow shelf sea, with connections to the Norwegian Sea to the west and the Kara Sea to the east. It is only 230 m deep on average, with deeper trenches up to $\sim$500 m deep and shallow banks of $\sim$100 m deep (Figure 2). Shallow areas such as the Great Bank and Central Bank influence the current patterns. Water mass modification and dense water formation occurs also in shallow areas. For instance, dense water formation by cooling and freezing is most effective on the shelves, Storfjorden and Novaya Zemlya bank (Vinje and Kvalmevik, 1991).

The ocean currents distribution is dominated by the presence of Atlantic Water (AW) and the Polar Water (PW). Their junction defines the location of the Barents Sea Polar Front: the dominant large scale feature of the central Barents Sea. The warm and saline AW, coming from the Norwegian Sea, splits into several branches while crossing the western boundary of the Barents Sea. One branch continues northward entering the Arctic Ocean west of Svalbard as the West Spitsbergen Current while another branch enters the Barents Sea as the North Cape Current.
The North Cape Current continues eastward or north-eastward while bifurcating into smaller branches. The position of the Polar Front is closely tied to the topography on the southern flank of Svalbard Bank and the width of the North Cape Current (Johannessen and Foster, 1978; Ingvaldsen, 2005). The latter is strongly modified by cooling, mixing and freezing during winter before entering the Arctic Ocean. In addition to its seasonal variability, its interannual variability is pronounced (Ingvaldsen et al., 2004; Furevik, 2001). The Persey and the East Spitsbergen Currents transport cold PW from Arctic origin southward into the Barents Sea (Pfirman et al., 1994). Tides in that region are among the strongest tides of the entire Arctic Ocean (Kowalik and Proshutinsky, 1993), and play an important role by contributing to the non uniform transformation of water masses. Maximum tidal amplitudes occur close to Spitsbergen, Bear Island, Hopen Island and in the White Sea (Kowalik and Proshutinsky, 1995).

Meteorological conditions of the Barents Sea are dominated by cyclones formed in the North Atlantic, which transport heat and moisture from lower latitudes towards the Barents Sea. In winter, close to the polar front, polar lows form with typical wind speeds reaching storm force in a short time. East Siberian Cyclones generate favorable conditions for strong northerly wind anomalies in the Kara and Barents Sea region (Sorteberg and Kvingedal, 2006). In summer the pressure gradients are weaker and the wind direction is more variable.

A large part of sea ice in the Barents Sea is formed locally, generally starting in September/October. It is affected by the salinity of the sea water, topography, wind, currents, sea state conditions, air temperature and the heat flux. The most active places of sea ice formation are on the shelves such as Storfjorden, southeast of Spitsbergen, Novaya Zemlya bank and Franz Josef Land. The annual cycle of the sea ice extent is characterized by a minimum in September and a maximum in April with large interannual variability (Sorteberg and Kvingedal, 2006).

Objectives

The main objective of this thesis is to implement and validate an advanced iceberg drift model for the Barents Sea and use the system to determine characteristics of iceberg distribution on seasonal and interannual time scales. The specific objectives are to:

- Implement an iceberg-sea ice-ocean nested model for the Barents Sea.
- Analyze and characterize the limitations of the system
- Provide a simulated climatology of iceberg characteristics in the Barents Sea to complement and extend sparse observations from oceanographic fields campaigns, ice reconnaissance flights and satellite observations in the region.
- Find mechanisms controlling the seasonal and interannual variability of iceberg extension and particularly the extreme southernmost extension.
- Improve the predictive skills of the model system by using an advanced data assimilation method in a step towards an operational iceberg drift model for the region.

Methods

Several modeling studies hindcast the drift of icebergs using forcing fields derived from observations (Smith and Banke, 1983; Kubat et al., 2005). Observations of ocean currents and
iceberg characteristics in the Barents Sea are largely insufficient to represent their complex dynamics. The lack of data motivated for the set-up of a coupled ice-ocean model to force the iceberg drift model. An ensemble is used to simulate the non-linear properties of iceberg trajectories.

**Iceberg model**

The dynamics of the iceberg module implemented in this thesis are based on the iceberg drift model introduced by Smith and Banke (1983) and further developed by Lichey and Hellmer (2001).

**Dynamics**

The iceberg acceleration is proportional to the forces from the atmosphere ($F_{AT}$), the water drag ($F_W$), the Coriolis force ($F_C$), the force due to the sea surface slope ($F_{SS}$) and the force due to interaction with the sea ice cover ($F_{SI}$):

$$M \frac{du}{dt} = F_{AT} + F_W + F_C + F_{SS} + F_{SI},$$  

(1)

where $M$ is the iceberg mass and $u$ is the iceberg velocity. The atmospheric force is

$$F_{AT} = \left[ \frac{1}{2}(\rho_a c_a A_{va}) + (\rho_a c_{da} A_{ha}) \right] |v_a - u|(v_a - u).$$  

(2)

The oceanic force $F_W$ is defined by the same quadratic drag law as $F_{AT}$ for each ocean model layer through the depth of the iceberg,

$$F_W = \sum_{k=1}^{n} \left[ \frac{1}{2}(\rho_w c_w A_{vw}(k))|v_w(k) - u|(v_w(k) - u) + (\rho_w c_{dw} A_{hw}(n))|v_w(n) - u|(v_w(n) - u).$$  

(3)

In Equations 2 and 3, $v_a$ (resp. $v_w(k)$) is the air (resp. oceanic) velocity. The index $k$ is the ocean layer number and $n$ is the number of ocean layers in contact with the iceberg. The term $A_{va}$ (resp. $A_{vw}(k)$) is the vertical cross section area in the air (resp. water). The air and water densities are $\rho_a$ and $\rho_w$ respectively, $c_a$ and $c_w$ are the form drag coefficients, $c_{da}$ and $c_{dw}$ are the skin drag coefficients set to 0.0022 and 0.0055 respectively, the same values as used in the sea ice model. $A_{ha}$ (resp. $A_{hw}(n)$) is the horizontal area of the iceberg in contact with the air (resp. ocean layer $n$). The wind is assumed to be constant with height above sea level. However, the ocean currents vary with depth as given by the ocean model. Note that the form drag coefficients $c_a$ and $c_w$ are commonly introduced in the drag forces to account for errors in the estimated area of the vertical plane in contact with the ocean currents and the atmosphere. In addition those parameters include errors in the forcing and the parameterizations. Here, the acceleration of water entrained in the turbulent wake of the iceberg is assumed to be only depending on the mass and is therefore included through an adjustment of the form drags. The effect of wind waves is also included implicitly in the atmospheric forcing through an adjustment of the air form drag coefficient (Paper I & II).

The third force acting on the iceberg is the Coriolis force,

$$F_C = 2M\Omega \sin \phi k \times u,$$  

(4)
where $\Omega$ is the angular velocity of the Earth, $\phi$ is the latitude, $\mathbf{k}$ is the unit vector perpendicular to the Earth’s surface and $\mathbf{u}$ the iceberg velocity. The force due to the sea surface slope is,
\[
F_{SS} = -Mg \sin \alpha,
\] (5)
where $g$ is the acceleration due to the gravity and $\alpha$ the tilt of the sea surface slope estimated from the modeled sea surface height. The force due to interaction with sea ice depends non-linearly on the sea ice concentration $f$, the sea ice strength $P$, a threshold $P_s$ above which the iceberg moves entirely with the sea ice, and the relative velocity of the iceberg with the sea ice:
\[
F_{SI} = \begin{cases} 
0 & \text{if } f \leq 15\%, \\
-(F_{AT} + F_W + F_C + F_{SS}) + \frac{dv_{si}}{dt} & \text{if } f \geq 90\% \text{ and } P \geq P_s, \\
\frac{1}{2}(\rho_{si}c_{si}A_{si})|\mathbf{v}_{si} - \mathbf{u}|(\mathbf{v}_{si} - \mathbf{u}) & \text{otherwise},
\end{cases}
\] (6)
where $c_{si}$ is the sea ice coefficient of resistance set to one, $A_{si}$ is the product of ice thickness by the iceberg width. The sea ice strength $P$ is a measure of the resistance of sea ice. It is defined by the standard formulation from Hibler (1979),
\[
P = P^*h \exp(-C(1 - f)),
\] (7)
where $h$ is the sea ice thickness. The empirical constants $P^*$ and $C$ are set to 20000 N m$^{-2}$ and 20. This formulation makes the ice strength strongly dependent on the sea ice concentration, while also allowing the ice to strengthen as the thickness $h$ increases.

Boundary conditions and stability criterion

When an iceberg hits the bottom, it remains stationary until it has either melted sufficiently to drift off or it is transported toward a deeper region by forces stronger than the frictional force. Based on experimental studies defining friction coefficients of a large ice block on a sand or gravel beach from Barker and Timco (2003), we used a static friction coefficient of 0.5 for grounded icebergs. When an iceberg is in a transition mode, moving from a deep region to a region shallower than its immersed part, we used a friction coefficient of 0.35. The icebergs are allowed to roll over, following the Weeks and Mellor (1978) stability criterion. Note that an iceberg is removed if any of its dimensions above the sea surface are less than 1 m.

Melting parameterizations

Among the mechanisms involved in the deterioration of icebergs, we consider only the most important ones: wave erosion, which is the primary source of melting (White et al. (1980) and Bigg et al. (1997)), lateral melting, and basal melting.

Wave erosion $V_{wave}$ parameterization is taken from Gladstone and Bigg (2001), who incorporated a dependency on the sea-surface temperature (SST) and the sea-ice concentration (m/day):
\[
V_{wave} = \left[\frac{1}{6}(T_w(1) + 2)\right]S_s\frac{1}{2}(1 + \cos(f^3\pi)),
\] (8)
where $T_w(1)$ is the SST, $f$ is the sea-ice concentration, and $S_s$ is the sea state derived from the wind speed. Thus, the wave erosion is damped in presence of sea ice.
Lateral melting $V_{lateral}$ is based on the parameterization of Kubat et al. (2007) over the iceberg draft. The empirical estimate of lateral melt rate (m/day) is:

$$V_{lateral} = \sum_{k=1}^{n} \left[ 7.62 \times 10^{-3} (\Delta T(k)) + 1.29 \times 10^{-3} (\Delta T(k))^2 \right],$$

(9)

where $\Delta T(k)$ is the difference between the sea-water temperature and the freezing-point temperature at the $k^{th}$ layer interface. The iceberg draft crosses the layers 1 to $n$ of the ocean model. The estimation of the basal turbulent melting rate $V_{basal}$ (in m/day) follows Weeks and Campbell (1973):

$$V_{basal} = 0.58 |v_w(n)| - u|^{0.8} \times \frac{T_w(n) - T(n)}{L^{0.2}}$$

(10)

where $v_w(n)$ is the water velocity at the iceberg base and $u$, the iceberg velocity. $L$ is the iceberg length, and $T(n)$ is the iceberg basal temperature. Similar to what it is done for sea-ice model basal-melting parameterizations, $T(n)$ is the local freezing-point temperature at the iceberg base and $T_w(n)$ is the local water temperature at the iceberg base.

Note that thermodynamics are considered only in in Paper II and Paper III.

**Wind forcing**

The iceberg-ice-ocean system is forced by wind fields from the 40-year European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) data (Uppala et al., 2005) for Paper I and Paper II with 1.125° grid cell resolution. For Paper III, the atmospheric parameters were taken from ERA-Interim, given on a 0.5° grid-cell resolution (Simmons et al., 2007).

**Ocean and sea ice forcing**

During the Global Ocean Data Assimilation Experiment (Dombrowsky et al., 2009, GODAE), ocean and sea ice forecasts over the North Atlantic and the Arctic Ocean were produced by the TOPAZ system (Bertino and Lisæter, 2008). Hence, the approach used here was to set-up a nested configuration of the Nansen/Mohn Sverdrup Center version of the Hybrid Coordinate Ocean Model (Bleck, 2002, HYCOM), where TOPAZ gives boundary conditions to a high resolution model (Barents model) covering the Barents and Kara Seas. The sea ice dynamics are based on the elastic-viscous-plastic rheology from Hunke and Dukowicz (1997). Thermodynamic fluxes over open water, ice-covered water and snow-covered ice are given in Drange and Simonsen (1996).

HYCOM has been tested on coastal and shelf ocean areas and found to reproduce the currents, salinities and temperatures well (Winther and Evensen, 2006). The Barents model has a 5 km horizontal grid cell resolution, 22 vertical hybrid layers and includes tides. The local Rossby radius of deformation is about 3 km (Løyning, 2001). As the model resolution is not sufficient to resolve mesoscale activity in the region, residual currents are underestimated. Nevertheless, the model has encouraging skills as shown in the KARBIAC project with a similar version (Bertino et al., 2007). For each study, the ocean sea ice model is forced by the same wind forcing as the iceberg model.
Validation data

For calibration and validation of iceberg drift, we used a dataset deployed during IDAP experiment (Spring, 1994). Icebergs from which information on their initial size was missing as well as those grounded during most of their recorded drift were excluded. Finally, we focused on data from 1990 where the time-averaged influence of the atmosphere and the ocean currents seemed to be the strongest, and thus most appropriate for model parameterization.

Data assimilation method

The data assimilation method used in this thesis is the Ensemble Kalman Filter (Evensen, 2009, EnKF). It is based on a Monte Carlo technique where error statistics are calculated from an ensemble of model states. It provides a spatial and temporal varying error covariance matrix which evolves according to model dynamics. It also allows for multivariate updates, and thus for the estimation of non-observed model parameters.

Summaries of papers

Paper I: Parameterization of an iceberg drift model in the Barents Sea

Paper I addresses the problem of parameter estimation to limit the errors in an ice-ocean-iceberg drift model for the Barents Sea. The system is forced by the ERA40 atmospheric reanalysis data from ECMWF and ocean and sea ice variables are from a nested configuration of HYCOM described in the previous section. The parameterization is tested and validated using four observed iceberg trajectories northwest of the Barents Sea from April to July 1990 (Spring, 1994). The model accuracy relies mainly on the quality of its forcing and the knowledge of the initial iceberg form and mass.

In the first part of the study, a measure of the sensitivity of the model to uncertainties in the mass of the iceberg and the forcing is given by jointly varying the iceberg mass and the form drag coefficients. As the iceberg mass increases, the optimal form drag coefficients increase linearly. A balance between the drag forces and the Coriolis force explains this behavior. It suggest that an optimal trajectory can be obtained by perturbing only the ocean and atmospheric form drags. The ratio between the optimal oceanic and atmospheric form drag coefficients is similar in all experiments, although there are large uncertainties on the iceberg geometries and errors in the forcing.

The second part of the study focuses on the impact of sea ice parameterization for the iceberg drift. Following Lichey and Hellmer (2001), there is a threshold value for the sea ice concentration and the sea ice strength from which the iceberg is moving entirely with the sea ice. We perturbed the threshold value of the sea ice strength, but the sea ice conditions East of Svalbard in winter 1990 were such that they exhibit no sensitivity.

Using optimal parameter values, the average distance from observation is: less than 20 km after two months of drift for the northernmost icebergs and less than 25 km after the first month and increasing rapidly to over 70 km thereafter for the two southernmost icebergs.
ERRATUM on Paper I publication

The Equation 6 presents a typographic error. The atmospheric force is written as $F_A$ while it should be $F_{AT}$.

The estimated iceberg masses are 10 times bigger than the values specified in the Table 1.

Paper II: Modeling dynamics and thermodynamics of icebergs in the Barents Sea from 1987 to 2005

A modeling study of iceberg drift characteristics in the Barents and Kara Seas based on a 19-year simulation is presented. In addition to the dynamics presented in Paper I, the effects of wave erosion, basal melting and lateral melting are included. An additional frictional term is also introduced in order to better simulate iceberg grounding, and the iceberg is allowed to roll over if it becomes unstable. Based on the estimates of iceberg production given by Dowdeswell et al. (2002), Hagen et al. (2003), Kubyshkin et al. (2006) and Dowdeswell et al. (2008), we define 11 calving sites representing the iceberg production per region. Initial iceberg sizes are generated randomly with a log-normal distribution following statistics on iceberg length width and height obtained during the 1988-1993 IDAP campaign (Spring, 1994). Seasonal variability of the calving rate is neglected despite indications of increased release from June through September (Kubyshkin et al., 2006), allowing an independent evaluation of the seasonal influence of the ocean currents, wind and sea ice on the iceberg characteristics and drift. Every day, icebergs are released according to a Poisson process, guarantying the independence of each calving event and a control on the average rate. The simulation starts in July 1985 and ends in December 2005. The first 1.5 years of the simulation are not used to allow sufficient time for the model to spin up.

Statistics of iceberg characteristics as a function of their origin are investigated. Maps of iceberg density and grounding locations complement sparse existing oceanographic and aerial field survey campaigns. Model results compare qualitatively well to the observations (Abramov and Tunik, 1996) and suggest preferential pathways and extensions from simulated calving sources. For example, icebergs originating from Franz Josef Land have the largest spread over the domain. Moreover the simulations show a seasonal cycle of the southernmost extent of the icebergs, even though the seasonality of the calving was not considered. Icebergs released in summer have the largest spreading. Hence, we suspect that introducing seasonal variability in the calving rate would intensify this latter pattern.

The interannual variability of the iceberg spread is analysed jointly with the iceberg extent. The latter shows a strong interannual variability. It is found that atmospheric forcing drives the extension of iceberg similarly to the sea ice extent (Sorteberg and Kvingedal, 2006). Anomalous northerly winds enhance the southward iceberg extension. They also produce a positive but delayed impact on the iceberg extent by limiting the inflow of Atlantic Water into the Barents Sea therefore reducing the heat content the following year and increasing the mean age of icebergs and thus their potential extension. This demonstrates that the thermodynamics also play an important role.

Finally, confidence in the system is reinforced as the model reproduces the observed extreme iceberg extension event, southeast of the Barents Sea in May 2003.
Paper III: Adaptive estimation of iceberg parameters using the Ensemble Kalman Filter

Paper I showed the ability of the model to reproduce iceberg drifts successfully, using optimal form drag coefficients. However, the model error grows in time due to inherent errors in the forcing, initial conditions and model parameterization. In this paper, we attempt to limit the error growth by correcting the icebergs position once per day using data assimilation. Such problem is non linear and requires advanced statistical methods.

In addition we intend to improve on the limitations regarding the estimate of the form drag coefficients in Paper I. First, the optimal values were estimated by minimizing the distance over the whole trajectory, although these parameters are expected to evolve with time due to the change in the shape of the iceberg. Second, the method employed was a classical Monte-Carlo approach with a regular sampling of initial parameters, which is computationally costly. The Ensemble Kalman Filter is an efficient data assimilation method based on Monte-Carlo that allows for multivariate update. Thus, the method corrects the position and gives the possibility to estimate non-observed parameters at a given time.

The system is composed of the iceberg drift model used in Paper II but the atmospheric forcing is updated to ERA-interim, the latest reanalysis product from ECMWF with 0.5° resolution. Observed trajectories of four icebergs from 1990, registered during the IDAP campaign (Spring, 1994), were assimilated in the model. The two northernmost simulated trajectories have a precision of 15 km and the two southernmost ones have a precision better than 25 km over the two months of drift. An analysis of the estimated form drag coefficients allows us to identify whether the shape was correct and when the forcing fields were failing.

Conclusions

This thesis has shown that a coupled ocean-sea ice-iceberg drift model system is able to describe physical processes, icebergs characteristics and trajectories in the Barents Sea. Its limitations were emphasized and the following conclusions are obtained:

- The proposed nested configuration has reasonably good prediction skills through the optimization of the iceberg mass and the ocean and atmospheric form drag coefficients. The best members reproduce observed iceberg trajectories for at least one month of drift with a precision better than 25 km.

- A 19 years simulation of iceberg trajectories in the Barents Sea complement and extend iceberg information from satellites and oceanographic and aerial field campaigns. Potential grounding location and preferential pathways from each of the principal calving sources are suggested.

- There is a seasonal variation of iceberg density in the Barents Sea, although the calving rate was set to be constant throughout the year.

- The interannual variability of the iceberg extent is strong and highly correlated with the sea ice area. This variability can be explained by two main mechanisms:
Northerly winds are the principal factor enhancing iceberg extent: higher than normal atmospheric mean sea level pressure over Svalbard region favors large iceberg extent.

Northerly winds also have a positive but delayed impact on the iceberg extent that shows the importance of the thermodynamics. They limit the inflow of Atlantic Water into the Barents Sea and therefore reduces the heat content during the following year, increasing the mean age of icebergs and thus their potential extension.

- The model is able to reproduce the extreme southernmost coverage of icebergs south east of the Barents Sea observed in May 2003.
- Part of the model errors can be compensated for by using the Ensemble Kalman Filter towards a pre-operational system for the region. It is shown that data assimilation clearly improves the prediction and has the advantage to give an estimation unknown model parameters.
- From the latter, we can also deduct information on the error in the iceberg mass and the forcing fields.

**Outlook**

The primary results have shown encouraging skills for the prediction of iceberg drift in the Barents Sea. The Barents Sea model is part of the operational TOPAZ system. The latter has undergone a significant upgrade in 2010, especially in the sea ice model formulation and in the data assimilation setup. It led to improvements both in the North Atlantic and in the high resolution Fram Strait model. In particular, the sea ice front is better described, and the heat balance is better represented. One can expect a better prediction in the Barents Sea model with a similar upgrade. Hence, those improvements are expected to be significant, especially for the thermodynamic part of the iceberg model.

In addition the latest version of the HYCOM model performs approximately twice as fast as the former one. It implies that running the model configuration with double resolution (∼2.5 km, i.e. eddy permitting) can be achieved at a four times the computational cost of the current model version. This model upgrade would impact the iceberg dynamics. Running an eddy resolving model operationally (∼750 m horizontal grid cell resolution) is today still out of reach for such a large domain.

Forecasting capabilities can be improved by including data assimilation in the ocean model. EnKF is advised for assimilation of sea ice concentration (Lisæter et al., 2003), but too costly to be applied in a regional model. Nevertheless, sea surface temperature could be assimilated using Ensemble Optimal Interpolation, a method that has yielded good results in regional models (Evensen, 2003), and is computationally less expensive.

The effect of waves can be important on icebergs. In the current model implementation, the wave drag is included implicitly through the atmospheric drag force. The Barents Sea is a small and relatively closed basin. It is unclear how important swells are compared to the wind sea. Nevertheless, separating wave drag from atmospheric form drag may give more precision in the sources of errors and would help to give a more precise parameterization. Some icebergs are large enough to feel ocean swell since the swell wavelength is on the order of the iceberg...
lateral dimensions (~100-300 m). It is a point to consider in future studies, which would involve wave input from a wave model. With realistic wave information, the wave erosion could include calving process as suggested by Kubat et al. (2007).

Despite the model improvement, the forcing fields are not perfect. An approach to complement information from the forcing fields is to perturb them randomly. Such a perturbation would account for errors in the wind and ocean current direction as well as intensity. It would be easier to relate the evolution of the form drag parameters to the change in the iceberg shape.

A more advanced but computationally more costly approach is to use forcing from an ensemble run: for example, ensemble runs from the TOPAZ and ECMWF systems into a high resolution Ensemble Barents Sea system. The TOPAZ system provides a 10 days forecast of ocean and sea ice parameters using ECMWF 10 days forecast of atmospheric fields for the North Atlantic and Arctic Ocean. Hence, based on the approach described in Paper III, it is possible to provide a forecasting system of iceberg trajectory 10 days ahead.

Figure 3: Iceberg north of Kangerlussuaq fjord. Håkon Mosby cruise, September 2007.
Bibliography


Barnes, P., Lien, R., 1988. Icebergs rework shelf sediments to 500 m off Antarctica. Geology 16 (12), 1130.


Academy of Sciences Ice Cap, Severnaya Zemlya, Russian High Arctic. Journal of geophysical research 107 (B4).


Parameterization of an iceberg drift model in the Barents Sea

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Parameterization of an Iceberg Drift Model in the Barents Sea

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ABSTRACT

The problem of parameter estimation is examined for an iceberg drift model of the Barents Sea. The model is forced by atmospheric reanalysis data from ECMWF and ocean and sea ice variables from the Hybrid Coordinate Ocean Model (HYCOM). The model is compared with four observed iceberg trajectories from April to July 1990. The first part of the study focuses on the forces that have the strongest impact on the iceberg trajectories, namely, the oceanic, atmospheric, and Coriolis forces. The oceanic and atmospheric form drag coefficients are optimized for three different iceberg geometries. As the iceberg mass increases, the optimal form drag coefficients increase linearly. A simple balance between the drag forces and the Coriolis force explains this behavior. The ratio between the oceanic and atmospheric form drag coefficients is similar in all experiments, although there are large uncertainties on the iceberg geometries. Two iceberg trajectory simulations have precisions better than 20 km during two months of drift. The trajectory error for the two other simulations is less than 25 km during the first month of drift but increases rapidly to over 70 km afterward. The second part of the study focuses on the sea ice parameterization. The sea ice conditions east of Svalbard in winter 1990 were too mild to exhibit any sensitivity to the sea ice parameters.

1. Introduction

Icebergs in the Barents Sea present a threat for navigation and offshore installations. The main source of icebergs in the Barents Sea are the Franz Josef Land archipelago glaciers, particularly the Renown glacier on Vilzcek Land (Spring 1994). The Svalbard archipelago is the secondary source, particularly the Stonebreen glacier on Edgeoya. A smaller contribution of icebergs comes from glaciers of the northern tip of Novaya Zemlya. Observation campaigns during the Ice Data Acquisition Program (IDAP; Spring 1994) from 1988 to 1993 reported that icebergs have an average size of 91 m $\times$ 64 m $\times$ 15 m and a maximum observed size of 320 m $\times$ 279 m $\times$ 40 m. Although a great proportion of the icebergs stays and melts close to the calving area, icebergs can be found in more than half of the Barents Sea. Further, based on aerial surveys covering the period 1933–90, Abramov (1992) studied the seasonal cycle of the southern extension of the iceberg distribution. The southernmost extension is found to occur during January–May and the northernmost extension occurs during September–October. The interannual variability of the quantity and the geographical distribution of the icebergs depend on their calving rate and the wind. Predominantly northerly and northeasterly winds favor the southern extension of the icebergs.

Several studies in the Labrador Sea have successfully modeled the trajectories of icebergs by using forcing derived from observations (Smith and Banke 1983; Sodhi and El-Tahan 1980). Observations of ocean currents in the Barents Sea are largely insufficient for representing their complex dynamics. The lack of data has motivated us to use a coupled ice–ocean model to force the iceberg drift model. Bigg et al. (1997) produced a climatology of modeled iceberg trajectories over the Arctic by using a three-dimensional ocean circulation model and sea ice observations from Bourke and Garrett (1987). Lichey and Hellmer (2001, hereafter LH01) modeled the iceberg drift under the influence of sea ice in the Weddell Sea with a coupled ocean–sea ice circulation model. They focused on the relative importance of atmosphere, ocean, and sea ice forces that act on the iceberg drift in the Weddell Sea.

Following LH01, we adapt the model to the Barents Sea icebergs, which are much smaller than those in Antarctica. This has a consequence for drag force, because...
the form drag has a larger impact on trajectories of small icebergs. The model is configured to receive boundary conditions from the Towards an Operational Prediction System for the North Atlantic European Coastal Zones (TOPAZ) system (Bertino and Lisæter 2008). We study the parameterization of the iceberg drift model and focus on the dynamical forces that have the strongest impact on the iceberg trajectories as a first step toward forecasting the drift of icebergs.

The outline of this paper is as follows: section 2 provides a brief description of the model. Section 3 describes the dataset from the IDAP campaign that is used to evaluate our model system. Section 4 focuses on the sensitivity of the model to the mass of the iceberg and the ocean and atmospheric form drag coefficients. Section 5 describes the experiment testing the sea ice parameterization proposed by LH01. The work is discussed and summarized in section 6.

2. The model system

The iceberg drift model relies on the forcing components and their parameterization. Section 2a describes each forcing component: namely the ocean, atmosphere, and sea ice components. The implementation of the iceberg drift model is presented in section 2b.

a. Ocean and sea ice model

To model the ocean and sea ice, a version of the Hybrid Coordinate Ocean Model (HYCOM; Bleck 2002) is coupled to a dynamic sea ice model based on the elastic–viscous–plastic rheology by Hunke and Dukowicz (1997). The thermodynamic fluxes over open water, ice-covered water, and snow-covered ice are given in Drange and Simonsen (1996).

We used a one-way nested configuration that allows for high resolution and realistic boundary conditions in order to obtain a reasonable representation of the Barents Sea current system at limited computational cost. The large-scale model is a version of the TOPAZ3 forecasting system that covers the Atlantic and the Arctic Ocean and is run without data assimilation (Bertino and Lisæter 2008). The large-scale model provides boundary conditions to a regional model of the Barents and Kara Seas (Barents model). When nesting the slow boundary variables (i.e., baroclinic velocities, temperature, salinity, and layer interfaces), a simple relaxation technique is used. For the barotropic components (velocities and pressure), the boundary conditions are computed while taking into consideration both the waves propagating into the regional model from external solution and the waves propagating out through the boundary of the regional model (Browning and Kreiss 1982).

TOPAZ3 has an 11-km horizontal resolution in the Arctic with smooth transition toward lower resolution at the equator. Barents has a 4.5–5.8-km horizontal resolution. The model grids are created by using the conformal mapping algorithm of Bentsen et al. (1999). Barents was initialized from the Generalized Digital Environmental Model, version 3 (GDEM3) climatology (Teague et al. 1990) and spun up for four years. The atmospheric forcing is from 6-hourly 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) data (see Uppala et al. 2005). The regional model includes tides, which are specified as a barotropic forcing at the open boundaries. The tidal data are taken from the Finite Element Solution tidal atlas 2004 (FES2004; Lyard et al. 2006).

An overview of the mean ocean currents for our focus area of the top 60 m during the period of study is shown in Fig. 1. With a model that does not resolve mesoscale activity for the region and the inaccuracy of its boundary conditions, it is likely that residual currents are underestimated. Because the long-term average currents are weak in this region, it should not have a strong impact in this study. In addition, the sea ice edge is sharper and narrower in the sea ice model than in observations. Thus, the southern extension of low-concentrated sea ice is often underestimated in the model, but this should have minor impact on the iceberg trajectories.

b. Iceberg model

The iceberg model is based on LH01. The iceberg acceleration is proportional to the forces from the atmosphere \(F_{AT}\), the water drag \(F_W\), the Coriolis force \(F_C\); the force resulting from the sea surface slope \(F_{SS}\); and the force resulting from interaction with the sea ice cover \(F_{SI}\):

\[
M \frac{du}{dt} = F_{AT} + F_W + F_C + F_{SS} + F_{SI}.
\]

where \(M\) is the iceberg mass and \(u\) is the iceberg velocity.

The atmospheric and oceanic forces act on the cross-sectional area above and below the water line, respectively, in a vertical plane as form drag and a horizontal plane as surface drag (Smith and Banke 1983). The atmospheric force is

\[
F_{AT} = \left[ \frac{1}{2} (\rho_w c_w A_{vw}) + (\rho_a c_{aw} A_{aw}) \right] |v_a - u|(v_a - u). \]

The oceanic force \(F_W\) is defined by the same quadratic drag law as \(F_{AT}\) for each ocean model layer through the depth of the iceberg:
In Eqs. (2) and (3), \( v_a \) and \( v_w(i) \) are the air and oceanic velocities, respectively; the index \( i \) is the ocean layer number; and \( n \) is the number of ocean layers in contact with the iceberg. The terms \( A_ya \) and \( A_yw(i) \) are the vertical cross-sectional areas in the air and water, respectively. The air and water densities are \( r_a \) and \( r_w \) are the air and water densities, respectively; \( c_d \) and \( c_w \) are the form drag coefficients; \( c_{da} \) and \( c_{dw} \) are the skin drag coefficients, set to 0.0022 and 0.0055, respectively, as in the sea ice model we use; and \( A_{ha} \) and \( A_{hw}(n) \) are the horizontal areas of the iceberg in contact with the air and ocean layer \( n \), respectively. If the surface drag area exceeds about 250 times the sail area of the iceberg, the surface drag becomes larger than the form drags (Smith and Banke 1983). For the relatively small lateral dimensions of Barents Sea icebergs, the form drag will be the most important factor.

The wind is assumed to be constant with height above sea level. However, the ocean currents vary with depth, as given by the ocean model. By tuning the air form drag coefficient, we can partly compensate for errors in the sail area and the variation of wind speed with height. In the same way, by tuning the water form drag coefficient, we partly compensate for errors in the keel area and inaccuracies in the ocean currents. Note that we do not vary \( c_w \) with depth, though internal wave drag might occur in the pycnocline (Smith 1993).

The third force acting on the iceberg is the Coriolis force,

\[
F_C = 2M\Omega \sin \phi \mathbf{k} \times \mathbf{u},
\]

where \( \Omega \) is the angular velocity of the earth, \( \phi \) is the latitude, \( \mathbf{k} \) is the unit vector perpendicular to the earth’s surface, and \( \mathbf{u} \) is the iceberg velocity. The force resulting from the sea surface slope is

\[
F_{SS} = -Mg \sin \alpha,
\]

where \( g \) is the acceleration due to gravity and \( \alpha \) is the tilt of the sea surface slope estimated from the modeled sea surface height. The final force acting on the iceberg is due to interaction with sea ice. In LH01, the sea ice force depends nonlinearly on the sea ice concentration \( f \); the sea ice strength \( P \); a threshold \( P_s \), above which the iceberg moves entirely with the sea ice; and the relative velocity of the iceberg with the sea ice,
where $c_{si}$ is the sea ice coefficient of resistance set to 1, as in LH01 and $A_{si}$ is the product of the ice thickness and the iceberg width. The sea ice strength $P$ is a measure of the resistance of sea ice. It is defined by the standard formulation from Hibler (1979):

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

(7)

where $f$ is the sea ice concentration and $h$ is the sea ice thickness. The empirical constants $P^*$ and $C$ are set to 20 000 N m$^{-2}$ and 20, respectively. This formulation makes the ice strength strongly dependent on the ice concentration and also allows the ice to strengthen as the thickness $h$ increases. For regions of melting sea ice, $P^*$ should ideally be reduced, but we did not change it for simplicity.

Previous iceberg drift modeling studies Sodhi and El-Tahan (1980) and Smith (1993) include a term proportional to the iceberg mass to simulate acceleration of water entrained in the wake of the iceberg (i.e., track of the waves left by the iceberg moving through the water). Here, this force is taken into consideration implicitly in the experiment by testing three different iceberg masses (section 4).

Smith (1993) discussed the effect of a wave radiation force but did not include it explicitly in his dynamical model, whereas Bigg et al. (1997) did. Because the form of the wave radiation force in Bigg et al. (1997) is proportional to Eq. (2), wind waves are present implicitly in the atmospheric force parameterization by selecting higher values for the air drag coefficient. This parameterization will not capture the effect of swells. Note that during the first month of the period of study, effects of waves are damped by the presence of sea ice.

To force the model, we used daily averaged ocean current fields from the HYCOM system described in section 2a and daily averaged wind fields from ERA-40. The tidal oscillations could have been represented by using hourly input, but they have been excluded to limit the size of the forcing dataset. We thus simulate residual iceberg drifts.

3. Iceberg observations

We focus on four specific icebergs on which Argos buoys were placed during late April 1990 under the IDAP program (Spring 1994). The available dataset consists of 17 reliable observed trajectories over the following three years: 1988 (seven), 1989 (five), and 1990 (six). All the icebergs were located in the northwestern region of the Barents Sea. During the years 1988 and 1989, the observed icebergs were mostly driven by inertial tidal oscillations and mesoscale activity, with the latter not being represented in the model. Furthermore, little information about the size of icebergs in 1988 was available. Therefore, we focus on data from 1990 where the time-averaged influence of the atmosphere and the ocean currents seemed to be the strongest. The observed sizes and recorded drifts of the icebergs are given in Table 1. The iceberg shapes are unknown, leaving a wide range of uncertainty in the mass and form drag coefficients of each iceberg. The 1990 iceberg trajectories are presented in Fig. 2. All the icebergs are located in the western Barents Sea. Hereafter, each iceberg is referred to by its Argos buoy number. Two icebergs, 1872 and 7085, were grounded for most of their recorded drift and were excluded from our study. The four remaining icebergs were subject to different regimes: initially within tightly packed sea ice and then in contact with open water. None of these icebergs was grounded. The recorded drift lasted 65, 70, 82, and 67 days for icebergs 7086, 7087, 7088, and 7089, respectively, starting one day earlier for the two southernmost icebergs (7086 and 7087).

Trajectories of the four icebergs are very similar during the first month, but they differ toward the end of the period. Each iceberg starts by following a clockwise loop and then moves mainly southwestward until late May. The two northernmost icebergs have comparable trajectories until the end of the recorded drift. Both follow a southeastward trajectory until 21 and 19 June (for 7088 and 7089, respectively) and then move northwestward. The two southernmost icebergs have more complex trajectories during the second month (see Fig. 3). Iceberg 7087 moves in a chaotic manner until 3 June and then moves northward until 14 June. It then follows an anticyclonic loop from 14 to 24 June and moves northwestward until the end of the record, 29 June. Iceberg 7086 follows two cyclonic loops from 1 to 13 June, then moves southward until 18 June, and moves northeastward until the end of its recorded drift. The diameter

<table>
<thead>
<tr>
<th>Buoy</th>
<th>$L$</th>
<th>$W$</th>
<th>$H$</th>
<th>Mass*</th>
<th>Initial position</th>
<th>Drifting period</th>
<th>Final position</th>
</tr>
</thead>
<tbody>
<tr>
<td>7086</td>
<td>90</td>
<td>60</td>
<td>10</td>
<td>24300</td>
<td>78.11°N, 31.90°W</td>
<td>25 Apr–30 Jul</td>
<td>75.99°N, 25.13°W</td>
</tr>
<tr>
<td>7087</td>
<td>63</td>
<td>56</td>
<td>10</td>
<td>15876</td>
<td>78.07°N, 31.46°W</td>
<td>25 Apr–04 Jul</td>
<td>76.75°N, 26.62°W</td>
</tr>
<tr>
<td>7088</td>
<td>95</td>
<td>80</td>
<td>20</td>
<td>68400</td>
<td>78.92°N, 34.17°W</td>
<td>26 Apr–18 Jul</td>
<td>76.85°N, 29.68°W</td>
</tr>
<tr>
<td>7089</td>
<td>95</td>
<td>90</td>
<td>15</td>
<td>57712.5</td>
<td>79.03°N, 34.81°W</td>
<td>26 Apr–02 Jul</td>
<td>76.66°N, 32.10°W</td>
</tr>
</tbody>
</table>

* The mass is estimated for an idealized rectangular tabular iceberg with a depth of 4 times its height (Smith and Banke 1983).
and orbital period of the cyclonic loops correspond to typical eddy sizes in this area (Løyning 2001).

To select trajectories when the icebergs were within high concentration of sea ice, we used daily averaged sea ice maps from Special Sensor Microwave Imager (SSM/I) obtained with the Norwegian Remote Sensing Experiment (NORSEX) algorithm (Svendsen et al. 1983). The icebergs remained in areas where sea ice concentration was higher than 90% from their initial location until 17, 16, 20, and 24 May for icebergs 7086, 7087, 7088, and 7089, respectively.

4. Experiment 1: Sensitivity to iceberg mass and form drag coefficients

In this section, we measure the sensitivity of the model to uncertainties in the mass of the iceberg and the forcing by jointly varying the mass $M$, the ocean form drag coefficient $c_w$, and the atmospheric form drag coefficient $c_a$.

a. Setup

There is no information available regarding the shape of the icebergs being tracked. However, from statistics based on aerial stereo photography of 90 icebergs observed during 1990, most of the icebergs were tilting tabular (32%), pinnacled (31%), or tabular (16.7%; Spring 1994).

A tilting tabular iceberg is a tabular iceberg with its top surface no longer parallel to the ocean or sea ice surface. A pinnacled iceberg is an iceberg with a central spire or pyramid, with one or more spires. The only information we have from each iceberg is its initial length, width, and height. We chose to consider three different possible iceberg shapes in our simulations that would change the icebergs mass $M$ but not the cross-sectional areas $A_{sa}$ and $A_{sw}(i)$ in contact with the winds and ocean currents, respectively. We assume a tabular iceberg with the shape of a rectangular prism and a sail area equal to the product of the observed width and height. It will be referred to an iceberg with mass 100%$M$. An iceberg with a triangular prism shape, with the same $A_{sa}$ and $A_{sw}(i)$ as the tabular iceberg has a mass of 50%$M$. Finally, the last experiments with icebergs defined as 35%$M$ could correspond to an eroded iceberg with one large vertical wall. This way of changing the iceberg mass is a crude way of taking into account the geometrical configuration of the iceberg.
The values of $c_a$ and $c_w$ are sampled, as in Table 2. Because we focus on the ocean and atmospheric drag in this experiment, the sea ice force is defined as a regular drag relationship independent of $P$:

$$\mathbf{F}_{SI} = \begin{cases} 0 & \text{if } f \leq 15\%, \\ \frac{1}{2} \left( \rho_s c_s A_{si} \right) \mathbf{\nu}_si - \mathbf{u} (\mathbf{v}_si - \mathbf{u}), & \text{otherwise.} \end{cases}$$

(8)

### MEASURE OF THE ERROR IN THE MODELED ICEBERG DRIFT TRAJECTORIES

The performance of each experiment is analyzed by measuring the geographical distance between modeled and observed iceberg drift and by counting the successful simulations for different values of the parameters. The errors in the distance should increase with time as a result of unresolved ocean and atmospheric circulation processes and parameterization errors. Consequently, we define a linearly decreasing weighting function of time so that the earlier part of the trajectory counts more than the later part:

$$\Delta d = \frac{1}{2m(m+1)} \sum_{t=1}^{m} (m-t) \epsilon(t),$$

(9)

where $m$ is the number of time iterations, $t$ is the time iteration, and $\epsilon(t)$ is the geographical distance between the observed iceberg and the modeled iceberg at time $t$.

### TABLE 2. Parameter ranges for expt 1. Tested values for the mass, the atmospheric form drag coefficient $c_a$, and the oceanic form drag coefficient $c_w$ for the four icebers with sea ice force defined as in Eq. (8).

<table>
<thead>
<tr>
<th>Mass (%)</th>
<th>100, 50, and 35</th>
</tr>
</thead>
<tbody>
<tr>
<td>$c_a$</td>
<td>from 0.1 to 2.0; every 0.1</td>
</tr>
<tr>
<td>$c_w$</td>
<td>from 0.05 to 1.0; every 0.05</td>
</tr>
</tbody>
</table>

We apply this method over the whole drifting period to evaluate the modeled trajectories of the four icebers and store the best ones into classes of distances.

### b. Results

Sections 1 and 2 present a qualitative comparison between modeled and observed trajectories for the “full” mass experiment, 100% $M$. We obtained similar results with the 50% $M$ and 35% $M$ experiments (not shown). Further, section 3 highlights the relation between the optimal form drag coefficients and the mass. For the two northernmost icebers, 7088 and 7089, the best ensemble runs have mean weighted distances less than 20 km. For the two southernmost icebers, 7086 and 7087, the best ensemble runs have mean weighted distances less than 30 and 35 km, respectively.

1) TRAJECTORIES

The best trajectories for each iceberg are shown in Fig. 4. The two northernmost modeled icebers (7088 and 7089) closely follow the observations over the entire drifting period. The best members remain, on average, within 20 km of the observed icebers over the entire recorded drift.

For the two southernmost icebers (7086 and 7087), modeled and observed trajectories diverge after the first month. Nevertheless, we point out that even if the modeled trajectory did not reproduce features from 26 May to 26 June, the drift patterns in the last part of the trajectory (3 days for 7086 and 5 days for 7087) are parallel to the observed ones. The best members remain within 25 km, on average, over the first half of the trajectory of the observed icebers and within 45 km over the entire trajectory.

2) RELATIVE CONTRIBUTION OF FORCES

An evaluation of the contribution of each forcing component for each iceberg drift period highlights the importance of the three main forcing components: the wind forcing, the oceanic forcing over the drift of the iceberg, and the Coriolis force (see Fig. 5). The forces due to sea ice and sea surface slope are negligible for this simulation. Figure 6 shows how the position errors and the spread of the successful runs increased with time.
The exception is the period from 27 May to 5 June, the time when the two southernmost icebergs (7086 and 7087) were subject to forces that the model does not represent.

3) Optimal Mass and Form Drag Coefficients

Tuning the ocean and air form drag coefficients allows for a calibration of the modeled iceberg trajectories to the observed ones. These coefficients represent uncertainties related to the shape of the iceberg and uncertainties in the forcing fields: namely, the ocean and sea ice model and atmospheric data. Smith (1993) modeled short (from 11 to 73 h) iceberg trajectories off the east coast of Canada by using wind and current from nearby observations. He found little improvement by tuning the iceberg parameters. In our case, the parameterization is more critical, because trajectories are simulated over a much longer period and the modeled forcing fields have larger errors.

The successful model runs are presented in Fig. 7 as scatterplots of drag coefficients for three different iceberg mass experiments. For each experiment, there is an optimal range of values ($c_w$, $c_a$); as the mass increases, so do the optimal ($c_w$, $c_a$) values. In addition, the range of optimal values is broader for bigger icebergs. This is because a small change in $c_w$ or $c_a$ has a stronger impact on small icebergs and the effect of Coriolis force is larger on large icebergs. By linking the three clouds together with a linear regression, we found slopes of 0.35, 0.32, 0.31, and 0.29 for icebergs 7089, 7088, 7087, and 7086, respectively. These values represent the optimal ratio...
between \( c_w \) and \( c_a \) in our model, which is independent of the mass. The relation between the optimal \( c_w \) and \( c_a \) and the iceberg mass \( M \) is apparently linear. A linear relation is expected, considering the very small impact of \( F_{SS} \) and \( F_{SI} \) on the iceberg motions in our simulation. One can therefore assume the simplified force balance

\[ F_A + F_W = -F_C, \]  

(10)

where \( F_A \) and \( F_W \) are the atmospheric and oceanic forces, respectively, with each neglecting the horizontal friction. Considering the mean ocean velocity \( v_w \) across the iceberg drift and \( A_{sv} \) as the vertical area of the iceberg, then

\[ F_A = \frac{1}{2} (\rho_a c_a A_{sv}) |v_a - u| (v_a - u) \]  

and

\[ F_W = \frac{1}{2} (\rho_w c_w A_{sv}) |v_w - u| (v_w - u). \]  

(11)

(12)

If we assume known velocities at time \( t \) and a linear relation between \( c_w \) and \( c_a \), we obtain the following relation between the mass \( M \) and \( c_{a}: \)

\[ (c_a + K c_a) D = M \hat{C}, \]  

(13)

where \( K \) is constant, \( D \) is the drag forces divided by the drag coefficients, and \( \hat{C} \) is the Coriolis force divided by the mass.

Under these simplified assumptions, the relationships between, \( M, c_w, \) and \( c_a \) are linear. Also note that, as \( M \) increases, \( c_a \) and \( c_w \) increase accordingly, which explains the clustering of high (low) \( c_a \) and \( c_w \) values with high (low) mass \( M \).

5. Experiment 2: Influence of the sea ice strength

In this section, we evaluate whether the parameterization of the sea ice forcing (e.g., for the large Weddell...
Sea icebergs in LH01 is applicable to the much smaller icebergs in the Barents Sea.

\section*{a. Setup}

Calculated sea ice strength from the sea ice model shows that, during late winter 1990, the sea ice strength was rather weak in the western Barents Sea, less than 6500 N m\(^{-2}\) most of the time from late April to July. This motivated the choice of the sampling of \(P_s\) shown in Table 3. Note that the maximum sea ice strengths along all the modeled trajectories are 3721.5, 3779.9, 5980.33, and 6530.0 N m\(^{-2}\) for icebergs 7086, 7087, 7088, and 7089, respectively. We consider here only tabular icebergs with 100\%\(M\). Infinity corresponds to the experiments when the \(P_s\) parameterization is not used [i.e., \(F_{SI}\) is defined by Eq. (8)].

For \(P_s = 1000\) N m\(^{-1}\), the minimum ice thickness able to lock an iceberg in the sea ice is 0.37 m when the sea ice concentration \(f\) is 90\% and 0.05 m when \(f\) is 100\%, which is less than the typical ice thickness encountered in the Barents Sea. Hence, with \(P_s = 1000\) N m\(^{-1}\), the icebergs will tend to follow the movement of sea ice. For \(P_s = 6500\) N m\(^{-1}\), the corresponding minimum ice thickness is between 2.4 and 0.325 m. In this case, icebergs will at times move with sea ice, but other times not. As in LH01, shear and bending forces that occur because of iceberg tilt and rotation are neglected, which means that sea ice forces may be underestimated.

\section*{Measure of the Error in the Modeled Iceberg Trajectories}

Because the later parts of the trajectories are in low-concentrated sea ice, the parameterization of \(P_s\) has little influence on the icebergs. We therefore restrict the optimization to the first 10 days of drift. Only trajectories with a mean distance to the observations of less than 15 km are retained. To focus on the effect of the \(P_s\) parameterization, we need to remove the bias in the
modeled trajectory that comes from uncertainties of $c_w$ and $c_a$. Consequently, we make a first selection of trajectories keeping only the successful ($c_w$, $c_a$) couples for each iceberg, as obtained from experiment 1.

b. Results

From the histograms (Fig. 8), the proportion of representative trajectories with no $P_s$ parameterization [i.e., $F_{SI}$ defined in Eq. (8)] is comparable to the others for the two southernmost icebergs. For the two northernmost icebergs, the best simulations are given by experiments with no $P_s$ parameterization and experiments with $P_s = 6500 \text{ N m}^{-1}$. Thus, there is no evidence that the $P_s$ parameterization is significantly better for the four icebergs studied. However, sea ice characteristics have strong interannual variability in the area. The year 1990 was not particularly severe, and the sea ice can be thicker during other years. Though the proportion of multiyear ice is usually small in the western Barents Sea, it can be important in the eastern Barents Sea. Therefore, the sea ice force, as specified in Eq. (6), should not be excluded for an iceberg drift model in the Barents and Kara Seas. According to our experiments, a $P_s$ value of 13 000 N m$^{-1}$, as in LH01, remains acceptable despite the fact that Barents Sea icebergs are about 10 times smaller than the ones in the Weddell Sea. Thus, the iceberg would be locked into sea ice only for more severe sea ice conditions than in our experiments.

6. Discussion and conclusions

This study focuses on the parameter estimation of an iceberg drift model in the Barents Sea. Four icebergs drifting southeast of Svalbard observed for more than two months in 1990 have been modeled. The simulations have a precision better than 20 km during the two months of drift of the two northernmost icebergs and better than 25 km during the first month of drift for the two others. For the latter two icebergs, although they are only about 100 km south of the two northernmost at the
beginning, unpredicted circumstances such as melting, unresolved mesoscale features in the forcing, or a sudden change of shape and mass make the model error increase rapidly to 70 km in the second month.

In the first experiment, we sampled the iceberg masses and the atmospheric and ocean drag coefficients. The cross-sectional area of the iceberg was kept constant, whereas the ice mass was increased as a crude way of changing the geometry of the iceberg.

We found a common ratio of $c_w$ to $c_a$ between 0.29 and 0.35, independently of the chosen geometry. The ratio between the optimal atmospheric and ocean drag components is similar across the experiments, though we have little information about the geometry of the icebergs. Furthermore, the fact that the two southernmost icebergs have a much broader range of optimal $c_w$ and $c_a$ values highlights either some uncertainty in the chosen geometric configuration over the first month of integration, or it indicates that melting has been important for those two icebergs toward the end of the iceberg drift period.

As the iceberg mass increases, so do the optimal atmospheric and ocean form drag coefficients. A simple force balance between drag forces and the Coriolis force explains this linear behavior. The relation between optimal drag coefficients and the iceberg mass highlights the importance of the iceberg geometric configurations when modeling the trajectories. A value of $(c_w, c_a)$ that provides the best trajectory for an iceberg of mass $M$ may yield to a significantly different trajectory if the mass is 50% $M$, given that their cross-sectional area is the same. In a similar manner, the initial determination of the mass of the iceberg is important for successful modeling of their trajectories. In further studies, the ocean and atmospheric form drag should be a function of the iceberg mass.

It must be stressed that the optimal form drag factors calculated here do not correspond to actual form drag factors, because they depend on the mass of the iceberg. Thus, the optimal values reported here might be higher or lower than the actual values of the form drag factors. The optimized drag factors provide us with an interesting twist to our study. Can we estimate the iceberg mass based on the actual drag coefficients and the optimized drag values retrieved in a manner similar to this study? This approach is theoretically feasible, although we question the practicality of this result, if for nothing else, then for the difficulty in calculating “correct” drag coefficients for a given iceberg.

The second experiment focused on the sea ice force parameterization defined in Eq. (6). We looked at the sensitivity of the threshold value $P_s$ for our region of interest. The sea ice characteristics east of Svalbard in April–May 1990 did not have a direct impact on the iceberg trajectories. The ice flux from the central Arctic into the Barents Sea has strong interannual variations; therefore, it is possible that the parameterization may give better results for more severe ice years in the eastern Barents Sea.

In this study, we have tried to optimize some of the model parameters based on the relative importance of the forcing contributions. As in any model study, we depend on the accuracy of the forcing, which in this case is limited mainly by its spatial resolution. The bulk parameterization of the form and skin drag may also be too simple. Other force contributions, such as wake drag and wave radiation stress, are included implicitly through the variation of iceberg mass and drag force, albeit in a simplified form. A better parameterization of these mechanisms may improve our results.

Over a 2-month period, two of the icebergs are modeled with good accuracy, which illustrates that there is some skill in the system. Based on observed distributions of iceberg dimension, it is possible to estimate initial iceberg mass and corresponding form drag coefficients. In addition, including the thermodynamics will

<table>
<thead>
<tr>
<th>$P_s$ (N m$^{-1}$)</th>
<th>1000, 3000, 4500, 6500, and $\infty$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$c_w$</td>
<td>from 0.1 to 2.0; every 0.1</td>
</tr>
<tr>
<td>$c_a$</td>
<td>from 0.05 to 1.0; every 0.05</td>
</tr>
</tbody>
</table>
allow for modeling long-term Barents Sea iceberg trajectories under different climate regimes. The optimal values of the iceberg mass and the drag coefficients have another property, which is appealing from a forecasting perspective. For ensemble data assimilation techniques such as the ensemble Kalman filter (Evensen 2007), the ensemble spread should be on the order of the forecast error, a property that the “best model runs” in Fig. 6 appear to have for icebergs 7088 and 7089. It is too early to speculate on the success of an ensemble-based iceberg forecasting system, but the results presented here are nonetheless encouraging.

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REFERENCES

Paper II

Modeling dynamics and thermodynamics of icebergs in the Barents Sea from 1987 to 2005

Keghouche I., Counillon, F. and Bertino, L.

Journal of Geophysical Research, revised and re-submitted
Modeling dynamics and thermodynamics of icebergs in the Barents Sea from 1987 to 2005

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**Abstract**

A modeling study of iceberg drift characteristics in the Barents and Kara Seas for the period 1987-2005 is presented. Maps of iceberg density and potential locations subject to grounding complement sparse existing oceanographic and aerial field survey campaigns. The model suggests preferential pathways from the most important calving sources. Icebergs originating from Franz Josef Land have the largest spread over the domain. Simulations confirm the previously observed seasonal cycle of the southernmost extent of the icebergs. Strong interannual variability of the iceberg extent with a weak decreasing trend occurs, similar to the observed sea-ice extent. Analysis of the atmospheric forcing reveals that years with anomalous northerly winds enhance the southward iceberg extension. Northerly winds also have a positive delayed impact on the iceberg extent. They limit the inflow of Atlantic Water into the Barents Sea and therefore its heat content the following year, increasing the mean age of icebergs and thus their potential extension. Finally, the model is able to reproduce the observed extreme iceberg extension southeast of the Barents Sea in May 2003.

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1 Introduction

Icebergs in the Barents Sea present a threat for navigation and offshore installations. The threat could be in a direct form through a collision or via scouring of the ocean floor from grounded icebergs. The main source of icebergs in the Barents Sea is the Svalbard Archipelago and especially the Austfonna ice cap (Dowdeswell et al., 2008). The Franz Josef Land glaciers and in particular the Renown Glacier on Wilczek Land (Kubyshkin et al., 2006) are secondary sources. A smaller contribution of icebergs come from glaciers of the northern tip of Novaya Zemlya.

During the last century, icebergs in the Barents Sea have been observed through oceanographic fields campaigns, ice reconnaissance flights and satellite observations. Observational campaigns under the Ice Data Acquisition Program (Spring, 1994, IDAP) from 1988 to 1993 reported that their average and maximum size above the sea surface was: 91 m length \times 64 m width \times 15 m height and 320 m \times 279 m \times 40 m, respectively. Although a great proportion remain and melt close to the calving area, icebergs were found as far south as 67.2°N in the Barents Sea during the summer of 1929 (Abramov and Tunik, 1996). Further, based on aerial surveys covering the period 1933-1990, Abramov (1992) reported the variability of the iceberg distribution within the Barents Sea. The seasonal cycle of the extension of the iceberg distribution, with the southernmost extension occurring in January-May and the northernmost extension during September-October. The interannual variability of the quantity and the geographical distribution of the icebergs depend on their calving rate and the winds. Predominantly northerly and northeasterly winds favour the southern extension of the icebergs. These data have however shown their limitation because they are temporally and spatially sparse.

In order to complete information on icebergs characteristics, Bigg et al. (1997) modeled iceberg trajectories over the whole Arctic using a coarse resolution ocean circulation model and sea ice observations from Bourke and Garrett (1987). Here, we propose a more detailed extension of their work applied to the Barents and Kara Seas using a high resolution ice-ocean-iceberg drift model parameterized for the area (Keghouche et al., 2009). Our main goal is to evaluate the climate-related variability of icebergs in the region. We therefore carried out stochastic simulations of iceberg drift trajectories from July 1985 to December 2005. This allows us to quantify over a period of 20 years, the contribution of climatic factors to the average spatial distribution, the size, the source origin, seasonal and interannual variability of their extension.

The outline of this paper is as follows. Section 2 provides a brief description of the model and the experiment set-up. In Section 3.1, we focus on the main results by providing an extensive description of the statistics over the period of study. In Section 3.2, we describe the modeled seasonal extension of icebergs. In Section 3.3, we highlight mechanisms driving the interannual variability of iceberg drift and in particular the extreme southernmost extension. A summary of the study is given in Section 4.

2 Method

2.1 Model

The main characteristics of the iceberg drift model are detailed in Keghouche et al. (2009) and summarized in Section 2.1.1. For the long-term simulations, we added some model features that we describe in Sections 2.1.2 and 2.1.3, concerning the choices for the bottom friction and the melting parameterizations respectively.
2.1.1 Main dynamics

The basic equation describing the horizontal motion of an iceberg of mass $M$ is:

$$M \frac{du}{dt} = F_{AT} + F_W + F_C + F_{SS} + F_{SI}.$$  

(1)

where $u$ is the iceberg velocity. The atmospheric force ($F_{AT}$) and ocean force ($F_W$) act on the cross-sectional area above (resp. below) the water line in a vertical plane (form drag) and a horizontal plane (surface drag) as defined in Smith and Banke (1983). Optimal values for the atmospheric and ocean form drag are 0.70 and 0.25 respectively, based on the best fit for four icebergs observed during two months of drift, (Keghouche et al., 2009). In the latter study, the effect of wind waves is included implicitly through the optimization of the atmospheric form drag as in Smith (1993). $F_C$ is the Coriolis force, $F_{SS}$ is the force due to the sea-surface slope and $F_{SI}$ is the force due to interaction with the sea-ice cover. Note that $F_{SI}$ depends nonlinearly on the sea-ice concentration $f$ and the sea-ice strength $P$ (Lichey and Hellmer, 2001):

$$F_{SI} = \begin{cases} 0 & \text{if } f \leq 15\%, \\ -(F_{AT} + F_W + F_C + F_{SS}) + \frac{dv_{si}}{dt} & \text{if } f \geq 90\% \text{ and } P \geq P_s, \\ \frac{1}{2}(\rho_{si}c_{si}A_{si})|v_{si} - u|(v_{si} - u) & \text{otherwise.} \end{cases}$$  

(2)

$P_s$ is set to 13000 N m$^{-2}$, $v_{si}$ is the sea-ice velocity, $c_{si}$ is the sea-ice coefficient of resistance set to one as in Lichey and Hellmer (2001) and $A_{si}$ is the product of the sea-ice thickness and the iceberg width. If the sea-ice concentration $f$ is lower than 15%, no sea-ice force is applied. If $f$ is between 15% and 90% and the sea ice is strong enough, sea ice and the iceberg form a joint block and the iceberg drifts with the sea ice.

The model inputs are averaged daily to limit memory storage. The atmospheric parameters are from ERA-40 reanalysis of the European Center for Medium-range Weather Forecasting (Uppala et al., 2005, ECMWF), with 1.125° resolution. The ocean and sea ice variables are supplied by a nested configuration of the HYbrid Coordinate Ocean Model (Bleck, 2002, HYCOM). The inner model covers the Barents and the Kara Seas with a horizontal resolution of approximately 5 km and uses 22 vertical hybrid layers. It includes the tides but they are largely filtered out by daily averaging. On seasonal and interannual time scales, we expect the tides to have no influence. The outer model is a version of the TOPAZ3 forecasting system that covers the Atlantic and the Arctic Ocean, run without data assimilation (Bertino and Lisæter, 2008). The dynamic part of the sea-ice component is based on the elastic-viscous-plastic rheology by Hunke and Dukowicz (1997). The thermodynamic fluxes over open water, ice-covered water, and snow-covered ice are given in Drange and Simonsen (1996).

2.1.2 Boundary conditions and stability criterion

When an iceberg hits the bottom, it remains stationary until it has either melted sufficiently to drift off or it is transported toward a deeper region by forces stronger than the frictional force. Based on experimental studies defining friction coefficients of a large ice block on a sand or gravel beach from Barker and Timco (2003), we used a static friction coefficient of 0.5 for grounded icebergs. When an iceberg is in a transition mode, moving from a deep region to a region shallower than its immersed part, we used a friction coefficient of 0.35.

The icebergs are allowed to roll over, following the Weeks and Mellor (1978) stability criterion. Note that an iceberg is removed if its height is less than 5 meters or if it reaches the nesting zone of the
ocean and sea-ice model (thick gray lines in Figure 5). The nesting zone represents the last 20 grid cells of the ocean model. We neglect the icebergs that may re-enter the domain.

2.1.3 Melting parameterizations

Among the mechanisms involved in the deterioration of icebergs, we consider only the most important ones: wave erosion, which is the primary source of melting (White et al. (1980) and Bigg et al. (1997)), lateral melting, and basal melting.

Wave erosion \( V_{\text{wave}} \) parameterization is taken from Gladstone and Bigg (2001), who incorporated a dependency on the sea-surface temperature (SST) and the sea-ice concentration (m/day):

\[
V_{\text{wave}} = \left[ \frac{1}{6} (T_w(1) + 2) \right] S_s \left[ \frac{1}{2} (1 + \cos(f^3 \pi)) \right],
\]

where \( T_w(1) \) is the SST, \( f \) is the sea-ice concentration, and \( S_s \) is the sea state derived from the wind speed. Thus, the wave erosion is damped in presence of sea ice.

Lateral melting \( V_{\text{lateral}} \) is based on the parameterization of Kubat et al. (2007) over the iceberg draft. The empirical estimate of lateral melt rate (m/day) is:

\[
V_{\text{lateral}} = \sum_{k=1}^{n} \left[ 7.62 \times 10^{-3} (\Delta T(k)) + 1.29 \times 10^{-3} (\Delta T(k))^2 \right],
\]

where \( \Delta T(k) \) is the difference between the sea-water temperature and the freezing-point temperature at the \( k^{th} \) layer interface. The iceberg draft crosses the layers 1 to \( n \) of the ocean model.

The estimation of the basal turbulent melting rate \( V_{\text{basal}} \) (in m/day) follows Weeks and Campbell (1973):

\[
V_{\text{basal}} = 0.58 |v_w(n) - u|^{0.8} \times \frac{T_w(n) - T(n)}{L^{0.2}}
\]

where \( v_w(n) \) is the water velocity at the iceberg base and \( u \), the iceberg velocity. \( L \) is the iceberg length, and \( T(n) \) is the iceberg basal temperature. Similar to what it is done for sea-ice model basal-melting parameterizations, \( T(n) \) is the local freezing-point temperature at the iceberg base and \( T_w(n) \) is the local water temperature at the iceberg base.

The accuracy of the melting is strongly dependent on the initial estimate of the iceberg size and its location with respect to the sea-ice edge and the ocean front. The ocean model temperature and the sea-ice concentration compare reasonably well with observations (Bertino et al., 2007). In the model, the winter ice edge is well defined over the domain but it is overestimated west of Svalbard. During the summer, the model may underestimate the presence of sea ice in the northern region of the Barents Sea while overestimating it in the southern region of the Kara Sea. The ocean temperature error is less than 1°C southwest of the Barents Sea. In Keghouche et al. [manuscript in preparation, 2010], a data assimilation method is used to keep the modeled iceberg position close to the observed trajectory of four icebergs drifting in the western part of the Barents Sea from April to July 1990. The melting parameterization provided estimates of the iceberg lifespan which were comparable to the observations.

2.2 Experimental design

The iceberg discharge is controlled by the glacier speed and the geometric changes in the terminus region (Van der Veen, 2002). These processes occur on smaller scales than we consider in this study. The sources of icebergs in the Barents and Kara Seas are marine terminating glaciers and ice cap drainages from Svalbard, Franz Josef Land and Novaya Zemlya. A small amount of icebergs come from the Severnaya Zemlya archipelago, and Ushakov and Victoria islands. Estimates of iceberg production are given in Dowdeswell et al. (2002), Hagen et al. (2003), Kubyshkin et al. (2006), and Dowdeswell et al.
We define 11 calving sites representing the iceberg production of several outlet glaciers (Figure 1) and adjust the iceberg production rate according to their importance (Table 2). The calving sites are located at a single point near the most active regions, considering the bathymetry and distance to the coastline (≃ 20 km). (Personal communication in 2009 with Julian A. Dowdeswell and Andrey Glazovsky regarding the Svalbard and Russian Arctic regions respectively, confirmed that this repartition is plausible).

2.2.1 Size distribution

During the IDAP program (Spring, 1994), extensive observations of icebergs in the Barents Sea from 1988 to 1992 from Aerial Stereo Photography were made. We generate iceberg size characteristics randomly, through a log-normal distribution of the length, width and the freeboard height with mean and variance values, based on the distribution of the observations (Table 1). The width distribution is shown in Figure 2. The freeboard height and the length have similar distribution. Any idealized rectangular tabular iceberg has a depth of four times its freeboard height (Smith and Banke, 1983). Note that these estimates represent the size of icebergs drifting within the Barents Sea and not at their calving site, where they must be larger.
Table 1: Iceberg size characteristics in meters based on IDAP campaign (Spring, 1994).

<table>
<thead>
<tr>
<th></th>
<th>Mean</th>
<th>Standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length</td>
<td>91</td>
<td>53</td>
</tr>
<tr>
<td>Width</td>
<td>64</td>
<td>37</td>
</tr>
<tr>
<td>Freeboard height</td>
<td>15</td>
<td>7</td>
</tr>
</tbody>
</table>

![Image of Table 1](image)

Figure 2: Distribution of iceberg width generated from July 1985 to December 2005.

2.2.2 Release into the sea

In this study, we chose to release icebergs at a constant calving rate despite indications of increased release from June through September (Kubyshkin et al., 2006). This allows us to evaluate the seasonal influence of the ocean currents, wind and sea ice on the iceberg characteristics and drift, independently of the seasonal fluctuations in calving rate.

Knowing the annual calving rate and the mean size of icebergs, it is possible to estimate an averaged number of icebergs released each day from each source. We consider the release of icebergs as an event occurring at random with a known average rate and independently of the time since the last release. This is expressed by a Poisson process. For each source, the probability to release \( k \) icebergs at a given day, knowing the yearly averaged number of release \( \lambda \) is:

\[
f(k, \lambda) = \frac{\lambda^k e^{-\lambda}}{k!},
\]

(6)

The random length of a calved iceberg is picked up from the log normal distribution based on the mean and variance length given in the IDAP database (see Section 2.2.1). The width and freeboard height are assigned using the same procedure.

Based on the estimates given in Tables 1 and 2, the annual number of iceberg each year would be \( \approx 19000-20000 \). It is unclear which proportion of these remains grounded near the calving site and which proportion drifts freely in the Barents Sea. To limit the amount of computations, we divided the calving rates from all sites by a factor of 100 and hence launched about 200 icebergs per year. The release occurs from July 1985 to December 2005. Note that no icebergs were released from Victoria island through the period of study, so this minor source will not be discussed in the following.

3 Results

The simulation starts in July 1985 and ends in December 2005. The average age of icebergs is less than a year and relatively few icebergs persist for more than two years (see Section 3.1.2). Therefore, we consider the first year and a half as the "spin-up time" for icebergs to spread all over the domain and
thus we retain only the last 19 years of simulation. This section presents the results with statistics on different time scales: a 19-year mean for the climatology, monthly means for the seasonal variability and yearly means for the interannual variability.

### 3.1 Climatology for the period 1987-2005

#### 3.1.1 Grounded icebergs

Most of the grounded icebergs are located on shallow waters near their point of calving origin (Figure 3), especially around Nordaustlandet, the two Franz Josef Land sources and the northwest tip of Novaya Zemlya. A small proportion of the grounded icebergs are located along Svalbard bank and the bank starting from the Kara Sea shelf to the edge of Saint Anna Trough. Approximately 77% of the released icebergs become grounded. On average, the icebergs that become grounded spend about 42% of their lifetime motionless. Among the icebergs that are grounded at least once, on average 14% remain grounded for more than 80% of their lifetime. Except for the West Spitzbergen and East Severnaya Zemlya sources, all the others have a large probability (over 80%) of their icebergs becoming grounded at least once. During the IDAP campaign (Spring, 1994), the regions north of Hopen and Svalbard bank were identified as regions where icebergs grounded and melted in place. This is observed in the model as well: the proportion of grounded icebergs that spend most of their life grounded is 22% and 32% for iceberg comings from Nordaustlandet and Edgeøya, usually travelling southwest.

For some of the calving sites, a significant proportion drift out of the model domain, at a "matured" age, see Table 3. The exceptions are icebergs coming from West Spitzbergen where 45% of the icebergs are lost after 50 days of drift in average.

The friction parameterization had a limited effect on the grounding time. One year of simulation with and without the friction parameterization exhibits very little differences in the spatial density map of grounding sites and the average grounding time remained relatively similar (not shown).
3.1.2 Averaged age

The spatial distribution of the icebergs mean age is heterogeneous (Figure 4). The youngest icebergs are found close to the calving sites, whereas the oldest are often located along the southernmost positions or the northeastern model boundaries, especially southwest of the Barents Sea and along the bank starting from the Kara Sea shelf to the Saint Anna Trough, where icebergs might get grounded (Figure 3). Icebergs are on average 241 days old. About 20% of the icebergs survive more than a year and only 3.3% survive more than 2 years. The same statistics applied to each calving sources highlights large discrepancies from one source to another (Table 3). The mean age is maximum for icebergs coming from West Severnaya Zemlya (370 days) and East and West Franz Josef Land archipelago (329 and 314 days respectively). The oldest iceberg (5.9 years) comes from East Franz Josef Land. The lowest average lifetime of icebergs is 55 days and corresponds to icebergs coming from West Spitzbergen, which leave the model domain relatively quickly, see Section 3.1.1.

3.1.3 Annual probability of occurrence

A map of the probability of encountering an iceberg within the year shows that most of the icebergs are located close to calving sources and the probability of encountering an iceberg gradually decreases with the distance to the calving site (Figure 5). Abramov and Tunik (1996) processed a similar map for the period 1881 to 1993 based on observations from ships, aerial surveys and ARGOS buoys; 35% of the observations were gathered from 1987 to 1993. The direct qualitative comparison of both maps is challenging, due to the short time overlap and the sparse sampling of the observations. Qualitatively, we observe similar patterns in the iceberg distribution.
Table 3: For each iceberg source, estimates of the mean iceberg age, the percentage of icebergs leaving the model domain, their mean age at that time and the total number of icebergs based on simulations from 1987 to 2005. See Figure 1 for the location of the sources.

<table>
<thead>
<tr>
<th>Source number and name</th>
<th>Mean age in days</th>
<th>Percentage leaving the model boundary</th>
<th>Mean age in days when leaving the model boundary</th>
<th>Maximum age in days</th>
<th>Number and proportion</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-Nordaustlandet</td>
<td>260</td>
<td>16.2</td>
<td>250</td>
<td>1966</td>
<td>800 (22.2%)</td>
</tr>
<tr>
<td>2-West Spitzbergen</td>
<td>55</td>
<td>46.7</td>
<td>53</td>
<td>294</td>
<td>666 (18.5%)</td>
</tr>
<tr>
<td>3-Edgeøya</td>
<td>156</td>
<td>4.1</td>
<td>106</td>
<td>526</td>
<td>176 (4.9%)</td>
</tr>
<tr>
<td>4-West Franz Josef Land</td>
<td>314</td>
<td>34.2</td>
<td>214</td>
<td>1288</td>
<td>529 (14.7%)</td>
</tr>
<tr>
<td>5-East Franz Josef Land</td>
<td>329</td>
<td>14.8</td>
<td>325</td>
<td>2146</td>
<td>775 (21.6%)</td>
</tr>
<tr>
<td>6-East Novaya Zemlya</td>
<td>249</td>
<td>2.4</td>
<td>521</td>
<td>911</td>
<td>168 (4.7%)</td>
</tr>
<tr>
<td>7-West Novaya Zemlya</td>
<td>268</td>
<td>5.4</td>
<td>528</td>
<td>1284</td>
<td>276 (7.7%)</td>
</tr>
<tr>
<td>8-East Severnaya Zemlya</td>
<td>122</td>
<td>100.0</td>
<td>115</td>
<td>1476</td>
<td>149 (4.1%)</td>
</tr>
<tr>
<td>9-West Severnaya Zemlya</td>
<td>370</td>
<td>71.7</td>
<td>354</td>
<td>1407</td>
<td>53 (1.5%)</td>
</tr>
<tr>
<td>10-Ushakov Island</td>
<td>283</td>
<td>33.3</td>
<td>338</td>
<td>338</td>
<td>3 (0.1%)</td>
</tr>
</tbody>
</table>

3.1.4 Pathways

The spreading of icebergs inside the domain is complex and chaotic, see Figure 8. The total number of icebergs released from each source is reported in Table 3. The implementation allows us to follow the trajectory of icebergs from their calving site. A detailed description of the most important pathways from each source is presented in Figure 6. The modeled drift patterns compare well with the ones described in Spring (1994, p. 66), based on Argos buoys placed on 10 icebergs from 1988 to 1992 (not shown).

The icebergs that have the largest spread within the domain are coming from East Franz Josef Land and are found in the northern Kara Sea and the northern Barents Sea (Figure 8).

Icebergs coming from the eastern side of Svalbard region stay in the western part of the Barents Sea. They usually move southwest, along the coast, following the East Spitsbergen current over Svalbard bank or even closer to the coast of Svalbard.

Note that icebergs from West Spitsbergen have a relatively small spreading into the domain despite their important number. They have the tendency to quickly drift off the model boundary, see Section 3.1.1.

The icebergs drifting in the central Kara Sea originate mainly from East Novaya Zemlya.

3.2 Seasonal variability

The difference between the mean iceberg density during the summer months and the mean iceberg density during the winter months from 1987 to 2005 is an indicator of the seasonal variability, see Figure 7. Here, the summer is defined from 15 May to 15 November and the winter consists of the other months of the year.

In general, the number of icebergs close to the calving sources is higher in winter than in summer. The extreme southernmost positions occur during the summer period, in accordance with Abramov (1992). The presence of strong and highly-concentrated sea ice in winter slows down the motion of iceberg and limit their extension. Maximum iceberg extension occurs in June-July and the minimum iceberg...
extension occurs in October-November most of the years.
In Figure 7 (left), we also find a seasonal imbalance in the iceberg density on each sides of the islands producing icebergs (e.g. Franz Josef Land, Novaya Zemlya and Svalbard). There are more icebergs located on the east (west) of the island in winter (summer). This might be due to the preponderance of northeasterlies in winter, which transport the icebergs away if the source is on the west and maintain them close to their source if the latter is on the east of the island.
We observe a seasonal variability of the iceberg extension, though we did not consider the seasonality of the calving. It suggests that seasonal variability of forcing fields, i.e., the wind fields, ocean currents and sea ice are important for the seasonal spreading of icebergs. Finally, we looked at the distribution of iceberg released in summer and compared it with the distribution of iceberg released in winter (Figure 7, right). The icebergs released in summer have the largest spreading. Therefore, we suspect that introducing seasonal variability in the calving would intensify the seasonal variability of the iceberg spreading.

3.3 Interannual variability of the iceberg extension over the domain

In this section we focus on the mechanisms driving the spatial extension of the icebergs, by retaining them not far from the calving sites some years or transporting them very far south some others. In addition we analyze the extreme southernmost extension in our model.
3.3.1 Iceberg extension

The interannual variability of iceberg spreading is large. The year with the smallest spreading in the domain and the year with the largest one are shown in Figure 8. In order to evaluate the interannual variability of the iceberg spreading, we define a measure for the iceberg extension. The iceberg extension is the sum of any grid cell area where an iceberg has drifted independently of the number of passages within that grid cell. The annual time series have a strong variability and highlight a weak decreasing trend (Figure 9) over the period of study. Years with the largest spreading are 1988, 1987, 2003 and 1992 (sorted from the largest to the smallest), whereas years with the smallest extensions are 2000, 1997, 2001 and 1995 (from the smallest to the largest). If we remove the trend from the time series (not shown), the extrema appear in the second half of 1987-2005 period with a minimum kept in 2000 and a maximum in 2003. The variability of the iceberg extent has a time scale longer than two years (autocorrelation of 0.5, with one year lag), indicating a potential for prediction of this parameter.

In order to gain some insight on the mechanisms involved in the interannual variability of the iceberg spread in the domain, we study the relationship between the iceberg extent and the main forcing components of our model, i.e. the atmosphere, ocean currents and sea ice. Results from correlation analysis are summed up in Table 4. It reveals two mechanisms acting on the iceberg extent: one dynamic and the other thermodynamic.

Dynamics

We found large differences in the correlation between different sea ice quantities and the iceberg extent. The annual sea-ice area within the domain has a correlation of 0.64 with the iceberg extent, whereas the
Figure 6: Pathways of icebergs from their calving site based on the model runs covering the period 1985-2005. Main pathways are shown with larger arrows.

The annual sea-ice volume has no significant correlation. It implies that the iceberg extent is independent of the ice thickness within the domain. Note that our iceberg drift model is constraint non-linearly by the surrounding sea-ice concentration and thickness, as described in Equation 2. It is important to understand whether the high correlation between the sea-ice area and iceberg extent is due to a grasping effect of sea ice on the icebergs or rather a common reaction to an external force. The mean annual percentage that the icebergs drift with the sea ice varies between 15% and 30%. The time series are correlated at $\approx 0.37$ with the iceberg extent and $\approx 0.63$ with iceberg extent from the preceding year. Therefore, the intermittent movement of icebergs driven by sea ice is not the main mechanism controlling the iceberg extent. Concerning the sea ice volume transport into the domain, it is correlated at $\approx 0.41$ with the iceberg extent and at $\approx 0.46$ with the sea ice area. The correlation between the sea-ice volume-transport into the domain, the sea-ice area and the iceberg extent suggests a relationship with the atmospheric forcing, which is analyzed in more detail below.

Regression analysis of the winter mean sea level pressure (MSLP) anomalies over the North Atlantic and the Arctic with the winter sea ice extent over the Barents Sea during the period 1967-2002 (Sorteberg and Kvingedal, 2006), reveals that higher than normal MSLP over Svalbard region favors large sea ice extent. It leads to a weakening of the westerlies and more locally in the Barents region, to anomalous southward geostrophic winds over the period 1967 to 2002. We found similar results (not shown) by applying the regression analysis on the iceberg extent during our period of study: 1987-2005. We therefore suggest that the northerly winds are the principal actors enhancing iceberg extent.

**Thermodynamics**

The SST over the domain is anti-correlated with the iceberg extent with a value of -0.49. Interannual variability of the mean SST anomaly is anti-correlated at $\approx -0.66$ to the annual mean age anomaly of icebergs during the year of their calving. It indicates the importance of the thermodynamics in the age and therefore the extension of icebergs.
The warm Atlantic Water (AW) enters the Barents Sea between Norway and Svalbard. When passing through the Barents Sea, the AW is strongly modified by cooling, mixing and freezing during winter before entering the Arctic Ocean (Ingvaldsen et al., 2004a). The modeled positive transport into the Barents Sea has a negative correlation of -0.28 with the iceberg extent over the 19 years of simulation. It increases to -0.66 with a lag of one year. This one-year lag is similar the time needed for the heat anomalies to spread within the Barents Sea. The variability of the inflow to the Barents Sea is highly dependent on the winds (Ingvaldsen et al., 2004b). We processed a linear regression and correlation analysis of the MSLP anomalies with the iceberg extent lagged by one year (Figure 10). This indicates that the iceberg extent is correlated at ≃ 0.6 with MSLP anomalies over the Svalbard region. Positive MSLP anomalies over Svalbard-Nordic Seas region may limit the intensity of the Atlantic inflow to the Barents Sea due to a reduced cyclonic activity over the Nordic Seas or enhanced northerly winds. It limits the heat content of the Barents Sea the following year, increases the mean age of icebergs and thus their potential extension.

Note that we did not find any connection between the iceberg extent and large-scale atmospheric patterns such as the North Atlantic Oscillation or Arctic Oscillation with either a zero or a one-year time lag.

### 3.3.2 Extreme southernmost extension

In order to analyze the geographical locations of the southernmost icebergs, we first analyse the latitudinal distribution of the icebergs occurrence within different latitudinal ranges at yearly time scale (Figure 11). The iceberg position is stored once per day, thus an occurrence is a daily event. Any occurrence south of 75°N is considered to be "extreme". Considering the number of occurrences rather than the number of individuals allows us, to weight the icebergs with respect to the time spent south. We observe that the years 2003, 1993, 2002 and 1999 have the largest number of occurrences south of 75°N over the domain. To complete the information about the the spatial spread, we introduce another quantity: the meridional percentage of iceberg occurrence found south of 75°N (see Figure 12). The latter quantity exhibits discrepancies on a regional scale. Extreme southernmost extensions occurred at different times in the Barents and Kara Seas. For example, the 2003 and 1993 southernmost extension is large for both
regions, while the large spread during the years 2002 and 1999 was only in the Kara Sea. Most of the icebergs passing 75°N are located in the western part of the Kara Sea with a peak between 60°E and 65°E, except in 2002 when the peak shifted to 65°E-70°E. Southernmost Kara Sea icebergs come primarily from East Novaya Zemlya glaciers, and from West Novaya Zemlya and East Franz Josef Land for some years.

In the Barents Sea, the icebergs passing 75°N are concentrated in the eastern part, except in 1993 when a large amount of icebergs passed between 20°E and 30°E as well. The year with the largest southernmost extension is 2003, with icebergs mostly concentrated between 35°E and 55°E. Southeast Barents Sea icebergs come mainly from Franz Josef Land glaciers and West Novaya Zemlya. The same year (2003), icebergs from West Severnaya Zemlya were also found in the south of the Barents Sea (Figure 13). Southwest Barents Sea icebergs come mainly from West Franz Josef Land glaciers and possibly Edgeøya or Nordaustlandet. In relation to the dynamic and thermal mechanisms suggested previously, a positive MSLP anomaly was present over the Svalbard region for all of the years during which the iceberg extent in the Barents Sea was at its southernmost extreme (2003, 1993, 1988 and 1996). During these periods the transport of AW into the Barents Sea was also lower than the average value with the exception of the year 1996. In 1996 fewer icebergs drifted south of the Barents Sea compared to the other years, yet their lifespan allowed for them to drift past 75°N impacting the number of occurrences. The year 1996 is “extreme” in the sense that the anomalous southeasterly winds were observed.

Note that the northernmost extension occurred in 1989 for the Kara Sea, while for the Barents Sea, 1999, 2000, 2005 are years with the northernmost extension.

**Note for the southernmost extension in 2003**

The year 2003 is extreme in terms of the proportion of icebergs that drift southeast of the Barents Sea. A field campaign in May of that year observed a surprisingly large number of icebergs in that region (Zubakin et al., 2004). In our simulation, a group of icebergs is observed at the same location at the same time. This group came from East Franz Josef Land and most of the icebergs were calved during the same period, from late September 2002 until late October 2002. Despite of their large spread, they followed similar drift patterns. A persistent southward drift was followed by a northward (wind), transporting them northward. Each northward drift lasting approximately 20 days.
Figure 9: Annual mean of the iceberg extent (black) and ice area (dotted gray line) over the model domain.

4 Conclusion

This study is an attempt to describe the iceberg distribution in the Barents and Kara Seas from a modeling perspective. Lack of iceberg drift observations in the eastern part of the domain makes a thorough validation complicated. Using iceberg drift observations in the western region of the Barents Sea in 1990, Keghouche et al. (2009) obtained modeled trajectories, which were reasonably consistent and Keghouche et al. (manuscript in preparation, 2010) show that the melting parameterization is in accordance with the observed iceberg lifespan when icebergs are replaced close to the observed positions and when the initial mass is accurately estimated. The average age and extent of icebergs are somewhat uncertain but the qualitative mechanisms explained in this study should remain valid.

Simulations of iceberg trajectories were performed from July 1985 to December 2005. The first 1.5 years of the simulation are not used to allow sufficient time for the model to spin up. Maps of iceberg density, potential locations subject to grounding complement existing statistics of icebergs characteristics in the Barents Sea derived from sparse oceanographic and aerial field campaigns (Abramov and Tunik, 1996; Spring, 1994). The model suggests that the icebergs follow preferential pathways (Figure 6) from their respective calving sources. We found that icebergs originating from East Franz Josef Land have the largest spread over the domain.

Simulations confirm the seasonal cycle of the southernmost extension observed by Abramov (1992). We observe a large seasonal variability of the iceberg extension, though we did not consider the seasonality of the calving. It suggests that seasonal variability of forcing fields, i.e., the wind fields, ocean currents and sea ice are very important for the seasonal spreading of ice berg.

We observed large interannual variability of the iceberg extent with a weak decreasing trend, in accordance with the observed sea-ice extent. The latter two quantities are also strongly correlated when they are detrended (0.64). Sorteberg and Kvingedal (2006) found that MSLP higher than normal over Svalbard region enhance sea ice extent in the Barents Sea through enhanced northerly winds. Following their approach, regression and linear correlation analysis of the iceberg extent over our period of study leads to a similar conclusion for the iceberg extent.

We also suggest that the positive transport into the Barents Sea is a potential predictor of iceberg extent,
Table 4: Correlations between the annual iceberg extension and the annual sea ice area (SI area), the sea ice volume (SI vol), sea ice transport into the domain, from Svalbard to Severnaya Zemlya (SI trp), the positive transport along Norway-Svalbard section (N-S trp +), the sea surface temperature (SST), for the period 1987-2005.

<table>
<thead>
<tr>
<th>SI area</th>
<th>SI vol</th>
<th>SI trp</th>
<th>N-S trp+</th>
<th>SST</th>
</tr>
</thead>
<tbody>
<tr>
<td>With trend</td>
<td>0.68*</td>
<td>0.13*</td>
<td>0.44**</td>
<td>-0.30n.s</td>
</tr>
<tr>
<td>Without trend</td>
<td>0.64*</td>
<td>0.03*</td>
<td>0.41**</td>
<td>-0.28n.s</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>One year time lag</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>With trend</td>
<td>-0.65*</td>
</tr>
<tr>
<td>Without trend</td>
<td>-0.66*</td>
</tr>
</tbody>
</table>

* p ≤ 0.1, with a t-test that takes account the autocorrelation of the iceberg extent, p ≤ 0.05 with a t-test neglecting the autocorrelation of the iceberg extent.

** Same as * with p ≤ 0.2 and p ≤ 0.1 respectively.

n.s The correlation is not significant.

as they have a one-year lag correlation of -0.66. The AW affects the heat balance in the Barents Sea and therefore the iceberg lifetime. The inflow into the Barents Sea is highly dependent on the winds (Ingvaldsen et al., 2004b). A linear regression and correlation analysis of the MSLP anomalies with the iceberg extent lagged by one year shows that MSLP higher than normal east of the Greenland Sea and south of Svalbard region enhanced iceberg extent the following year. We suggest that this is due to the fact that a reduced cyclonic activity over the Nordic Seas limits the intensity of the warm inflow into the Barents Sea and reduces the heat content of the Barents Sea the next year, increasing the mean age of iceberg and thus their potential extension.

Finally, the model was able to simulate the observed extreme extension of icebergs originating from Franz Josef Land in May 2003.

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References


Barker, A., Timco, G., 2003. The friction coefficient of a large ice block on a sand/gravel beach. In:


Figure 10: Linear correlation (left) and regression (right) between the annual MSLP anomalies and the annual mean iceberg extent lagged by one year for the period 1987-2005.

Figure 11: Annual occurrence of icebergs at various latitudes south of 77°N for the period 1987-2005. The colored lines correspond to the years with the largest number of iceberg occurrence south of 75°N within the model domain.
Figure 12: Annual percentage of occurrence found south of 75°N depending on their longitude location for the period 1987-2005. The number was split by classes of 5° of longitude. The colored lines correspond to the years with the largest number of occurrence south of 75°N within the Barents Sea area.

Figure 13: Annual percentage of icebergs from each source found south of 75°N for the period 1987-2005. The colored lines correspond to the years with the largest number of occurrence south of 75°N within the Barents Sea area.
Paper III

Adaptive estimation of iceberg parameters using the Ensemble Kalman Filter

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to be submitted
Adaptive estimation of iceberg parameters using the Ensemble Kalman Filter

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To be submitted

Abstract

Icebergs in the Barents Sea are a potential threat for navigation and offshore installations. Hence, there is an increasing demand for accurate monitoring of icebergs in that region. An iceberg drift model has been implemented for the Barents Sea using a nested configuration of the Hybrid Coordinate Ocean Model (HYCOM) and ERA-interim as forcing fields. Previous modeling studies in that region showed large sensitivity of the results to the scaling factors that control the form of the iceberg and the amplitude the forcing fields. Here, the Ensemble Kalman Filter (EnKF) data assimilation method is used to correct the position of the iceberg and estimate these parameters. Observed trajectories of four icebergs from 1990 were assimilated once per day in the model. The two northernmost simulated trajectories have a precision of 15 km and the two southernmost ones have a precision better than 25 km over two months of drift. In addition, the estimation of non-observed parameters by the EnKF allows to identify the error in the estimated iceberg form as well as in the forcing fields.

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1 Introduction

Navigation is expected to increase with the reduction of sea ice in the Arctic Ocean and more specifically in the Barents Sea. Hence, for safety of marine operations, there is an increasing demand for accurate monitoring of icebergs, considered as a potential threat. An operational iceberg drift model (Kubat et al., 2007) is already used by the Canadian Ice Service and the International Ice Patrol to track individual icebergs over Great Banks region. There is a need for similar system in the Barents Sea. In that region, ocean dynamics occurs on small scales so observations are not sufficient. Hence, forcing of the iceberg model needs to be provided by a numerical model. Keghouche et al. (2009) modeled successfully icebergs trajectories in the Barents Sea. Their iceberg drift model is forced by a high resolution, coupled sea ice-ocean, model. The atmospheric forcing fields are taken from the European Center for Medium range Weather Forecasting (ECMWF). Iceberg drift models become inaccurate due to errors in the forcing fields, the uncertainty on the iceberg form and the melting parameterization. A common approach is to estimate constant parameters that compensate for these errors. In Keghouche et al. (2009) they are estimated over the whole iceberg trajectory for four observed iceberg drift, using a Monte Carlo approach with regular sampling of the parameters. They found that the accuracy of the results was very sensitive to these parameters. However, it is expected that these parameters should evolve with time with the change in the iceberg form and the errors in the forcing. This study is an evaluation of a data assimilation method for monitoring iceberg drift and time-dependent parameter estimations. Because of the non-linear properties of the iceberg model, we use the Ensemble Kalman Filter (Evensen, 2003, EnKF) that relies on the statistics of an ensemble not only to correct the position of the iceberg, but also the non-observed variables. The motivation is to correct the model parameters, knowing the model error, so that the next prediction would be more accurate. For example, if data assimilation is able to identify from the trajectory that the mass of the iceberg is overestimated, such information could be greatly beneficial for improving the following forecast. Two model parameters will be estimated along the iceberg trajectory, to account for errors in the form, errors in the forcing, and mechanisms that are not considered in the model. The model is an improved version of the one used in Keghouche et al. (2009) that includes melting parameterization and allows the iceberg to roll over. As a preliminary test, we focus on the analysis of four iceberg drifts in the northwest part of the Barents Sea in 1990. They were measured during IDAP campaign (Spring, 1994) using ARGOS buoys. Section 2 presents the main characteristic of the iceberg model, and the forcing. A brief description of the dataset is given in section 3, while section 4 presents the principles of the Ensemble Kalman Filter and its implementation for our experiment. The results are presented and discussed in section 5, and the conclusions are given in section 6.

2 Model

The basic equation describing the horizontal motion of an iceberg of mass M is:

\[
M \frac{du}{dt} = F_{AT} + F_W + F_C + F_{SS} + F_{SI},
\]  

(1)

where \( u \) is the iceberg velocity. The atmospheric force (\( F_{AT} \)) and ocean force (\( F_W \)) act on the cross-sectional area above and below the water line, respectively in a vertical plane (form drag) and a horizontal plane (surface drag) as defined in Smith and Banke (1983). \( F_C \) is the Coriolis force, \( F_{SS} \) is the force due to the sea surface slope and \( F_{SI} \) is the force due to interaction with the sea ice cover. The atmospheric
force is
\[ \mathbf{F}_{\text{AT}} = \left[ \frac{1}{2} (\rho_a c_a A_{va}) + (\rho_a c_{da} A_{ha}) \right] |v_a - u|(v_a - u). \quad (2) \]
The oceanic force \( \mathbf{F}_{\text{W}} \) is defined by the same quadratic drag law as \( \mathbf{F}_{\text{AT}} \) for each ocean model layer through the depth of the iceberg,
\[ \mathbf{F}_{\text{W}} = \sum_{k=1}^{n} \left[ \frac{1}{2} (\rho_w c_w A_{vw}(k)) |v_w(k) - u|(v_w(k) - u) \right. \\
left. + (\rho_w c_{dw} A_{hw}(n)) |v_w(n) - u|(v_w(n) - u) \right]. \quad (3) \]
In Equations 2 and 3, \( v_a \) and \( v_w(k) \) are the air and water velocities, respectively. The term \( A_{va} \) (resp. \( A_{vw}(k) \)) is the vertical cross section area in the air (resp. water). The wind is assumed to be constant with height above sea level, but the ocean currents vary with depth according to the ocean model. The index \( k \) is the ocean layer number and \( n \) is the number of ocean layers in contact with the iceberg. The air and water densities are \( \rho_a \) and \( \rho_w \) and \( c_{da} \) and \( c_{dw} \) are the skin drag coefficients set to 0.0022 and 0.0055, the same values as used in the sea ice model (Hunke and Dukowicz, 1997). \( A_{ha} \) (resp. \( A_{hw}(n) \)) is the horizontal area of the iceberg in contact with the air (resp. ocean layer \( n \)) and \( c_a \) and \( c_w \) are the corresponding form drag parameters. Their usefulness is emphasized in section 2.2. The mechanisms involved in the deterioration of icebergs included in the model are wave erosion and lateral and basal melting parameterizations. If it is unstable, the iceberg is allowed to roll over. The iceberg is considered melted if the freeboard height or its lateral dimensions are less than one meter. We refer to Keghouche et al. (2009) for more details on model dynamics and Keghouche et al. (2010) for model thermodynamics. The iceberg-drift trajectory is calculated using a predictor-corrector method with a constant time-step of 600 s.

2.1 Forcing

The wind fields from ERA-Interim are given on a 0.5° grid-cell resolution and updated every 6 hours. The ocean and sea ice variables are supplied by a nested configuration of the Hybrid Coordinate Ocean Model (Bleck, 2002, HYCOM). Sea ice dynamics are simulated using the elastic-viscous-plastic formulation of Hunke and Dukowicz (1997). Thermodynamic fluxes over open water, ice covered water, and snow covered ice are described in Drange and Simonsen (1996). The inner model covers the Barents and the Kara Seas with a horizontal resolution of approximately 5 km, 22 vertical hybrid layers and includes tides. The tides are simulated as a barotropic pressure forcing at the open boundary of the inner model. It is calculated with eight constituents: K1, O1, P1, Q1, M2, N2, S2 and K2, taken from the Finite Element Solution global atlas (Lyard et al., 2006, FES2004). The outer model is a version of the TOPAZ3 forecasting system that covers the Atlantic and the Arctic Ocean, and is run without data assimilation (Bertino and Lisæter, 2008). The ocean forcing fields are updated in two steps in order to limit the input/output. The slow varying variables, i.e. the baroclinic velocity, temperature, and salinity are updated every day whereas the fast varying variables, i.e. the barotropic velocities, the ice thickness, ice velocities and ice concentration are updated hourly. All the variables are interpolated linearly at the position of the iceberg.

2.2 Importance of the form drag parameters

The form drag coefficients \( c_a \) and \( c_w \) are commonly introduced in the drag forces (Equations 2 and 3) to account for errors in the estimated area of the vertical plane in contact with the ocean currents and the atmosphere. In addition those parameters include errors in the forcing and the parameterization. If the iceberg form is perfectly known and the forcing fields exact, both form drags are unnecessary and set to
one. In this study, we attempt to analyze their evolution and their behavior for each type of error.

Keghouche et al. (2009) reported that when the motion of iceberg is not fully driven by the sea ice, the forces driving the iceberg are the atmospheric, oceanic, and the Coriolis forces. For the small Barents Sea icebergs, one can assume that the forces acting on the horizontal cross sectional area are negligible compared to the one acting on the vertical face of the iceberg. It leads to the following relation:

\[
M \frac{du}{dt} = c_a A_v F_{AT} + c_w A_{vsw} F_{WA} + M F_C,
\]

where \( F_C \) is the Coriolis force divided by the mass. \( F_{AT} \) and \( F_{WA} \) are the atmospheric and ocean form drag forces that are divided by their respective estimated form drag parameter \( c_a \) and \( c_w \). \( A_{vsw} \) is the total vertical area in contact with the ocean currents. Note that we do not vary \( c_w \) with depth, though internal wave drag might occur in the pycnocline (Smith, 1993).

An iceberg is assumed to have a constant geometry (tabular). This assumption induces inaccuracies in the modeled trajectory even with perfect forcing since in reality the shape of the iceberg varies in time and may be complex. Hence, one can define the following “true” parameters:

\[
A'_{va} = k_a A_v, \quad A'_{vsw} = k_w A_{vsw} \text{ and } M' = k_m M,
\]

where \( M', A'_{va} \) and \( A'_{vsw} \) are the “unknown true” values for the mass and the areas in contact with the atmosphere and the ocean respectively. The scalars \( k_a, k_w \) and \( k_m \) are scaling factors. The form drag coefficients considering only the error in the form at one particular time are:

\[
\tilde{c}_a = \frac{k_a}{k_m} \quad \text{and} \quad \tilde{c}_w = \frac{k_w}{k_m}.
\]

Thus, an overestimation of \( A_va \) (resp. \( A_{vsw} \)) leads to a decrease of \( k_a \) (resp. \( k_w \)) and, if the mass is overestimated, \( k_m \) is smaller than one and both \( \tilde{c}_a \) and \( \tilde{c}_w \) are increased. These form drags are expected to vary with time as the iceberg is exposed to deterioration processes.

A sea ice-ocean model and atmospheric reanalysis are used to force the iceberg model. They contain biases and errors due to their coarse resolution and their parameterization. The model biases are independent of time and location and will thus be similar for every iceberg. They can be compensated for by introducing a scaling factor in the form drag. Note that there is a component of the forcing error that varies in time and space. Adjustment of \( c_a \) and \( c_w \) compensates only partly for this error.

Finally, form drags account for parameterizations not included explicitly in the iceberg model, i.e. the “added mass” and the wave radiation (Keghouche et al., 2009; Smith, 1993). The “added mass” is a term proportional to the iceberg mass that simulates acceleration of water entrained in the wake of the iceberg (Sodhi and El-Tahan, 1980). This minor effect (Smith, 1993) is assumed to be only dependent on the mass and is represented by an artificial adjustment of the form drags. When no wave information is available, the wave radiation force is parametrized based on the Beaufort scale (Bigg et al., 1997), and is proportional to the wind drag force. In our case, it is implicitly represented by adjusting the air drag coefficient. Note that this latter effect is damped in presence of packed sea ice. Hence, \( c_a \) and \( c_w \) can be defined as follow:

\[
c_a = \frac{k_a}{k_m} C_{sta} \epsilon_a \quad \text{and} \quad c_w = \frac{k_w}{k_m} C_{stw} \epsilon_w,
\]

where \( C_{sta} \) and \( C_{stw} \) are constant in time and space. They include the term that encompass forcing biases and “added mass”, while \( \epsilon_a \) and \( \epsilon_w \) include errors in the forcing and the effect of wind waves that vary in time and space.

### 3 Iceberg observations

This study focus on the four icebergs previously studied in Keghouche et al. (2009). These icebergs were observed using ARGOS buoys during IDAP campaign (Spring, 1994). They drifted in spring 1990.
over the northwest part of the Barents Sea. Their initial size characteristics and positions are specified
in Table 1. We do not have any information about their shape and consider them as tabular icebergs.
The density of pure ice suggests that the freeboard height should represent 13% of the total height of a
tabular iceberg but in the following, it is set to 20% to account for irregularities in the iceberg shape, as
suggested by Smith and Banke (1983).

<table>
<thead>
<tr>
<th>Buoy</th>
<th>L</th>
<th>W</th>
<th>H</th>
<th>Initial position</th>
<th>Initial time (GMT)</th>
<th>Final position</th>
<th>Final time (GMT)</th>
</tr>
</thead>
<tbody>
<tr>
<td>7086</td>
<td>90</td>
<td>60</td>
<td>10</td>
<td>31.48°W, 78.14°N</td>
<td>24/04, 11 am</td>
<td>25.13°W, 75.99°N</td>
<td>30/06, 00 am</td>
</tr>
<tr>
<td>7087</td>
<td>63</td>
<td>56</td>
<td>10</td>
<td>31.06°W, 78.12°N</td>
<td>24/04, 11 am</td>
<td>26.62°W, 76.75°N</td>
<td>04/07, 05 am</td>
</tr>
<tr>
<td>7088</td>
<td>95</td>
<td>80</td>
<td>20</td>
<td>34.22°W, 78.92°N</td>
<td>25/04, 09 pm</td>
<td>29.68°W, 76.85°N</td>
<td>18/07, 07 am</td>
</tr>
<tr>
<td>7089</td>
<td>95</td>
<td>90</td>
<td>15</td>
<td>34.86°W, 79.03°N</td>
<td>25/04, 09 pm</td>
<td>32.10°W, 76.66°N</td>
<td>02/07, 03 am</td>
</tr>
</tbody>
</table>

Table 1: Initial length (L), width (W), freeboard height (H) in meters, and the drifting period of the
observed icebergs.

4 Data assimilation method

The Ensemble Kalman Filter (EnKF) is a sequential filter method based on Monte Carlo approach. At
a given time, the method uses the statistics drawn from an ensemble of model simulations, a set of
observations and their uncertainty in order to estimate a new ensemble that is statistically closer to the
“true” value. Our current system is a non-linear Lagrangian type of model, with a tendency towards
bifurcation of close states. The EnKF is a robust method that has demonstrated its ability to handle
non-linear problems using a simple formalism (Evensen, 2009). It is multivariate and therefore well
suited for parameter estimation. It also provides an additional error estimate of the prediction (Evensen,
2009). For doing so, an ensemble of simulations is run forward with perturbed values of the uncertain
parameters. It is then possible to estimate a value favoring an agreement with the observation. The
parameter values are then updated linearly with the misfits to the observation. This estimation is repeated
for each new observation. Observations considered here are the positions of the iceberg (lon, lat), on a
daily frequency. Once per day, the following analysis is solved:

\[
X^a = X^f + (H^f (H^f)^T)^{-1} H^f (Y - H X^f). \tag{8}
\]

\(X^a\) and \(X^f\) denotes the analysis and the forecast ensemble composed of \(p\) model states. The model state
consists of the prognostic variables lon and lat and is enlarged with the model parameters \(c_a\) and \(c_w\). The
prime indicates that the ensemble is centred, i.e. when the ensemble mean is removed. The superscript T
denotes a transpose matrix. The ensemble size is \(m\), \(Y\) corresponds to the perturbed measurement matrix
(lon, lat), and \(R\) is the corresponding measurement error covariance matrix considered to be diagonal.
The error variance is set uniformly to 11 km\(^2\), i.e. to 0.1° for lat and 0.1° × \(\cos(lat)\) for lon. It includes
the measurement errors (negligible here) and the representation error that accounts for the difference of
resolution between the forcing fields and the measurements. \(H\) is the measurement operator relating the
prognostic model state variables (lon, lat) to the measurements. Finally, we used an ensemble inflation of
10% in order to circumvent ensemble spread collapse caused e.g. by the suboptimalities from a limited
ensemble size (Anderson and Anderson, 1999).
<table>
<thead>
<tr>
<th>Initial sample name</th>
<th>Mean/variance for ( c_a )</th>
<th>Mean/Variance for ( c_w )</th>
</tr>
</thead>
<tbody>
<tr>
<td>IS 1</td>
<td>0.7 / 0.14</td>
<td>0.25 / 0.05</td>
</tr>
<tr>
<td>IS 2</td>
<td>0.4 / 0.14</td>
<td>0.28 / 0.05</td>
</tr>
<tr>
<td>IS 3</td>
<td>0.6 / 0.14</td>
<td>0.25 / 0.05</td>
</tr>
<tr>
<td>IS 4</td>
<td>0.6 / 0.14</td>
<td>0.50 / 0.05</td>
</tr>
<tr>
<td>IS 5</td>
<td>0.8 / 0.14</td>
<td>0.30 / 0.05</td>
</tr>
</tbody>
</table>

Table 2: Mean and variance for the initial normal distribution of ocean and atmospheric from drags \( c_a \) and \( c_w \) used in experiments.

5 Experiments

In the present study, we use data assimilation to limit the time error growth of simulated iceberg trajectories by correcting the iceberg position once per day. In addition, we intend to estimate the time evolution of the ocean and air form drags over the simulated trajectories. The ensemble size needed by the EnKF and sensitivity of the method to the initial conditions are uncertain and need to be investigated. The skills of a data assimilation experiment are commonly tested, analyzing the Root Mean Squared (RMS) error and the ensemble spread. The RMS error is the geographical distance \( \epsilon \) between the measured position \( \mathbf{d} \) and the ensemble mean position \( \overline{\mathbf{X}} \) at a given time:

\[
\text{RMS} = \epsilon(\mathbf{d}, \overline{\mathbf{X}}),
\]

The ensemble spread is the standard deviation of the \( m \) ensemble members compare to the ensemble mean at a given time:

\[
\text{Spread} = \sqrt{\frac{1}{m-1} \sum_{i=1}^{m} \epsilon(\mathbf{X}_i, \overline{\mathbf{X}})}.
\]

In the following, the method is tested for different ensemble size (section 5.1), and for different initial distributions of the form drag parameters (section 5.2). Next, we try to extract information about the shape of each iceberg from the estimated values (section 5.3), and discuss the trajectory and the evolution of the parameters for each iceberg (section 5.4).

5.1 Sensitivity to the ensemble size

Theoretically, the ensemble size required by the EnKF to converge should be at least equal to the model subspace dimensions. The latter is usually large and unknown. Here, we tried to identify a sufficient ensemble size for reaching best accuracy at reasonable computational coast. For one iceberg (7089), we run simulations with ensemble size of 50 and 100, with IS 5 initial distribution (Table 2) for \( c_a \) and \( c_w \). Figure 1 shows that the RMS error and the ensemble spread over the simulation is relatively similar. The ensemble mean of the ocean and air form drag are also comparable (Figure 2). It seems that the error is slightly bigger using 100 members. Such behavior is common when the ensemble size gets larger than the model subspace. Hence, running simulations with only 50 members seems to be sufficient and will be used in the following.
5.2 Sensitivity to initial conditions

The stability of the experiment is assessed using different initial distributions of the ocean and air form drag coefficients for iceberg 7089 (Table 2). For each simulation we change either the mean value of the distribution of $c_a$ or the mean value of the distribution of $c_w$. The standard deviations remain equal to 0.14 and 0.05 for $c_a$ and $c_w$, respectively. The RMS error and the ensemble spread are relatively similar in the five experiments with some small discrepancies at the end of the simulations (Figure 3). Time series of $c_a$ and $c_w$ for the different simulations are shown in Figure 4. After 15 days, the form drags mean distributions ($c_a$, $c_w$) have a similar evolution over the time suggesting that the results are independent from the initial conditions. It is interesting to see that the variability of the RMS error is not directly correlated with the evolution of the form drags. Simulations with two different initial Gaussian distributions (not shown) for the other three icebergs also lead to similar conclusions; after a period of 15 days, the experiment is not very sensitive to the initial value of the form drags.

The sections 5.3 and 5.4 focus experiments run with 50 members and initial form drag parameter distribution from IS 5 sample, i.e. with an ensemble mean of 0.8 for $c_a$ and 0.3 for $c_w$.

5.3 Information on the iceberg form

Considering the very small dynamical impact of sea surface slope and sea ice for the four icebergs (Keghouche et al., 2009), we assume the four icebergs are driven by the atmospheric force, the oceanic force and Coriolis (Equation 4), and expect to collect information on the shape of the iceberg from the form drags. For all icebergs considered, we select the time when the model is no longer dependent on the initial conditions and when the model and observations are in good agreement in order to minimize the impact of the forcing error. For the four icebergs, the trajectory is initially complex with rapid changes in direction, but “stabilize” after 10 days with a southward drift for about 10 days. We therefore picked up the form drag values after 15 days of drift, i.e. on May 8 for 7086 and 7087 and on May 9 for 7088 and 7089. At this particular time the sea ice concentration was still over 80% (Figures 5, 6, 7 and 8),

Figure 1: Iceberg 7089 RMS error and ensemble spread with 50 and 100 members.
Figure 2: Sensitivity of evolution of ocean and atmospheric from drags to number of members for iceberg 7089. The initial distribution of the form drags is IS 5 (Table 2).

hence, we assume that the effect of waves was minor.

Based on the formulation of $c_a$ and $c_w$ in Equation 7, and assuming that the model errors $\epsilon_a$ and $\epsilon_b$ are minor, the ratio between $c_w$ and $c_a$ includes information on the iceberg shape and a model bias that is constant for all the icebergs. Table 3 shows the ratio between $c_w$ and $c_a$ for all the icebergs. They are relatively similar, except for iceberg 7087. Hence, we can expect that 7086, 7088 and 7089 have comparable form while, iceberg 7087 has a different one, with a large sail area compared to the vertical area in contact with ocean currents. Iceberg 7088 has slightly bigger ratio than 7086 and 7089, while $c_w$ is very close. It suggests that the sail area has less importance for this iceberg compare to 7086 and 7089. Note that adjustment of the mass can be analyzed when icebergs have the same form drag ratio and different ocean and atmospheric form drag coefficients. Despite the same ratio between $c_a$ and $c_w$ for icebergs 7086 and 7089, one can not comment on the mass adjustment since they have the same form drag values.

<table>
<thead>
<tr>
<th>Buoy</th>
<th>$c_a$</th>
<th>$c_w$</th>
<th>$c_w/c_a$</th>
</tr>
</thead>
<tbody>
<tr>
<td>7086</td>
<td>0.59</td>
<td>0.22</td>
<td>0.37</td>
</tr>
<tr>
<td>7087</td>
<td>0.76</td>
<td>0.10</td>
<td>0.13</td>
</tr>
<tr>
<td>7088</td>
<td>0.48</td>
<td>0.21</td>
<td>0.44</td>
</tr>
<tr>
<td>7089</td>
<td>0.60</td>
<td>0.22</td>
<td>0.37</td>
</tr>
</tbody>
</table>

Table 3: $c_a$ and $c_w$ estimated after 15 days for each simulated iceberg.

5.4 Analysis of the model parameters for each trajectory

In the previous section, we estimated ($c_a$, $c_w$) at a given time, when the error in the forcing was assumed to be small. These parameters evolve in time to compensate for errors in the forcing fields, melting inaccuracies and the parameterization. In the following, we try to relate the evolution of these parameters
5.4.1 Iceberg 7086

The results of the simulations of iceberg 7086 trajectory are shown in Figure 5. The mean distance error is 20 km. The Ensemble mean trajectory can be separated into four periods. The first period corresponds to the first 20 days, during which the ensemble mean trajectory is comparable to the observations, with a circular drift in the first 10 days. The first adjustment \((c_a, c_w)\) mentioned in the previous section is made at the end of this period. The second period is characterized by a rapid drift lasting about 10 days, 17-25 May. The initial position on the 17 May, is very close to the observations. It is clear from Figure 5 that the model has a tendency to deviate from the observed trajectory. Accordingly, the EnKF adjusts the ratio between \(c_a\) and \(c_w\), suggesting an error in the form. The third period is from 26 May to 17 June. Form 26 May to 13 June, the observed iceberg undergo cyclonic trajectories not represented in the model. The EnKF sets values of \(c_a\) and \(c_w\) close to 0, implying that the forcing is not beneficial. During the last period, from June 18 to the end of the trajectory the iceberg travels westward in agreements with the model forcing, but deviates to the right of the observed trajectory during the last days. The value of \(c_a\) increases rapidly, while \(c_w\) oscillates around the same value. An adjustment of the form is occurring. At the end of the simulation the modeled iceberg has lost 95% of its initial volume. Discrepancies in the mass may be linked to the melting parameterization. For example, calving deterioration events are not considered in this study.

5.4.2 Iceberg 7087

Simulations of iceberg 7087 are shown in Figure 6. The mean distance error is 24 km. Note that for this particular iceberg, the modeled ensemble mean trajectory stops on 22 June while the ARGOS buoy on the observed iceberg stops transmitting 13 days later, on 4 July. An observed iceberg is assumed...
melted if the ARGOS buoy stops transmitting while in the model, it is when any of its lateral dimension or freeboard height is less than one meter. These timing imprecisions can not explain discrepancies on the estimated iceberg lifespan. Hence, either the mass is underestimated or the melting is overestimated. The simulated sea ice concentration drops below 50% five days earlier than in observations. One can suspect that the erosion from the waves starts too early. However, the simulated iceberg 7086, which drifted in the same region at the same time, did not melt before the observed iceberg. Therefore, we suggest that the main reason for the precipitated melting for iceberg 7087 is the underestimation of its initial mass. Note as well that this iceberg has a relatively different shape compared to the three others (section 5.4). The melting parameterization is probably not suited for irregularly shaped icebergs. The modeled ensemble mean trajectory follows reasonably well the observations until 10 May. Then it starts to drift southward rapidly. From 23 May to June 3, the observed iceberg drifts erratically. The atmospheric forcing keeps trying to push the iceberg southward, and the EnKF constrains the iceberg trajectory by decreasing \( c_a \) drastically. The value of \( c_a \) even becomes negative indicating that a reversal of the atmospheric forcing would improve the prediction.

### 5.4.3 Iceberg 7088

The ensemble mean trajectory of 7088 iceberg shows very good agreement with the observed trajectory (Figure 7). The mean distance error is 15 km. The main features of the drift are represented despite a southwestward shift from the modeled trajectory from 1 June to 29 June. From 24 May to 10 June, both form drags are increasing suggesting that the mass is overestimated. Then, during 11-22 June period, \( c_a \) keeps on decreasing, while \( c_w \) increases slightly indicating an adjustment of the form. During 22-26 June period, the EnKF constrains both form drags to decrease, suggesting an adjustment of the mass in that period. The air form drag drops rapidly during 27-30 June period, implying that atmospheric forcing fields are erroneous at this particular time. The simulated sea ice concentrations decline below 50% from 30 May, 15 days earlier than in the observations which suggests that wave erosion may start too early in the model. However, the modeled iceberg lost 96% of its initial volume at the end of the

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**Figure 4:** Evolution of \( c_a \) and \( c_w \) for five different initial distributions of \( c_a \) and \( c_w \). The characteristics of each distribution are specified in Table 2.
Simulation.

5.4.4 Iceberg 7089

Results for iceberg 7089 are shown in Figure 8. As for the preceding iceberg, the ensemble mean trajectory of this iceberg shows very good agreement with observations all along the trajectory. The mean distance error is 13 km. From 26 May to 4 June, EnKF constrains a decrease of $c_a$ in order to adjust the sail area of the iceberg. At the end of the trajectory, both $c_a$ and $c_w$ increase, which might reveal that the mass is overestimated but the modeled iceberg melted less than a day before the end of the simulation. Note that sea ice concentration was 50% 10 days before the observations.

6 Conclusion

This study aimed at reproducing observed iceberg trajectories and estimating model parameters in order to gain some insight into the relative form of the observed icebergs and the accuracy of the forcing. This was achieved using an advanced data assimilation technique: the Ensemble Kalman Filter. This method is computationally more efficient than the classical Monte Carlo method used for parameter estimation in Keghouche et al. (2009) and allows for a time varying estimation of the parameters. The assimilation occurs once a day to limit the model error growth. It consists of updating the position of the iceberg, and estimating the air and ocean form drag coefficients. Simulations with different ensemble sizes and different initial conditions demonstrate the robustness of the method. We performed simulations of four observed trajectories from 1990, registered during the IDAP campaign (Spring, 1994) and drifting northwest of the Barents Sea. Assuming a simplified force balance between the Coriolis force and the form drag forces, we try to extract the iceberg characteristics. In the first part of the analysis, discrepancies between the optimal form drag with the four icebergs reveal that one (7087) was probably not tabular. Finally, we tried to relate the evolution of $c_a$ and $c_w$ for each iceberg with the known weaknesses of the model, e.g. inaccuracy in the melting, error in the forcing, processes not considered such as calving. The time evolution of the ocean and air form drag helps to identify the periods when the forcing fields were inaccurate or when the form/mass changed. It gives information on the relative geometries of the four icebergs.

The two northernmost simulated trajectories (7088 and 7089) have a precision of 15 km and the the two southernmost trajectories (7086 and 7087) have a precision better than 25 km over the two months of drift.

Acknowledgments

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References


Figure 5: Iceberg 7086. Top: Simulated ensemble trajectories (grey lines), ensemble mean trajectory (black line) and observed trajectory (magenta line). Filled circles are plotted every 10 days in magenta for the observation and in black for the ensemble mean trajectory. **Middle:** Time series of the distance between observed and simulated trajectories in grey. The RMS is shown in black. **Bottom:** Time series of the ensemble mean of form drag coefficients $c_a$ (black line) and $c_w$ (grey line). The iceberg volume proportion relative to its initial volume is shown in green. The blue (resp. red) vertical lines show the time when simulated (resp. observed SSM/I) sea ice concentration drop to 80% (solid line) and 50% (dotted line).
Figure 6: Iceberg 7087. Top: Simulated ensemble trajectories (grey lines), ensemble mean trajectory (black line) and observed trajectory (magenta and green line). The part of the observed trajectory shown in green was not simulated due to precipitated melting of the simulated iceberg. Filled circles are plotted every 10 days in magenta for the observation and in black for the ensemble mean trajectory. Middle: Time series of the distance between observed and simulated trajectories in grey. The RMS is shown in black. Bottom: Time series of the ensemble mean of form drag coefficients $c_a$ (black line) and $c_w$ (grey line). The iceberg volume proportion relative to its initial volume is shown in green. The blue (resp. red) vertical lines show the time when simulated (resp. SSM/I) sea ice concentration drop to 80% (solid line) and 50% (dotted line).
Figure 7: Iceberg 7088. Top: Simulated ensemble trajectories (grey lines), ensemble mean trajectory (black line) and observed trajectory (magenta line). Filled circles are plotted every 10 days in magenta for the observation and in black for the ensemble mean trajectory. Middle: Time series of the distance between observed and simulated trajectories in grey. The RMS is shown in black. Bottom: Time series of the ensemble mean of form drag coefficients $c_a$ (black line) and $c_w$ (grey line). The iceberg volume proportion relative to its initial volume is shown in green. The blue (resp. red) vertical lines show the time when simulated (resp. observed SSM/I) sea ice concentration drop to 80% (solid line) and 50% (dotted line).
Figure 8: Iceberg 7089. Top: Simulated ensemble trajectories (grey lines), ensemble mean trajectory (black line) and observed trajectory (magenta line). Filled circles are plotted every 10 days in magenta for the observation and in black for the ensemble mean trajectory. Middle: Time series of the distance between observed and simulated trajectories in grey. The RMS is shown in black. Bottom: Time series of the ensemble mean of form drag coefficients $c_a$ (black line) and $c_w$ (grey line). The iceberg volume proportion relative to its initial volume is shown in green. The blue (resp. red) vertical lines show the time when simulated (resp. observed SSM/I) sea ice concentration drop to 80% (solid line) and 50% (dotted line).