EFFECTS OF DATA ASSIMILATION OF ICE MOTION IN A BASIN-SCALE SEA ICE MODEL

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ABSTRACT
A sea ice thickness distribution model is modified to include data assimilation of satellite-derived sea ice motion. The model is applied to the Arctic basin for 1990 and 1998, two years with anomalously low summer sea ice extent and coverage. Daily ice area observations derived from SSM/I satellite data are used to initialize the model and for comparison with model-predictions. In both simulated years, the use of assimilated sea ice motion constrains the 15% sea ice extent closer to SSM/I-observed values, but excessive ice melt occurs within the central pack. Compared to the baseline, the assimilated ice motion field varies the dynamic transport of ice and thereby the production of open water and ridged ice, resulting in dramatic differences in the seasonal evolution of the ice pack. This suggests that the sea ice model is overly sensitive to changes in dynamic forcing. This has implications for the inclusion of data assimilation of sea ice motion in climate models.

INTRODUCTION
Historically, the treatment of sea ice in general circulation models has been, at best, based upon a relatively simple constitutive law describing a linear (e.g., Flato and Hibler, 1992) or elliptical (Hibler, 1979) yield curve. These parameterizations, mainly developed 20 years ago, are based on a limited amount of observations of ice motion primarily from drift buoys, considering the ice on the large scale to be a continuous fluid as opposed to a field of discrete floes and plates. Nevertheless, modeled fields of ice area, ice extent, and ice volume using these dynamic parameterizations compare favorably with observations. Moreover, the large-scale ice drift pattern agrees with that seen in the drift buoy record. Recently, techniques have been developed which allow sea ice motion to be obtained from the satellite record. While the errors in ice motion are larger than those obtained with buoy measurements, these data represent a substantial increase in the temporal and spatial coverage of sea ice motion observations in both polar regions. When averaged over months or years, the spatial pattern of ice drift varies smoothly. However, the

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day-to-day ice motion is much more discontinuous, with localized areas of shear and divergence consistent with the discrete floe nature of sea ice.

The use of data assimilation techniques (in this case, optimal interpolation) improves the hindcasting of sea ice motion, correcting for errors associated with the physical assumptions governing the model dynamics. Deficiencies in other model parameterizations may also be uncovered. Furthermore, there are implications for coupled models, as errors in the sea ice model will propagate into the atmosphere or ocean (see, for example, Maslanik et al., 2000).

Previous studies by Maslanik et al. (2000), Lynch et al. (2001), and ongoing research by A. Rinke (personal communication, 2002) investigate extreme sea ice retreat in 1990 (East Siberian Sea, e.g., Serreze et al., 1995) and 1998 (Beaufort Sea, e.g., Maslanik et al., 1999). These years are characterized by unusually warm surface air temperatures and central Arctic sea level pressures substantially lower than normal (e.g. Serreze et al., 1995). They conclude that while thermodynamic melt can account for the bulk of the ice retreat, there are several important factors that are crucial for the formation of the anomaly. In their model simulations, dynamic transport of the ice is crucial to its initiation and final distributions. However, internal ice physics (i.e., thermodynamics) are also important. Here, we revisit these studies, investigating the effect of assimilated sea ice motion on the simulation of these events.

**MODEL AND DATA DESCRIPTION**

The sea ice model used here is similar to that described by Bitz et al. (2000). It is implemented on a 109 × 109 50 km² Cartesian grid derived from the Equal Area Scalable Earth (EASE) projection (e.g., Armstrong et al., 1997). Each grid cell contains up to 5 ice categories each, with 2–4 internal temperature points, depending on the category thickness (Bitz and Lipscomb, 1999). A viscous-plastic elliptical rheology (e.g., Hibler, 1979) is used here, with the alternate-direction implicit solver employed (Zhang and Rothrock, 2000). Ridging and vertical redistribution of the sea ice within each grid cell are determined using formulation for ice strength (Rothrock, 1975) and ice divergence within each cell (Thorndike et al., 1975; Hibler, 1980). A 3-hour timestep is used.

Forcing data are obtained primarily from the POLES sea ice model forcing data set (Zhang et al., 1998). These data consist of winds derived from NCEP numerical models, while the radiative fields are based on in situ observations of surface air temperature. Downwelling shortwave and longwave values in this study are derived using relations from Parkinson and Washington (1979) using daily spatially-varying values of cloud amount based on AVHRR observations (Maslanik et al., 1997, 2001) rather than the spatially-invariant climatological monthly means used in POLES. For the shortwave radiation, a diurnal cycle is simulated by calculating fluxes every time step based on position (latitude) and time of day (mean hour angle) for each point. Basin mean calculated values compare well with point observations from the SHEBA field project (figure 1). Ocean currents and heat flux from the POLES dataset are utilized. The ocean data are based on ice-ocean model simulations using the POLES atmospheric forcing (Zhang et al., 1998).

The ice model is initialized using observations of ice area from SSM/I measurements (Comiso, 1999) for 1 Jan. 1990 or 1 Jan. 1998. Ice thicknesses are obtained from mean
January values of a climatological simulation by Maslanik et al. (2001). Where SSM/I observations are unavailable, the climatological mean area is used. When data assimilation is employed, daily ice motion data derived from SSM/I (Emery et al., 1997; Maslanik et al., 1997) are incorporated into the model. The observed ice motions are not directly substituted; rather, the observed and calculated ice motions are blended using an Optimal Interpolation scheme (Meier et al., 2000; Meier and Maslanik, 2001). The model first calculates an ice motion field based on the existing distribution of thickness and area. This motion field is then blended with the calculated field.

RESULTS
We first consider the summer ice edge retreat. As seen in figure 2, in 1990, the observed meltback first appears in the East Siberian Sea and north of the Bering Strait. The ice edge, defined to contain a region in which ice-covered area is at least 15%, retreats northward, and by the summertime minimum is a continuous swath extending the Laptev Sea to the Chukchi Sea, with a narrower strip of open water reaching through the Beaufort Sea to the Canadian Archipelago. These observations are not well reproduced by the baseline model case for 1990. The ice extent is too far south at midsummer (day 180), and although the Bering Strait meltback is
reproduced well, the ice edge in the Chukchi Sea never retreats from the Siberian coast. The assimilation case fares better in that regard, although a tongue of ice remains in the Chukchi Sea. However, areas of unrealistically low ice concentration appear in the central ice pack due to excessive ice melt. Their cause will be discussed in the next section.

The results for 1998 are similar (figure 3). The Observations show that meltback first occurs in the Beaufort Sea, progressing northward and westward to the Bering Strait by the summertime minimum at day 240. The baseline case performs better this year than in 1990, but the ice edge still does not retreat as far as observations. The simulated ice edge in the assimilation case is better in places, but overall too much ice melts in the Beaufort Sea and, as in the 1990 simulation, holes again appear in the central ice pack. Thus, the effect of data assimilation on the modeled solution is mixed; it improves the ice edge calculation but has detrimental effects on the calculation of ice area (the total amount of ice cover, excluding leads and open water) in the interior. This is illustrated in figure 4. The basin total ice-covered area in the baseline cases (gray lines) of both 1990 and 1998 follows closely with the SSM/I observations. The area in the assimilation cases decreases to nearly half that of the observations or the baseline cases. (The Arctic basin is defined, as in Gloersen et al. (1992), to include all of the western Arctic Ocean north of Bering Strait and the eastern Arctic north of Fram Strait, excluding the Barents and Kara Seas.)

![Figure 3: Same as figure 1, for 1998.](image)

![Figure 4: Total ice-covered area (excluding leads) in the Arctic basin for (a) 1990, (b) 1998: SSM/I (dotted), Baseline (gray), Assimilation (black). Missing points in the SSM/I observations were excluded from the model means.](image)
DISCUSSION

It would seem that for the present model formulation, the use of assimilated sea ice motion improves some aspects of the simulation but results overall in a poorer reproduction of the ice area than for the baseline runs. Because the only difference in external forcing between the baseline and assimilation cases in each year is the assimilation of ice motion vectors (see figure 5), the altered ice motion must have a secondary effect that alters the modeled ice properties. In this model, as in typical basin-scale dynamic sea ice models, sub-grid scale ridging and open water production are directly proportional to the large-scale horizontal dynamic transport of ice, i.e., its convergence and divergence (e.g., Rothrock, 1975; Hibler, 1980). Therefore, given identical initial conditions and thermodynamic forcing and varying only the ice drift field, a different rate of ice divergence is calculated, resulting in a different amount of ridged ice or open water production at a point and timestep. As the model progresses forward, this feeds back into the model thermodynamics, as different distributions of ice thicknesses and areas within a grid cell result in different growth/melt rates. The effects of this can be seen in the ice volume (figure 6), where in both 1990 and 1998, the assimilated case has more ice volume (thicker ice overall) in the basin in the first half of the year. In the summer, the rate of ice melt is greater, resulting in less ice volume (thinner ice overall) for the remainder of the year. The result is a positive feedback accentuated by overprediction of lateral and basal melt and/or heat absorption within the open water area. It is also possible that random noise in the motion data overestimates divergence, but this noise is substantially reduced via the assimilation (Meier et al., 2000).

![Figure 5: Mean ice drift speed in the Arctic basin for (a) 1990, (b) 1998: Baseline (gray), Assimilation (black).](image)

We stress that the difference between modeled and assimilated ice motion fields on a daily basis is locally very pronounced. The effects on ice volume vary spatially as well; there is not a uniform increase or decrease in any calculated ice field across the Arctic basin. Thus, the effects of constrained ice motion fields affect other calculated sea ice fields, and would undoubtedly propagate into the atmosphere or ocean in a coupled model scenario. Considering that the ice motions in the assimilated case have been altered (and are more realistic) than in the baseline case, which employs an idealized elliptical rheology, their inclusion has revealed a strong sensitivity of the ice model to variability in the ice drift. Previous results have shown similar sensitivities of basin-scale dynamic ice models to perturbations in wind forcing (e.g., Arbetter et al., 1999). Given the large variability of atmospheric datasets (e.g., Walsh et al., 1998; Randall et al., 1998) and the uncertainty in some sea ice physical parameterizations (e.g., Arbetter et al., 1997), ice models are typically tuned to produce reasonable results for a particular forcing dataset, rather than a variety of inputs, and to allow components of the model to compensate for shortcomings elsewhere in the model.
SUMMARY

A basin-scale sea ice model of a level of complexity similar to those used in coupled climate models of regional and global scale is combined with an optimal interpolation ice motion data assimilation scheme. Simulations are performed investigating anomalous Arctic summer ice retreat in 1990 and 1998, following results presented in Maslanik et al. (2000) and Lynch et al. (2001) indicating that correct dynamic transport of sea ice is essential to the accurate reproduction of the observed ice conditions. The modeled ice edges with assimilated sea ice motion are more consistent with those seen in SSM/I-derived observations than are the non-assimilated (baseline) cases, but the modeled ice-covered area is substantially less than that seen in either the baseline cases or observations. It is shown that, due to the varied dynamic transport feeding back into other model components, the altered sea ice motion fields have a cumulative effect on the ice volume and thereby the ice area.

Following this, we believe that the ice model is tuned to give reasonable results given an idealized ice dynamic parameterization and that aspects of the thermodynamic treatment are overly sensitive to perturbations in the sea ice drift. Thus, while there has been considerable progress in the modeling of sea ice, more work is needed to improve its response given inaccuracies in the measurement of other climate components (e.g., Randall et al., 1998). While we have not improved the overall simulation of Arctic ice retreat in 1990 (East Siberian Sea) or 1998 (Beaufort Sea), we have demonstrated the utility of using assimilated ice motion to identify aspects of ice models that may not be as accurate as assumed or that at least need to be investigated further using more detailed and accurate ice motion data.

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