Sea-Ice Deformation Rates from Satellite Measurements and in a Model

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ABSTRACT *The deformation of sea ice is an important element of the Arctic climate system because of its influence on the ice-thickness distribution and on the rates of ice production and melt. New data obtained from the Radarsat Geophysical Processor System (RGPS) using satellite synthetic aperture radar images of the ice offers an opportunity to compare observations of the ice deformation to estimates obtained from models. The RGPS tracks tens of thousands of points, spaced roughly at 10-km intervals, for an entire season in a Lagrangian fashion. The deformation is computed from cells formed by the tracked points, typically at 3-day intervals. We used a coupled ice/ocean model with ice-thickness and enthalpy distributions that covers the entire Arctic Ocean with a 40-km grid. Model-only and model-with-data-assimilation runs were analysed. The data assimilation runs were analysed in order to determine the validity of the comparison techniques and to find the comparisons under the best of circumstances, when many buoy measurements are available for assimilation. This step is necessary because the RGPS and model data differ in spatial and temporal sampling characteristics. The assimilated data included buoy motion and Special Sensor Microwave/Imager (SSM/I)-derived ice motion. The Pacific half of the Arctic Basin was analysed for a 10-month period in 1997 and 1998. Comparisons of ice velocity observations to the modelled velocities showed excellent agreement from the model-with-data-assimilation run but poorer agreement for the model-only run. At a scale of 320 km, the deformation from the data assimilation run was in modest agreement with the observations but where many buoys were available for assimilation the agreement was quite good. Both model runs showed poor agreement during summer. Comparisons of the deformation distribution functions suggest why the agreements were poor even though the velocity agreements were good. Decreasing the ice strength parameter in the model improved the deformation comparisons for the model-only runs.*

RÉSUMÉ [Traduit par la rédaction] *La déformation des glaces de mer est un élément important du système climatique de l'Arctique, car elle influe sur la répartition de l'épaisseur des glaces et sur leur vitesse de formation et de fonte. De nouvelles données fournies par le système de processeur géophysique de RADARSAT (RGPS) à l'aide d'images de la glace prises par radar à antenne synthétique à bord d'un satellite permettent de comparer la déformation des glaces observée et la déformation évaluée à partir de modèles. Le RGPS suit des dizaines de milliers de points espacés plus ou moins de 10 km et ce, pendant une saison entière et de façon lagrangienne. La déformation est calculée à partir de cellules formées par les points suivis typiquement à des intervalles de trois jours. Nous avons utilisé un modèle couplé glace/océan avec distribution de l'épaisseur et des enthalpies des glaces qui couvre l'ensemble de l'océan Arctique avec une grille de 40 km. Les résultats du modèle seul et du modèle avec assimilation des données ont été analysés. Les résultats du modèle avec assimilation des données ont été analysés afin de vérifier la validité des techniques de comparaison et de trouver des comparaisons dans les meilleures circonstances, lorsque nous disposons de nombreuses mesures de bouées pour fins d'assimilation. Cette étape est nécessaire puisque les caractéristiques d'échantillonnage spatial et temporel du RGPS et du modèle sont différentes. Les données assimilées comprennent des données sur le mouvement des bouées et des données sur le mouvement des glaces tirées du capteur hyperfréquences spécialisé/imageur (SSM/I). La moitié du bassin Arctique qui se trouve du côté du Pacifique a été analysée pendant une période de 10 mois en 1997 et 1998. Les vitesses des glaces observées et les vitesses modélisées ont été comparées; la concordance est excellente dans le cas du modèle avec assimilation des données, mais moins bonne dans le cas du modèle seul. À une échelle de 320 km, la concordance entre la déformation modélisée avec assimilation des données et la déformation observée est modeste; toutefois, lorsque les données à assimiler proviennent de nombreuses bouées, la concordance est très bonne. Pour les deux modèles, nous avons constaté une mauvaise concordance pendant l'été. La comparaison des fonctions de répartition de la déformation expliquerait pourquoi la concordance est mauvaise même si la concordance entre les vitesses est bonne. En réduisant le paramètre de la force des glaces dans le modèle, nous avons amélioré les comparaisons de la déformation dans le cas du modèle seul.*

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1 Introduction

Sea-ice deformation is a fascinating and unique component of the Arctic geophysical environment. The deformation rate of pack ice, determined from the spatial gradients in the velocity, is a key parameter in determining the formation of leads and open water, as well as the formation of rafted and ridged ice. The amount of open water and the thickness of the ice are key parameters in the climate system because of the strong effect ice thickness has on albedo, heat exchange, and ice growth rates. In sea-ice models, the ice velocity is established through a balance of forces that depends on the winds and currents (forcing), the model state (mean thickness), the model physics (drag and constitutive laws), as well as the model resolution. Accurate modelling of the ice velocity and deformation rate is essential if ice is to be properly represented in climate models.

It is possible to test the mean motion of ice calculated from models by comparing it to the trajectories of buoys and drifting ice stations (e.g., Thomas, 1999; Zhang et al., 2003; Meier et al., 2000). But it has not been possible to test ice deformation computed in the models adequately for lack of an appropriate comparison dataset. The buoys routinely deployed in the Arctic are too few and too widely separated to measure ice deformation accurately at small scales. Thomas (1999) compared buoy-derived and model-derived deformation estimates at large scales, 400 to 600 km. The model he used is similar to the one used here. He investigated different model wind drag formulations as well as a best-fit linear model based only on the geostrophic wind. He found modest correlations for vorticity and shear but the correlations of divergence were statistically insignificant. The best correlations were found for the simple linear model. There was insufficient data to determine the spatial or temporal variability of the correlations. However we now have a new dataset, based on the tracking of ice motion in satellite radar backscatter images, that offers an excellent opportunity to compare modelled and measured deformation rates over a wide area of the Arctic Ocean, for all seasons of the year, and over a wide range of scales.

This new dataset is from the Radarsat Geophysical Processor System (RGPS) (Kwok, 1998; Kwok et al., 1999). The RGPS uses synthetic aperture radar images from the Canadian Radarsat satellite to track thousands of points during a full season. The points are on an initial 10-km grid and are tracked in a Lagrangian fashion, typically at 3-day intervals. The trajectories are grouped into 4-point cells, and the area changes and strain rates of the cells are computed each time the positions of the four corners are observed.

Here we seek to determine how well comparisons can be made between two very different datasets and to determine whether the comparisons can be used to improve the model simulations. To do this we compare the RGPS strain-rate estimates to those obtained from a coupled ice/ocean model that is run in two modes: with and without assimilation of observed ice displacement measurements from buoys and from the Special Sensor Microwave/Imager (SSM/I) passive-microwave satellite sensor. The data assimilation runs are

included in order to assess the level of correspondence that can be expected under the best of circumstances, when the model is able to assimilate many buoy velocity measurements. This comparison will help to assess the effects of the interpolation and smoothing required to align the RGPS and model datasets. The assimilation techniques, based on optimal interpolation, are presented in Zhang et al. (2003) (hereinafter, Z2003). They compare model ice velocity to buoy motions and compare ice-thickness outputs to submarine ice-draft measurements. Here we extend that study to include comparisons of modelled deformation rates to the RGPS observations. We concentrate on the model runs analysed in Z2003, but we also show results that indicate how the model ice strength can be reduced to improve the comparisons.

Spatial scale is a central concept of ice deformation because the ice velocity field is spatially discontinuous. The estimated deformation is a function of the area over which the determination is made and, in general, the deformation rate is reduced as the area is increased. The model grid size is 40 km, so that the smallest area for the deformation rate estimates is a 40×40 -km square that includes approximately 16 of the 10-km RGPS cells. We concentrate our efforts on four spatial scales: 40, 80, 160 and 320 km. The area covered is the square of the scale.

In the following, we first review the RGPS dataset, the model characteristics, the data assimilation procedures, and how the datasets are processed to produce matched pairs of measurement and model estimates. The datasets are then compared as a whole and as functions of scale, location and time. An example is given of how a model parameter, the ice strength, can be changed to improve the comparison. Differences found in the comparisons with and without data assimilation and the impact of buoy density on the deformation correlations are discussed.

2 Data sources and model description

a RGPS

The RGPS begins with sequential 100-m resolution Radarsat backscatter images of the ice. A maximum-correlation technique is used to determine ice displacement (Fily and Rothrock, 1990; Kwok et al., 1995; Kwok, 1998). Processing is performed on one of two systems. One is operated by the Polar Remote Sensing Group at the Jet Propulsion Laboratory (JPL) and a second at the Alaska Synthetic Aperture Radar (SAR) Facility (ASF). The data are available on the web site www-radar.jpl.nasa.gov/rgps/radarsat.html.

The displacements from image to image determine the trajectory of a point on the ice. The trajectories are grouped into cells. Each cell consists of four corner points. Additional trajectories are initiated at the mid-point of a cell edge if the side of a cell becomes more than twice its initial length. The cell area and the four components of the velocity gradient

$\left(\frac{\partial u}{\partial x}, \frac{\partial u}{\partial y}, \frac{\partial v}{\partial x}, \frac{\partial v}{\partial y} \right)$ are determined from an approximation of

the line integral around the boundary of the cell, similar to what is outlined below for the determination of the model-based deformation.

The tracking accuracy for the individual trajectories is excellent. We conducted a validation study (Lindsay and Stern, 2002), and found that the displacements measured by the RGPS match those measured by buoys with a squared correlation of 0.996. The absolute tracking error is 286 m or less. Errors in the deformation estimates arise from within-image tracking errors that are insensitive to the small geolocation errors of the images; for these purposes they are estimated at 100 m. The error in the deformation comes from both tracking errors and boundary-definition errors, which arise from the approximation of the velocity along a material boundary of a cell from the velocity at just the four corners. The combined error for the 10-km cells is estimated to be 3.5%. The errors are reduced by a factor of $1/N^{3/4}$ when averaging over N cells. Here, large areas are considered with between 16 and 1000 cells (40 to 320 km). At these scales the accuracy of the RGPS deformation estimates is very good (0.43 – 0.02%).

The RGPS data have strengths and weaknesses that must be considered in a comparative study with model ice velocity and deformation. The temporal and spatial sampling is complex and variable. Most trajectories are tracked with three-day intervals, but the intervals may be less than one day or more than 10 days. The spatial sampling is initially in 10-km squares, but over the course of the season the squares drift and deform; some cells become very large, small, or distorted. Highly distorted cells may even have computed areas that are negative. But the spatial coverage is excellent. While the entire Arctic Basin is not yet available, a large region of the Arctic Ocean including the Beaufort, Chukchi, and Lincoln seas is available with coverage ranging from 2.5×10^6 km² to 4.8×10^6 km². The temporal coverage extends from November 1996 to May 1999 and is expanding as processing continues at both JPL and ASF. Here we concentrate on the November 1997 to August 1998 period, corresponding to most of the year of the Surface Heat Budget of the Arctic (SHEBA) field experiment (Perovich et al., 1999). The number of cells observed and the area covered decrease significantly in August because of difficulties in tracking the ice under the low-contrast backscatter conditions of summer.

b Model

The model runs used here are the same as those analysed in Z2003. The thickness and enthalpy distribution sea-ice model consists of five main components: a momentum equation that determines ice motion, a viscous-plastic ice rheology with an elliptical yield curve that determines the relationship between ice deformation and internal stress, a heat equation that determines ice growth or decay and ice temperature, two ice-thickness distribution equations for deformed and undeformed ice that conserve ice mass, and an enthalpy distribution equation that conserves ice thermal energy. The first two components are described in detail by Hibler (1979). The ice momentum

equation was solved using Zhang and Hibler's (1997) numerical model for ice dynamics. The heat equation was solved, over each category, using Winton's (2000) three-layer thermodynamic model, which divides the ice in each category into two layers of equal thickness beneath a layer of snow. The ice-thickness distribution equations are described in detail by Flato and Hibler (1995).

The model domain covers the Arctic, Barents, and Greenland-Iceland-Norwegian (GIN) seas. It has a horizontal resolution of 40 km $\times$ 40 km, 21 ocean levels, and 12 thickness categories each for undeformed ice, ridged ice, ice enthalpy, and snow. The ice-thickness categories, the model domain, and bottom topography can be found in Zhang et al. (2000).

Daily surface atmospheric forcing from 1992 to 1998 was used to drive the model. The forcing consists of geostrophic winds, surface air temperature, specific humidity, and longwave and shortwave radiative fluxes. The geostrophic winds are calculated using the sea level pressure (SLP) fields from the International Arctic Buoy Program (IABP) (see Colony and Rigor, 1993). The 2-m surface air temperature data are derived from the IABP-POLES 2-m Air Temperature dataset which is derived from buoys, manned drifting stations, and land stations (Rigor et al., 2000). The specific humidity and longwave and shortwave radiative fluxes are calculated following the method of Parkinson and Washington (1979) based on the SLP and air temperature fields. Model input also includes river runoff and precipitation detailed in Hibler and Bryan (1987) and Zhang et al. (1998).

Velocity and deformation estimates are determined both from a model-only (MO) run and a data assimilation (DA) run. Daily buoy motion data and SSM/I-85-GHz, 2-day ice motion data from 1992 to 1998 are used for data assimilation. The daily buoy data were provided by the IABP (Colony and Rigor, 1993). On any given day, there are 10 to 30 buoy motion measurements irregularly and sparsely distributed in the Arctic. The 2-day SSM/I ice motion dataset was provided by the JPL Remote Sensing Group (Kwok et al., 1998). The SSM/I data are gridded and have better spatial coverage than the buoy data, but the number of available data vary temporally (with no coverage in summer from 1 June to 30 September), and the error standard deviation is about eight times larger than for the buoys: 0.058 m s^{-1} vs. 0.007 m s^{-1} (Thorndike and Colony, 1980; Kwok et al., 1998).

Data assimilation of the buoy and SSM/I displacement observations is accomplished through optimal interpolation at each daily time step. The correlation function assumed for the ice velocity allows for the possibility of divergent flow with separate formulations for the correlation parallel and perpendicular to the line of separation between two points. Details are given in Z2003. The Z2003 study considered assimilations of just buoy measurements and just SSM/I measurements as well as a combination of both. The velocity and the ice-thickness comparisons are best for the combined assimilations, so, for simplicity, we consider just the combined assimilation case here.

c Computing the Deformation

The strain rate tensor of the velocity has three invariants

$$divergence = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}, \quad (1)$$

$$shear = \left[\left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial u}{\partial y} - \frac{\partial v}{\partial x} \right)^2 \right]^{1/2}, \quad (2)$$

and

$$vorticity = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}. \quad (3)$$

The total deformation rate is the quantity

$$total\ deformation = [shear^2 + divergence^2]^{1/2}. \quad (4)$$

The RGPS determines the components of the tensor by calculating line integrals around the boundary of each cell. The components for a collection of RGPS cells are obtained from area-weighted averages of the individual cell components.

The RGPS measures the total derivatives of the velocity by following individual elements of ice in a Lagrangian fashion, and does not measure the partial derivatives for fixed locations as estimated by the model. This distinction is unimportant when the size of the region analysed is large compared to the net displacement of the elements. The mean speed of the ice is on the order of 5 km d⁻¹, and over three days this is 15 km, a substantial portion of the smallest scale analysed, 40 km. Occasionally the mean velocity is much larger; since RGPS measures the total derivative, not the partial derivative at a fixed location, this may contribute to the discrepancy between the two estimates at the smallest scales.

To obtain the elements of the strain-rate tensor from the model velocity fields, an approximation is used of the line integral around the outside of a set of model grid points. If (x_i, y_i) are the locations and (u_i, v_i) the velocity components for n points forming the boundary of a region (the indices increase when proceeding counter-clockwise around the cell and $x_{n+1} = x_1, y_{n+1} = y_1$, etc.), then

$$\frac{\partial u}{\partial x} = \frac{1}{2A} \sum_{i=1}^n (u_{i+1} + u_i)(y_{i+1} - y_i) \quad (5)$$

with the other derivatives formed in a similar manner. The area is given by

$$A = \frac{1}{2} \sum_{i=1}^n (x_i y_{i+1} - y_i x_{i+1}). \quad (6)$$

The area bounded by the set of points determines the scale S of the deformation measurement, $S = \sqrt{A}$.

d Aligning the Datasets

Some interpolation and smoothing in time and space are required to account for the irregular sampling of the RGPS

data. A two-step process is used. First, daily estimates of the deformation from both the observations and the model are obtained at all model grid points. For a target model day and grid point, all RGPS observations are selected whose centre positions are within a range of $S/2$ (where S is the scale of the estimate) in both the x and y directions. The observation interval must be between 1 and 4 days, the initial time must be before the start and the final time must be after the end of the target day, and the size of any cell at the time of the initial observation must be between 20 and 180 km², thus avoiding cells that may have near-zero or negative areas. The observation intervals in the sample may not all be the same, but all intervals include at least the entire span of the target day. In practice, many of the observation intervals for a particular time and location are identical because of the large spatial extent of the backscatter images that form the raw data for the RGPS (see Fig. 1a). The area-weighted average of the strain-rate tensor components and the mean centre time of the observations are then found for all of the selected observations. At least half of the possible area must be represented for the observation to be retained. Note that because of the nominal 3-day sampling, most of the mean centre times will be constant for three days and then jump three days (see Fig. 1b).

The model deformation is computed daily by first smoothing the velocity components with a 3-day running-mean filter. At each grid point the deformation is then found from the line-integral approximation applied to the smoothed velocity components for points within $S/2$ of the target point in each coordinate as in Eq. (5). If any of these points includes land, the deformation for the point is not determined.

The second step accounts for the irregular observation times by grouping both the RGPS and the model daily values to remove what amounts to redundant RGPS observations created by the daily sampling. The resampling consists of averaging the time, velocity, and strain-rate information of points for which the RGPS centre times differ by less than one day. The model velocity and deformation values are then interpolated to the average RGPS centre times. This final dataset consists of matched pairs of 3-day RGPS and model velocity and deformation values: pairs matched in time, in location, and in spatial scale (see Fig. 1c).

The correlation between the velocity estimates from the two datasets increased after the resampling procedure because of a more accurate temporal matching of the model and RGPS data. Before the alignment the correlation was $R^2 = 0.76$ ($N = 218,620$) and after the alignment the correlation increased to $R^2 = 0.83$ ($N = 96,534$). The number of samples was reduced because of the grouping of observations in the resampling process.

To reiterate, the model velocity values are smoothed to 3-day intervals, the deformation is computed at spatial scales ranging from 40 to 320 km, and the deformation is resampled in time to match the RGPS observations. The RGPS samples are also computed for spatial scales ranging from 40 to 320 km, and redundant observations are removed by the resampling process. Figure 1 shows a sample time series for

Fig. 1 Illustration of the resampling process for a sample time series at a single fixed 40-km grid point over a 50-day period: a) the RGPS sample times showing cells moving through the region; b) the daily RGPS and the daily smoothed model u component of the ice velocity; and c) the resampled RGPS and model u component. The model results are from the data assimilation model run.

a single grid point and just one component of the velocity. The RGPS sample times are mostly the same for the different cells so the velocity values remain constant for about 3-day periods. The resampled time series removed the redundant observations. For this 50-day sample, the correlation between the datasets of the u component increased from 0.76 for the daily time series to 0.94 for the resampled time series.

3 Results of the comparisons

The mean of the total deformation in winter decreased slowly with the size of the area over which it was calculated (Fig. 2) while the correlations between the RGPS measurements and the model estimates increased. However, the velocity correlation was not strongly dependent on the scale because these scales are small compared to the autocorrelation length scale of the ice velocity (about 700 km). Since deformation events are often highly localized, the improvement in the correlation with increasing scale may be due, in part, to mislocations of deformation events in the model. With spatial averaging, these mislocations are less significant in reducing the correlation. Because the correlations

Fig. 2 Scale dependence of the total deformation and correlations of the invariants during winter: a) mean total deformation for the RGPS measurements and for the model over the entire domain; and b) the correlation of the model velocity, vorticity, shear, and divergence with RGPS measurements. Model values are for the run with data assimilation. The velocity correlation is for vectors expressed as complex numbers.

were best at 320 km, we concentrate on that scale in most of the rest of the paper.

We begin by reviewing the basic statistics of the velocity and deformation rate as measured on a scale of 320 km by the RGPS and as estimated by the model without and with data assimilation. In winter (November through May) there were a total of $N = 86,664$ observation pairs and in summer (June through August) a total of $N = 29,657$. These are not independent samples because their spacing is the same as the model grid, 40 km, even though the spatial scale of the velocity averages and the deformation determinations is 320 km. The mean and standard deviation over all times and locations of the ice speed and the four invariants of the deformation rate are shown in Table 1 for winter and summer. The correlation values for the velocity and the deformation rates are shown in Table 2.

a Mean Ice Velocity Comparisons

The mean speed of the ice, as measured by the RGPS, was a little higher in summer than in winter but the variability is a little lower. The MO simulation had lower winter mean ice speeds than measured by RGPS while the DA simulations showed a much better match to the observations. This is consistent with the findings of Z2003 in comparisons with unasimulated buoys. However, in summer the mean speeds of both model runs matched the observations quite well. The

TABLE 1. Mean and standard deviation of the speed and the strain-rate invariants at a spatial scale of 320 km and a temporal interval of 3 days, as measured by the RGPS and as estimated by the model with and without data assimilation. Winter is November to May, summer is June to August.

	Speed (m s ⁻¹) Mean, Stdev.	Shear (% d ⁻¹) Mean, Stdev.	Divergence (% d ⁻¹) Mean, Stdev.	Vorticity (% d ⁻¹) Mean, Stdev.	Total Def. (% d ⁻¹) Mean, Stdev.
Winter:					
RGPS	0.051, 0.039	0.87, 0.60	0.02, 0.30	-0.66, 1.45	0.91, 0.61
Model only, MO	0.040, 0.039	0.74, 0.83	0.07, 0.24	-0.47, 1.10	0.77, 0.84
Data assimilation, DA	0.051, 0.035	1.10, 0.78	0.05, 0.43	-0.52, 1.29	1.19, 0.78
Summer:					
RGPS	0.060, 0.035	1.13, 0.62	0.00, 0.40	-0.32, 1.66	1.20, 0.62
Model only, MO	0.059, 0.034	1.17, 0.71	-0.01, 0.48	-0.31, 1.35	1.27, 0.72
Data Assimilation, DA	0.061, 0.033	1.46, 0.76	0.03, 0.67	-0.33, 1.67	1.62, 0.75

TABLE 2. Correlation of RGPS and model values of the velocity and strain-rate invariants at a spatial scale of 320 km and a temporal interval of 3 days.

	Velocity*	Shear	Divergence	Vorticity	Total Deformation
Winter:					
Model only, MO	0.63	0.14	0.32	0.37	0.15
Model with DA	0.95	0.62	0.47	0.78	0.63
Summer:					
Model only, MO	0.77	0.15	0.38	0.58	0.12
Model with DA	0.89	0.35	0.51	0.70	0.35

*The velocity correlation is for vectors represented as complex numbers.

velocity correlation was excellent for the DA run in winter. This is no surprise, since the tracking accuracy of the RGPS is excellent and because there are sufficient buoys within the autocorrelation length scale of most comparison points to constrain the model velocity to be near the true ice velocity. Furthermore, the high correlation in the winter indicates that the alignment techniques used to compare these very different datasets were effective. Interestingly, the correlation with data assimilation was diminished from winter to summer, but without data assimilation it improved. The diminished correlation in summer for the DA simulation may be related to the loss of the SSM/I ice motion data. For the MO case, the correlation was higher in summer than in winter because the winter model pack was more locked up, with lower mean speed and deformation than observed, while in summer the MO speed matched the observed speed quite well. The velocity correlations for both the MO and the DA runs were somewhat higher than found for daily unassimilated buoy motions by Z2003; the likely reason is the temporal and spatial smoothing applied to this dataset.

The spatial patterns of the velocity correlation for the MO and the DA runs, winter and summer, are shown in Fig. 3. The model with no data assimilation showed lower correlations in all regions, while the model including data assimilation showed markedly increased correlations. As also seen in Table 2, the velocity correlation for the MO run increased substantially in summer but was still less than that of the DA run, which decreased from winter values. The DA correlation was highest in regions with many buoy reports available for assimilation. The maximum is $R = 0.98$, showing that the

velocity comparisons can be excellent where the model is forced to match the ice motion as measured by the buoys. Figure 4 shows a map of the total number of daily buoy velocity observations used for assimilation. The large number of observations in the Beaufort Sea is related to the SHEBA field experiment. However, the correlation remained good even in many regions where there were few buoy observations, except along the coasts in summer. In the summer the maximum correlation was $R = 0.88$, indicating more difficulty assimilating buoy data in the summer, a result also seen in Z2003 with just the buoy data. The reduction in correlation along the coasts may have been due, in part, to the assimilation scheme, which reduces the influence of the observations within 200 km of land. Another additional cause could be the model's incorrect accounting of the shear stresses near the coast.

b Deformation Rate Comparisons

The deformation rate, as seen in the mean shear or the total deformation, was larger in summer than in winter. The vorticity showed the largest variability of the deformation invariants in both seasons. There was net anticyclonic rotation. In winter the MO simulation underestimated and the DA simulation overestimated the total deformation. Again, this is because the MO simulation had a mean speed that was too low. In summer the mean total deformation was substantially better for the MO simulation than for the DA simulation, although the correlation, while small, was somewhat better for the DA simulation.

The deformation rates found here are somewhat smaller than the values reported by Thorndike (1986, Table 5) for the Arctic Ice Dynamics Joint Experiment (AIDJEX) array of buoys. He reports a length scale for the array of 800 km and uses daily buoy velocities to find mean shear values of $1.0\% \text{ d}^{-1}$ for winter and $1.6\% \text{ d}^{-1}$ for summer, compared with the present results from the 320-km 3-day RGPS data of $0.87\% \text{ d}^{-1}$ for winter and $1.13\% \text{ d}^{-1}$ for summer. The smaller values found here could be due to the longer averaging time in the RGPS data but could also be a result of sampling different years and different regions.

Of the different measures of the deformation rate, the correlation was highest for the vorticity. This is because the standard deviation of the vorticity was largest (Table 1) and

Fig. 3 Correlation of the model ice velocity with the RGPS ice velocity estimates for model runs without (model only, MO) and with data assimilation (DA) for a scale of 320 km. The data assimilation included buoy 1-day and SSM/I 2-day ice displacement measurements. The colour scale represents the range $R = 0.5$ to 1.0 . Winter is defined here as November 1997 to May 1998, and summer as June 1998 to August 1998. Land masses are identified in the first panel and NP is the North Pole.

because large-scale solid-body rotation, which can be well represented by a wind-driven model, contributed a significant amount of the vorticity variability. In winter, the correlations were markedly higher for model runs that include data assimilation, as would be expected. The summer DA model run showed lower correlations than winter and there was not as much improvement in the summer correlations of the DA over the MO runs. A possible reason is that in winter the coasts and a few buoy motions put significant constraints on the deformation, while in summer, when the ice pack is nearly in free drift, a few buoy motions do not constrain the deformation significantly. Also, the summer DA model run did not have the benefit of the SSM/I ice motion measurements, which, even with their large error variances, aid the deformation estimates in winter. Interestingly, in summer the correlation for the divergence was higher than that for the total deformation or shear.

In winter the mean total deformation measured by RGPS for a scale of 320 km was largest in the Beaufort Sea and smallest near the pole (Fig. 5). In summer there were broad regions of high deformation in the central Arctic Ocean and a

region of small deformation, perhaps fast ice, in the East Siberian Sea. Recall from Table 1 that mean total deformation was $0.91\% \text{ d}^{-1}$ in winter and $1.20\% \text{ d}^{-1}$ in summer. The mean total deformation for the MO and the DA runs was larger in summer than in winter as in the observations. The MO run resulted in much lower deformation than the DA run except in a few coastal regions. In both winter cases the deformation was largest near the coasts and smallest in the central region of the ocean. The deformation for the DA run was larger than in the RGPS measurements, particularly near the coasts (for example in the East Siberian Sea and off Point Barrow), and during summer.

Maps of the correlation between the observed total deformation and that in the MO and the DA runs are shown in Fig. 6. The correlation was very poor for the MO run at all locations and in both seasons, but quite good for the DA run, particularly during winter, far from the coasts, and where there were many buoys available for assimilation (Fig. 4). The maximum correlation was $R = 0.87$. This value is perhaps a good indicator of the best comparison possible between the two very different datasets even when the model is constrained to

Fig. 4 Number of daily buoy observations in each 40-km grid square during the 10-month study period. The concentration of observations in the Beaufort Sea is largely due to the buoys deployed in the vicinity of the drifting SHEBA ice camp.

follow the buoy-observed ice motion. The improvement that assimilation caused in the winter correlation is expected but in summer there was no such improvement over much of the region. Both had poor overall correlations, however, locally the DA simulation reached R values of 0.84 in the vicinity of the SHEBA camp where many buoys were available, similar to the maximum value in winter.

The winter deformation, both in the observations and in the model, was greater near the coasts than in the interior, while the correlations of both velocity and deformation were lower. This larger deformation is undoubtedly due to the constraints the land puts on the ice motion. Because coastal deformation was larger, open water formation, ridging, and ice production rates were also larger in these regions. However, these are just the regions where the correlation was greatly diminished.

Thomas (1999) found correlations between the deformation of 627 buoy clusters 400 to 600 km in diameter and a number of different model formulations over a 5-year period. The highest correlations were for the vorticity, and the

Fig. 5 Mean total deformation rate estimated by RGPS observations, the model only (MO), and the model with data assimilation (DA) at a spatial scale of 320 km and a temporal interval of 3 days.

Fig. 6 The correlation between the total deformation measured by RGPS and that estimated by the model only (MO) and with data assimilation (DA) at a spatial scale of 320 km and a temporal interval of 3 days. The colour scale represents the range $R = 0.0$ to 1.0 .

correlations for the shear were lower than for the divergence. He found the highest correlation for a simple best-fit linear model of the ice velocity and the geostrophic wind. This model had a vorticity correlation of 0.78. Model formulations very similar to those used here (but with a grid size of 160 km instead of 40 km) showed vorticity correlations of about 0.67. His results are not differentiated by season. The results from Table 2 show that the MO simulation had a vorticity correlation of 0.37 in winter and 0.58 in summer. These lower values may be related to the difficulty in matching the temporal sampling of the RGPS with that of the model or perhaps to the much shorter time period analysed. Figure 2 also suggests that the larger scale he used may increase the correlation.

c Time Series of Speed, Total Deformation, and Correlations

The temporal changes in the average speed and deformation are shown in Fig. 7 for the entire region. In the top plot, the mean ice speed in the DA model run matched that measured by the RGPS very well during most of the period, as expected. The MO model run showed a speed mostly lower than the observations until mid-May and matched the observations better during summer. The largest speeds were in November,

before the ice was well consolidated, and in mid-summer during free drift. The lowest speeds were in spring. The total deformation in both the models and the observations falls from a peak in November to lower values from January to April, and climbs again to a high in late summer. The DA model deformation tracked the observations better in winter than the MO results, but did not do as well in summer. The velocity correlation for the DA model run was very high in winter and decreased slightly in summer. As might be expected, the correlation of the deformation was loosely tied to the correlation in the ice velocity, with lower correlations in the deformation when the velocity correlations drop during summer.

d Deformation Distribution Functions

For many climate model applications it may be more important to estimate the distribution function of the deformation accurately than to obtain a good correlation with the observations. This is because the statistics of the deformation, the mean values and the shapes of the tails of the distribution function, determine ice growth rates in winter and open water formation in summer. Getting a good match between model results and observations for individual events may not be as important as obtaining a good estimate of the mean and of the

Fig. 7 Time series for the RGPS (black line) and the model (both model only, MO, blue line, and with data assimilation, DA, orange line) of the mean speed, the mean total deformation, the velocity correlations, and the total deformation correlations. The statistics are computed for 10-day windows every 5 days over the entire domain for the 320-km scale estimates.

variability. The RGPS data offer an opportunity to investigate the match in the distribution functions of the deformation between the model and observations. The distribution functions of the total deformation rate and of the divergence as measured by RGPS and as estimated by the DA model were compared for a scale of 80 km (Fig. 8). The smaller scale size was used because it was closer to the resolution of the model. At a scale of just 40 km, the deformation is poorly determined for a model with 40-km grid spacing, hence the choice of 80 km, which allowed the use of eight model grid points and roughly 64 RGPS cells to determine the deformation.

RGPS measurements in winter have a substantial percentage of very small deformation values (Fig. 8). We interpret the very low deformation values in the RGPS observations as

evidence of the occurrence of semi-rigid plates (Moritz and Ukita, 2000). The MO model run has much more of the ice exhibiting low deformation and the DA model has much less. The difference between the DA model and the measurements points to two problems with the model. One is that the DA model appears to smooth the velocity field too much so that narrow lines of deformation are smoothed to the extent that there is virtually no ice left in the rigid plates, plates that are often apparent in the RGPS data. A higher resolution version of the model, run at 10 km, showed very narrow slip lines and substantial regions of low deformation, so that the discrepancy in the distributions may be partly due to the model resolution. (A figure showing slip lines from the 10-km model can be found in Overland and Ukita, 2000.) The MO model run

Fig. 8 Comparisons of the distribution functions of the total deformation from RGPS and from the model only (MO) run and the model with data assimilation (DA) run for winter and summer for the entire area at a spatial scale of 80 km.

showed too much of the ice with very low deformation due to the low mean speeds of the MO run. The ice was moving too slowly and deforming too little. A second but related difficulty is the heavier tail of the DA model run. This discrepancy may be due to buoy motion that is quite different from that of the model solution and in the model a zone of ice is forced to move at a different velocity from the surrounding ice. This creates a zone of artificial deformation around the buoy. The large correlation length scale (700 km) used for assimilating the buoy velocities will tend to reduce this effect but in some conditions the creation of artificial deformation stemming from the conflict between isolated buoy data and the model field may be inevitable. The incorrect solutions may be the result of errors in the model physics, the model state (e.g., thickness), or the model forcing fields. The large difference in the sampling characteristics of the observations and the model results also must be considered as a possible source of the discrepancy in the distribution tails.

In summer the model-only run matched the observed distribution better than the data assimilation run. Under summer free-drift conditions, when the MO model was no longer locked up (see the speed distribution plots in Z2003) the

impositions of buoy motions appear to add to the deformation estimate of the DA model. The correlation of the total deformation also suffered (Fig. 7). Even in summer, some of the ice in the observations exhibited near-zero deformation while there was almost none in either model run.

4 Discussion and conclusions

These comparisons between model-based and observation-based deformation rates are an example of the use of RGPS data to determine which model formulations provide a better match with observations. Clearly the data assimilation run provided a better velocity correlation (as expected) in winter but surprisingly, in both winter and summer, data assimilation did not help with the mean deformation rate and indeed made the total deformation about 20% too large in winter and 40% too large in summer (Table 2). Much of the difference between the model only and the data assimilation runs can be traced to the fact that in winter the model-only run has the ice moving 20% too slowly resulting in lower mean total deformation rates.

The correlation between the model estimates and the RGPS measurements of the velocity was very good, particularly

with assimilation of buoy displacement data, throughout the period and in most of the region analysed, while the correlation in the deformation was much worse. How can velocity correlations be relatively high, but the deformation correlations be relatively low? The deformations are related to the spatial derivative of the velocity, and, although the large-scale velocity field is well represented in the model (as long as data assimilation is included), the smaller-scale spatial variations in velocity are not so well represented. Other important factors that can reduce the correlation of the deformation rates computed for the model and the measurements include errors in the model location of deformation events, and differences in the determination of the deformation arising from the alignment process. The RGPS deformation rate measurements may also be biased by excluding cells that have experienced very large deformation. Given the quite different spatial and temporal sampling of the model simulations and of the RGPS observations, it is perhaps remarkable that the deformation correlations were as high as we found. The high correlation of the velocity between the data assimilation run and the RGPS observations adds to the confidence that can be placed in the analysis and comparison techniques.

Although a high correlation in the deformation is desirable, for many climate applications it may be sufficient for ice production estimates if the distribution functions of the deformation rates, in particular the divergence, are well represented in the model. However our analysis indicates that there were important differences in the distribution functions for both the MO and the DA simulations.

The differences between the observed deformation rates and the DA model deformation rates were particularly large in the summer. The DA model appears to have larger deformation rates, in both the mean and in the standard deviation and in both shear and divergence. These comparisons indicate that simply assimilating limited quantities of high-quality ice motion data does not ensure accurate deformation rate estimates and can, in some situations, make the estimates worse. However, where there were many buoys to assimilate near the SHEBA experiment, the correlation between the observed and the DA model deformation was good, showing that with enough ice motion data the assimilation procedures can reproduce the observed deformation. More sophisticated data assimilation schemes than used here might reduce this problem.

These comparisons are most useful in showing aspects of the model that need improvement. What can these comparisons tell us about the model physics? Many factors influence the model estimates of the deformation rates: wind forcing fields, ocean currents, air and water drag formulations, the

ice-thickness distribution, ice redistribution formulations, and model constitutive laws relating deformation rates to stress. The ice thickness distribution is built up over time, so the thermodynamic formulations relating to ice growth and melt are also significant. As a consequence of the interdependence of many model units, one cannot change just one element of the model without seeing reflections in many aspects of the model response. One possibility is to focus on the aspect of the model most closely related to the deformation: the constitutive law that relates strain rates to stress. Different constitutive laws will produce different model deformation rates. Which law provides the best fit with the observations?

As an example of the way the RGPS data might be used to this end, we performed a model simulation with the ice strength reduced by 25%. In this simulation no data assimilation was included. The mean speed of the ice increased 28% and the mean total deformation rate increased 24% and both exceeded the RGPS-measured values. The velocity and deformation correlations did not change significantly. The ice thickness for the five-year period of 1993 to 1997, when compared to submarine-measured ice draft as in Z2003, showed no change in the bias and a small reduction in the correlation. The best model ice strength depends on many elements of the model, including the forcing wind field and the wind drag coefficients used. In this case, the deformation and velocity comparisons suggest the best ice strength value to be about 20% smaller than in our standard simulation.

The RGPS dataset provides highly accurate deformation rate measurements that can provide a useful comparison dataset for sea-ice models when the two datasets are properly interpolated and aligned. We have not exhausted the possibilities of using the data to improve our sea-ice model, but have pointed out a direction that we or others can follow. The deformation of the ice plays an essential role in the evolution of the ice cover and models will serve us best if they can accurately reproduce the observed deformation.

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