SENSITIVITY OF MODELED SEA ICE TO EXTERNAL FORCING AND PARAMETERIZATIONS OF HEAT EXCHANGE PROCESSES

A.P. Makshtas¹, S.V. Shoutilin¹,² and V.F. Romanov¹,²

ABSTRACT
A dynamic–thermodynamic sea ice model with 50-km spatial and 24-hour temporal resolution and zero-dimensional thermodynamic sea ice models are used to investigate the spatial and temporal variability of the sea ice cover and the surface energy exchange in the Arctic Basin. The models satisfactorily reproduce the averaged main characteristics of the sea ice and the sea ice extent in the Arctic Basin. At times the evaluation of atmospheric forcing data from NCEP, generally accepted for sea ice climate models, shows large disagreement with data from meteorological observations from drifting stations. The numerical experiments with zero-dimensional sea ice model reveal that negative feedback existing in nature and reproduced in the models artificially reduce the influence of inaccuracy of forcing parameters on the results of modeling sea ice thickness, but distort information about main surface heat fluxes. The possible approach to improve NCEP atmospheric surface data, based on the Rossby-number similarity theory and universal dimensionless functions is outlined.

INTRODUCTION
Recent reports of the decrease in sea ice extent (e.g. Parkinson et al., 1999) and thickness (Rothrock et al., 1999) in the Arctic Basin stimulate attempts to explain this remarkable change in the main features of the Arctic climate system. One of the main ways to understand the reasons for climate changes is to use numerical modeling to reproduce observed environmental characteristics and to study the possible mechanisms responsible for such changes. Explanation of these changes, however, depends crucially on the model used for investigating long-term sea ice changes in the Arctic and the external forcing applied in the model. Many authors (e.g. Maslanik et al., 2000) explain the decrease in sea ice cover in the eastern part of the Arctic Basin during last decades as a consequence of increased open water and thin ice, forced primarily by ice dynamics. As a result, the surface albedo decreases; the absorbed solar radiation at the surface and in the oceanic mixed layer likewise increases; and, finally, the lateral and bottom melting increases. Hilmer and Lemke (2000), on the other hand, conclude from

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sensitivity simulations that most of the thinning can be attributed to changes in the surface-level air temperature rather than to changes in the circulation.

Using Thorndike’s simple model, Rothrock et al. (1999) discuss possible thermodynamic processes that could produce the observed thinning, such as an increase in the oceanic heat flux, an increase in the poleward atmospheric heat transport and the consequent increase in the incoming longwave radiation, or an increase in the downwelling shortwave radiation. Other proposed reasons for the decrease in sea ice thickness and extent are changes in snow precipitation and snow depth or an increase in ice export from the Arctic Basin.

Unfortunately, existing data sets from different reanalyzes, which are needed as forcing for climatic sea ice models, continue to have uncertainties, particularly in the surface fields and fluxes, as described by the World Meteorological Organization (1999). The same problems restrict the use of coupled models for studying climate anomalies in the Arctic sea ice (Maslanik et al., 2000). Because of these shortcomings in the various data sets, almost all of the sea ice models use the NCEP/NCAR reanalysis data (Kalnay et al., 1996) entirely or use only the near-surface air temperature and surface-level pressure together with different climatic mean values of other atmospheric forcing parameters. The restrictions of such approaches for projecting the long-term variability of the sea ice cover will be shown below. We will analyze results from numerical experiments with a dynamic-thermodynamic and a zero-dimensional thermodynamic sea ice models, forced with atmospheric data from NCEP (1948–1997) and with data of field observations executed on the Russian and US drifting stations (1954–1990), to study the sensitivity of modeled sea ice and air-sea ice interaction processes to external forcing and some parameterizations of heat fluxes used in sea ice models.

DATA
To investigate how NCEP reproduces recent climate in the Central Arctic Basin we put together daily averaged data of standard meteorological observations executed on drifting stations, further named as NP data, and NCEP data, interpolated to the points where drifting station sailed each particular day. Comparison had been carried out for January and July, typical winter and summer months.

Figure 1: Comparison of the main meteorological parameters of atmospheric surface layer in the Canadian Arctic, measured on drifting stations and interpolated from NCEP data for January (1,3) and July (2,4).

Figure 1 and Table 1 illustrate the results of comparison for atmospheric surface pressure (P), air surface layer temperature (T) and wind velocity (U), and total cloudiness (N). The statistical characteristics of differences between NP and NCEP data demonstrate the best agreement between measured and calculated atmospheric surface pressure, and calculated on the basis of NCEP pressure fields surface wind velocity during both seasons. The reproduction by NCEP of the spatial-temporal variability of T
is much worse, especially in summer, when correlation falls to 0.56 and positive bias increases up to 0.7 °C. Additionally, Figures 1.1, 1.2 show the positive trends of difference between measured and calculated T for both seasons, reached 0.2 °C per decade. The last supports the Kalnay et al. (1996) remark about the incorrectness to use NCEP data for calculations of long-term trends, caused by different reanalysis routine used before and after 1978.

Table 1: Mean values and mean square deviations (msd) of differences between NCEP and NP data and correlation coefficients between time series of meteorological parameters.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>January</th>
<th>July</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mean</td>
<td>msd</td>
</tr>
<tr>
<td>DP, mb</td>
<td>–1.4</td>
<td>2.1</td>
</tr>
<tr>
<td>DU, m/s</td>
<td>–0.3</td>
<td>1.5</td>
</tr>
<tr>
<td>DT, °C</td>
<td>0.2</td>
<td>3.5</td>
</tr>
<tr>
<td>DN, tenth</td>
<td>–0.3</td>
<td>3.1</td>
</tr>
</tbody>
</table>

NCEP reproduction of cloudiness in the Arctic Basin is quite inadequate. Correlation between measured and calculated N does not exceed 0.5 and follow Figures 1.3 and 1.4 there are the principal distinctions in the shapes of N frequency distributions for both seasons. Additional information about problems related to cloudiness in the Arctic basin can be found in paper of Makshtas et al. (1999). In the next parts we will show the response of modeled sea ice to based on NCEP data inadequate atmospheric forcing.

MODEL

The sea ice cover in our large-scale dynamic-thermodynamic model (Makshtas et al., 2001) is described at each grid point by the relative areas of level or undeformed ice (thickness \( h_i \), area \( N_i \)), which undergoes thermodynamic growth and melting; ridged ice (area \( N_h \)) with fixed effective thickness \( h_h = 12 \) m, and leads (area \( N_0 \)). The main equations of the model are the momentum equation, which includes a parameterization of internal ice stresses in the framework of a cavitating fluid (Flato and Hibler, 1992); the quasi-steady heat conduction equation, which describes heat processes in the level sea ice and its growth or melting; and the nonstationary mass balance equation. This last equation is

\[
\frac{\partial m_i}{\partial t} + \text{div}(\bar{u}m_i) + f = 0, \tag{1}
\]

where \( m_i = h_iN_i + h_hN_h \) and \( \bar{u} \) is ice drift velocity. Finally, here,

\[
f = N_i \left( \frac{\partial h_i}{\partial t} \right)_T + h_i \left( \frac{\partial N_i}{\partial t} \right)_T + h_h \left( \frac{\partial N_h}{\partial t} \right)_T + N_h \left( \frac{\partial h_h}{\partial t} \right)_T \tag{2}
\]

is a function describing the thermodynamic growth or melting of level ice, the lateral melting of level and ridged ice in leads, and melting at the upper and lower surfaces of ridges, recalculated in turn to reduction of area occupied by ridged ice and increase the area occupied by level ice.

Our model describes the growth and melting of level ice with a zero-dimensional thermodynamic sea ice model, similar to the Semtner (1976) model. We described the energy fluxes between the atmospheric surface layer and sea ice surface following
Jordan et al. (1999) and used the following values of surface albedo: 0.81 for dry snow, 0.70 for melting snow, and 0.63 for melting bare ice. We model heat processes in leads following Ebert and Curry (1993). To calculate the redistribution of lateral heat fluxes between ridged and level ice, we use an algorithm proposed by Doronin, (1969). We describe the bottom and surface melting of ridges following Thorndike et al. (1975, their Table 1). The redistribution of the areas occupied by leads, level ice, and ridges due to thermodynamic and dynamic processes is calculated with an algorithm based on the method of large particles (Belotserkovskii, 1984).

The model is driven using daily 2-meter air temperature and relative humidity, atmospheric surface pressure, total cloudiness amount, monthly mean solid precipitation and dynamic heights of the ocean surface. The model has a horizontal resolution of 50 km by 50 km and a temporal resolution of 24 hours. With the model, we can calculate the temporal-spatial distribution of the sea ice thickness and area, the snow depth, the turbulent sensible (H) and latent (LE) heat fluxes, longwave (R) and shortwave (F) radiation balances at the upper surface for each kind of ice cover, the temperature in leads and in the upper ocean, and the heat flux from the ocean mixed layer to the bottom of the sea ice.

**THERMODYNAMIC SEA ICE MODEL AND EXTERNAL FORCING**

To investigate sensitivity of modeled sea ice to external forcing we used thermodynamic part of our dynamic-thermodynamic model. The model was integrated for 50 years with NCEP data about T and U for the North Pole with the same atmospheric forcing for 1958 and an oceanic heat flux of 1 W/m². The results of the standard run (Figure 2, Tables 2 and 3) show agreement with the results of Maykut and Untersteiner (1971) for equilibrium sea ice thickness. Our numerical experiments show that the equilibrium sea ice thickness is much more sensitive to changes in summer air temperature than in winter air temperature. Figure 2a demonstrates that restricting the summer air temperature from the NCEP data to be 0 °C or less (note, that it is corresponded to differences between NCEP and NP) and –0.5 °C or less leads to increases in equilibrium thickness of 1.5 and 3 times, respectively. In contrast, changing the surface-level air temperature in winter by +10 °C or –10 °C changes the equilibrium sea ice thickness by only 30 % (Figure 2b).

![Figure 2: (a) Seasonal variability of equilibrium sea ice thickness near the North Pole, calculated with the thermodynamic sea ice model and the NCEP data for surface-level air temperature $T_A$ (diamonds) and also when the air temperature is restricted to be $T_A = 0$ °C if $T_A > 0$ °C (squares) and $T_A = -0.5$ °C if $T_A > -0.5$ °C (circles). (b) Same as for (a) except during the winter the surface-level air temperature differs from the NCEP data by 10 °C (circles), 2.5 °C (squares), –2.5 °C (diamonds) and –10 °C (triangles).](image-url)
In winter there is a strong negative feedback between the turbulent sensible heat flux (H) and the longwave radiation balance (R). As a result, the conductive heat flux in the sea ice is not very sensitive to air temperature changes; and, in turn, the growth of ice on the bottom of the ice cover is not either. In summer, on the other hand, the surface temperature of the ice is fixed at the freezing point, and the negative feedback between the turbulent and longwave radiative heat fluxes cannot operate. As a result, any increase in the downward sensible heat flux caused by a small increase in the surface-layer air temperature leads directly to increased melting of the ice cover. By the way, the same effect explains the relatively high sensitivity of the modeled ice cover to summer clouds when averaged or daily cloudiness data are used in simulations (Makshtas et al., 1999). In this case the increase or decrease of cloudiness and related changes in longwave radiative heat flux are not compensated by changes in turbulent heat fluxes.

One weakness of the above results is the calculated equilibrium sea ice thickness is a cumulative product of the energy exchange between the sea ice and the atmosphere during 50 years of integration with a fixed annual cycle of external forcing, while the corresponding heat fluxes, in turn, depend on the changes in ice thickness. We, therefore, made some numerical experiments with a fixed ice thickness of 3 m to avoid this effect. At each time step during forty years of integration with NCEP forcing, we calculated ice surface and bottom growth or melting, and the corresponding heat fluxes. But after each step, we returned the snow and sea ice thickness to the value 0.2 and 3 meters. The results of these numerical experiments are shown in Tables 2, 3. Table 2 shows a very modest impact of SAT on modeled sea ice cover in winter. Changes in the SAT of 10 °C above and below the baseline NCEP data during December–February decrease or increase the ice growth for the period by only about 0.1 m. The strong insulating properties of snow-covered sea ice and the above-mentioned negative feedback between the turbulent and radiative fluxes explain the small changes in ice growth.

Table 2: Modeled average surface heat fluxes (W/m²) and corresponding growth rates (m/period) during winter (December–February) at the North Pole with the sea ice thickness fixed at 3 m.

<table>
<thead>
<tr>
<th>Tₐ +10 °C</th>
<th>Tₐ +5 °C</th>
<th>Tₐ +2.5 °C</th>
<th>Tₐ</th>
<th>Tₐ –2.5 °C</th>
<th>Tₐ –5.0 °C</th>
<th>Tₐ –10 °C</th>
</tr>
</thead>
<tbody>
<tr>
<td>H</td>
<td>–13.3</td>
<td>–12.9</td>
<td>–12.6</td>
<td>–11.6</td>
<td>–11.2</td>
<td>–10.1</td>
</tr>
<tr>
<td>LE</td>
<td>1.4</td>
<td>1.2</td>
<td>1.0</td>
<td>0.9</td>
<td>0.8</td>
<td>0.7</td>
</tr>
<tr>
<td>R</td>
<td>25.0</td>
<td>26.4</td>
<td>27.1</td>
<td>27.5</td>
<td>27.9</td>
<td>27.7</td>
</tr>
<tr>
<td>Dh</td>
<td>0.30</td>
<td>0.36</td>
<td>0.38</td>
<td>0.40</td>
<td>0.43</td>
<td>0.46</td>
</tr>
</tbody>
</table>

In contrast, Table 3 shows that, during the melting period, the increase in averaged turbulent latent heat flux and the change in direction of the turbulent sensible heat flux between experiments when Tₐ ≤ 0.5 °C and Tₐ ≤ –0.5 °C increase the cooling of the upper ice surface and, therefore, decrease the melting rate. It is interesting that the increased cooling due to the turbulent fluxes is almost equal to the decrease in the absorbed solar radiation (F) during the same period caused by albedo differences related to the shortening of the melting period from 78 to 48 days. Table 3 shows that these mean differences are 7.3 W/m² for H, 7.1 W/m² for LE, 1.4 W/m² for R, and 8.3 W/m² for F. We need to point out also the asymmetry in surface melting rates between the
experiments with $T_a \leq 0.5 \, ^\circ C$ and $T_a = 0.0 \, ^\circ C$ and with $T_a = 0.0 \, ^\circ C$ and $T_a \leq -0.5 \, ^\circ C$: 0.17 m/period and 0.27 m/period, respectively. In summary, we find that a decrease in SAT below the melting point in summer has a much stronger impact on sea ice cover than a comparable increase.

Table 3: Modeled average surface heat fluxes (W/m²), corresponding melting rates (m/period), and duration of bare ice melting (days) during summer at the North Pole with the sea ice thickness fixed at 3 m.

<table>
<thead>
<tr>
<th>$T_a \leq 1.0 , ^\circ C$</th>
<th>$T_a \leq 0.5 , ^\circ C$</th>
<th>$T_a \leq 0.0 , ^\circ C$</th>
<th>$T_a \leq -0.5 , ^\circ C$</th>
<th>$T_a \leq -1.0 , ^\circ C$</th>
</tr>
</thead>
<tbody>
<tr>
<td>H</td>
<td>–1.8</td>
<td>–1.5</td>
<td>0.3</td>
<td>3.0</td>
</tr>
<tr>
<td>LE</td>
<td>6.4</td>
<td>6.8</td>
<td>8.6</td>
<td>10.9</td>
</tr>
<tr>
<td>R</td>
<td>22.4</td>
<td>22.4</td>
<td>22.8</td>
<td>23.2</td>
</tr>
<tr>
<td>F</td>
<td>–58.7</td>
<td>–58.7</td>
<td>–58.3</td>
<td>–53.2</td>
</tr>
<tr>
<td>dh total</td>
<td>–0.77</td>
<td>–0.74</td>
<td>–0.56</td>
<td>–0.29</td>
</tr>
<tr>
<td>Duration of melting</td>
<td>78</td>
<td>78</td>
<td>76</td>
<td>48</td>
</tr>
</tbody>
</table>

We applied the same approach to investigate interannual variability of the thermodynamic sea ice model response for January and July to different atmospheric forcing from the data sets described in part 2.

Figure 3: Monthly mean growth rate of modeled sea ice in January (a) and melting rate in July (b) under NCEP (solid line), NP (dash line), and NCEP-NP (points) forcing.

The main outputs of the model, forced by daily averaged NCEP and NP data (Figure 3, Table 4), confirm the above conclusions. In January negative feedback between H and R (correlation –0.93 and –0.83 for NCEP and NP respectively) stipulated a good agreement between monthly mean sea ice growth rate (Fig. 3a) and relatively small difference between mean values of heat fluxes (Table 4). In summer the difference between results is much larger in the mean melting rates and its trends (Fig. 3b) as well as in the values of heat fluxes. The absence of mentioned feedback (correlation between H and R equal 0.19 for NCEP and –0.37 for NP) due to fixed surface temperature near melt point and positive bias NCEP air surface level temperature in summer are the main reasons of such difference. Additionally wrong description of cloudiness (Part 2) in NCEP determine large values of longwave and shortwave radiation fluxes, having opposite directions. Ironically the combination of NCEP data about air surface level temperature and wind velocity with relatively correct data about cloudiness and relative humidity from different climatologies, usually used in climate sea ice model, leads to
even more worse results in calculations of melting rate than NCEP data only, due to different influence of cloudiness on R and F values.

Additionally to comparison the turbulent heat fluxes and longwave radiation balance parameterizations, described by Jordan et al. (1999) and Makshtas et al. (1999), our numerical experiments (Table 4) shows that employment of popular Shine (1984) and Marshunova (in Doronin, 1969) parameterizations for calculations shortwave radiation balance do not disturb strong the results of sea ice modeling.

Table 4: Average for 1954–1990 surface heat fluxes in July and January, its msd (W/m²), and corresponding melting rates (m/day) calculated for sea ice with fixed 3 m thickness.

<table>
<thead>
<tr>
<th>Data Set</th>
<th>H</th>
<th>Msd</th>
<th>LE</th>
<th>Msd</th>
<th>R</th>
<th>msd</th>
<th>F</th>
<th>Msd</th>
<th>Dh</th>
</tr>
</thead>
<tbody>
<tr>
<td>NCEP</td>
<td>−7.1</td>
<td>1.9</td>
<td>−4.8</td>
<td>1.4</td>
<td>35.3</td>
<td>3.0</td>
<td>129</td>
<td>12.4</td>
<td>−0.037</td>
</tr>
<tr>
<td>NP (S)</td>
<td>1.1</td>
<td>5.0</td>
<td>8.7</td>
<td>5.7</td>
<td>16.9</td>
<td>3.3</td>
<td>123</td>
<td>12.2</td>
<td>−0.033</td>
</tr>
<tr>
<td>NCEP-NP</td>
<td>−7.2</td>
<td>1.8</td>
<td>0.5</td>
<td>2.9</td>
<td>15.9</td>
<td>3.2</td>
<td>123</td>
<td>12.3</td>
<td>−0.039</td>
</tr>
<tr>
<td>NP (M)</td>
<td>1.0</td>
<td>5.0</td>
<td>8.6</td>
<td>5.6</td>
<td>16.9</td>
<td>3.3</td>
<td>118</td>
<td>11.6</td>
<td>−0.031</td>
</tr>
<tr>
<td>NCEP (w)</td>
<td>−7.1</td>
<td>2.7</td>
<td>2.2</td>
<td>1.3</td>
<td>26.4</td>
<td>2.4</td>
<td>0</td>
<td>0</td>
<td>0.005</td>
</tr>
<tr>
<td>NP(w)</td>
<td>−9.8</td>
<td>4.2</td>
<td>1.4</td>
<td>0.7</td>
<td>25.7</td>
<td>4.0</td>
<td>0</td>
<td>0</td>
<td>0.005</td>
</tr>
</tbody>
</table>

Note: letters S and M in first column indicate calculation of shortwave radiation balance by Shine (1984) and Marshunova (in Doronin, 1969) parameterizations respectively; letter “w” indicate model results for January.

DISCUSSION
In previous parts we showed that an inaccuracy of NCEP reanalysis forcing leads to rather large differences in values of main parameters of modeled sea ice and, or magnitudes of corresponding heat fluxes. Same time the relatively correct NP data, used for comparison, covered 1954–1990, and characterized very restricted area of the Arctic Basin, could be used only for validation of different routines, employed for calculations the main parameters of polar atmosphere, used as a forcing for large scale climate models of the Arctic ocean. From point of view the temporal-spatial scales and resolution NCEP-reanalysis is the best among others archives. Fortunately, the numerical experiments with dynamic-thermodynamic sea ice model show much more weaker sensitivity of modeled sea ice cover to atmospheric forcing in comparison with pure thermodynamic models. Even numerical experiments with increase or decrease of air surface level temperature in summer, crucial for thermodynamic model (Fig. 2, Table 3), leads to very modest changes of modeled mean sea ice thickness in different parts of the Arctic basin (Fig. 4). The complicate interaction of level ice with leads and ridges (two other main features of sea ice cover) due to numerous dynamic and thermodynamic processes, which analysis is beyond our paper due to restricted volume, determines relative stability of modeled ice cover.
The errors in computing surface heat fluxes over sea ice, related with inadequate presentation of the polar atmospheric boundary layer (ABL), could be crucial for projection of climate change with coupled atmosphere-sea ice-ocean models. Preliminary numerical experiments with new ABL parameterization including NCEP data from 850 and 925 hPa, accounting horizontal nonhomogeneity of underlying surface and advection in the low atmosphere, allow to hope on significant improvement of interactive ABL-sea ice-ocean models.

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REFERENCES


