Arctic decadal variability from an idealized atmosphere-ice-ocean model:

1. Introduction

1.1. Problem Formulation

[2] Summarizing results from different studies of Arctic climate variability, Proshutinsky et al. [2002] formulated a conceptual model where freshwater and heat fluxes between the Arctic Ocean and GIN Sea play a crucial role in generating self-oscillating climate regimes with a period of 10–15 years. There are several ways to test this hypothesis. One of the most convincing approaches may be to analyze observations, but there are not enough data, and analysis of observational data would not allow us to investigate the mechanisms and roles of different factors that influence system behavior. Another approach is numerical modeling and using results from model runs. Climate models of different complexity (from very simple box models to extremely complex coupled global climate models) could be employed [North, 1975; Saltzman, 1985; Henderson-Sellers and McGuffie, 1987; Häkkinen, 2000; Curry and Lynch, 2002]. State-of-the-art coupled general circulation models of the atmosphere and the ocean have been developed and used to study decadal variability of Arctic climate, and its interaction with the North Atlantic Ocean, as well as the interaction between sea ice and global thermohaline circulation [Manabe and Stouffer, 1999; Gordon and O’Farrell, 1997; Flato et al., 2000; Gordon et al., 2000]. These types are considered to be “the most complete type of climate models currently available” [Henderson-Sellers and McGuffie, 1987]. However, in terms of delicate balances between the Arctic climate components and their poorly documented sensitivities, it is not surprising that climate models have not succeeded in explaining the observed changes in the Arctic climate system. Comparison with observations shows that “models have difficulty reproducing both the trends and the mean state of the Arctic, raising doubt over their ability to elucidate causative connections or predict future trends” [Shindell, 2003]. Claussen et al. [2002] warns against overestimating the reliability of the “state-of-the-art” models and argues that the description of the climate system should rely on a spectrum of climate system models, including simple conceptual models.

[3] The advantage of idealized climate models is their relatively simple formulation where only the major forces are included. Simple climate models have low computational cost and are designed to investigate the basic relationships between components of the climate system, the factors driving climate change and the overall response of the system [Claussen et al., 2002].

[4] There are also models of intermediate complexity that still have enough simplicity to be used to understand the role of different forcing factors. Applications of this type of model for studies of Arctic climate variability [e.g., Goosse et al., 2003; Mysak et al., 2005] have shown promising results. Nevertheless, it is still difficult to determine where and how internal mechanisms work in the coupled atmosphere-ice-ocean system, so model output needs to undergo

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complex data analysis to reduce uncertainty and find the major relationships and causes of climate variability.

[5] A simple model reproducing the major links and interactions among components of the Arctic climate system similar to Proshutinsky et al. [2002] must be able to simulate the mechanisms of decadal variability. Obviously such a model should include several simple, coupled models describing the essential processes. This could be done by employing an existing “multibox” climate model, however, one disadvantage of such models (discussed in the next section) is that, being constructed to solve a particular problem, they may be inadequate for our purposes. To meet our needs, we develop, validate, and calibrate a simple model of an idealized atmosphere-ice-ocean climate system.

1.2. Overview of Simple Models of the Arctic

[6] Relatively small horizontal gradients of temperature and salinity within the real Arctic Ocean enabled researchers to apply simple one-dimensional and box models to generate a realistic first-order description of the stratification and evolution of the upper Arctic Ocean [Stigebrandt, 1981; Killworth and Smith, 1984; Lemke and Manley, 1984; Björk, 1989]. For the Greenland-Iceland-Norwegian Seas (GIN Sea), simple models are mostly employed to simulate convection in the Greenland Sea [Killworth, 1979; Houssais, 1987; Visbeck et al., 1995]. Table 1 lists a number of simple models of the Arctic climate study and their characteristics.

[7] The first simple dynamical models of the Arctic Ocean appeared in the early 1980s, and the simplest were characterized by few interacting components. Most of the simple models of the Arctic Ocean simulated processes in the upper ocean (above 200 m) [Björk, 1989; Lemke and Manley, 1984] and consisted of two layers: a homogeneous mixed layer and a pycnocline. Some models included the deep ocean [Killworth and Smith, 1984; Alekseev and Ryabchenco, 1996]. Deepening of the upper layer in the mixed layer models was based on Kato-Phillips [Kato and Phillips, 1969] and the Niiler and Kraus formulation [Niiler and Kraus, 1977] which parameterizes the rate of entrainment of fluid across the lower interface of the mixed layer, based on the balance of potential and kinetic energies in the layer.

[8] Entrainment depends on the density jump at the base of the mixed layer and the thermohaline structure of the underlying pycnocline. In some simple ocean models, the pycnocline was treated as a homogeneous layer [Walin, 1993; Robitaille et al., 1995]. In other mixed layer models, the pycnocline structure was approximated by an analytical function, like the exponential function in Lemke and Manley [1984] or the parametric fitting of Miroslowsky et al. [1969] used in Alekseev and Ryabchenco [1996]. In the model of Björk [1989], the thermohaline structure in the pycnocline was determined by vertical advection-diffusion equations for temperature and salinity.

[9] Following the classification of Claussen et al. [2002], the most advanced among simple models are simple “climate system models” which are applied to study long-term climate variability. The simple climate system models are constructed by coupling several box and one-dimensional models that simulate component processes in the climate system. These models have more interacting components and a higher value of cumulative dimension.

[10] Several simple climate system models have been developed for the high latitude environment [e.g., Robitaille et al., 1995; Wang and Mysak, 2000]. Among simple climate models of the Arctic Ocean – Greenland Sea region, the one by Robitaille et al. [1995] would seem appropriate for our problem. However, this model was not designed to reproduce oscillations. Also the formulation of the model contradicts our view of the problem: in Robitaille et al. [1995], the atmosphere completely controlled the system. In our mechanism the ocean affects the atmosphere through the heat fluxes from the GIN Sea. Also, the thermohaline structure of the Arctic Ocean cannot be resolved in a two-layer model such as Robitaille et al.’s.

[11] From the above it follows that in order to simulate our climate system there is no “ready-to-use” simple model. This caused us to develop our own simple model of the Arctic Ocean - Greenland Sea atmosphere-ice-ocean climate system. The description of the model components is presented in section 2. Validation of the model is given in section 3.

2. Model Description

[12] The Arctic system model consists of two modules: the Arctic module and the Greenland Sea module (Figure 1). The Arctic module includes a shelf box model coupled with a thermodynamic sea ice model and an Arctic Ocean model coupled with a sea ice model and an atmospheric box model.

[13] In the GIN Sea we focus on ocean overturning, air-sea flux, ice formation, and surface air temperature (SAT) anomalies in the convection region (Greenland Gyre). The Greenland Sea module includes an oceanic model of the central Greenland Sea region coupled with a sea ice model and a statistical model of SAT in the Greenland Sea. The module interaction between the Arctic Ocean and the Greenland Sea is realized through the freshwater flux (ice and water) from the Arctic Ocean, and atmospheric heat advection from the Greenland Sea. The ice-ocean components interact with the atmospheric models via the net surface heat flux.

[14] The model is forced by solar (shortwave downwelling) radiation, wind stress, river runoff, Atlantic inflow, Pacific Water inflow through Bering Strait, cloudiness, air humidity, and ice/snow albedo. The daily shortwave downwelling radiation values have been computed for 75°N and 85°N latitudes using equations of the Sun-Earth astronomical relationships [Iqbal, 1983]. Other forcing parameters have been prescribed with monthly means linearly interpolated into daily data [Gorshkov, 1980; Lindsay, 1998; Polyakov et al., 1999]. The Notation section lists most notations used in the model.

2.1. Arctic Module: Arctic Ocean Model

[15] The Arctic Ocean model (Figure 2) describes seasonal and interannual variability in the upper ocean. The model is based on the pycnocline model of Stigebrandt [1985] modified and applied for the Arctic Ocean by Björk [1989]. In this study, the model has been further modified by new algorithms for: parameterization of forced convec-
Table 1. Simple Models of the Arctic and GIN Sea Region

<table>
<thead>
<tr>
<th>Year</th>
<th>Model Description</th>
<th>Simulated Regions</th>
<th>Research Goals</th>
<th>Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>1979</td>
<td>Quasi-static and time dependent models of vertical convection based on energy, salinity and temperature budgets of mixed layer and sea ice [Killworth, 1979]</td>
<td>Weddel Gyre, Greenland Sea Gyre</td>
<td>Study the development of chimney in the convection regions</td>
<td>Mixed layer thickness, ice production</td>
</tr>
<tr>
<td>1981</td>
<td>Two-layer (mixed layer and halocline) ocean box model [Stigebrandt, 1981]</td>
<td>Arctic Ocean</td>
<td>Study sensitivity of ice cover to the freshwater supply</td>
<td>Constant wind, water and ice transport at the lateral boundaries</td>
</tr>
<tr>
<td>1984</td>
<td>1-D halocline model [Killworth and Smith, 1984]</td>
<td>Arctic Ocean and shelf</td>
<td>Study vertical thermohaline structure</td>
<td>Constant river runoff, Bering Strait inflow, shelf water outflow, and Atlantic water inflow, surface salt flux</td>
</tr>
<tr>
<td>1984</td>
<td>1-D upper ocean model (mixed layer and halocline) [Lemke and Manley, 1984]</td>
<td>Arctic Ocean</td>
<td>Simulate seasonal variability of the upper Arctic Ocean</td>
<td>Monthly wind stress and surface salt fluxes</td>
</tr>
<tr>
<td>1984</td>
<td>1-D mixed layer model coupled to a thermodynamic sea ice model [Lemke, 1987]</td>
<td>Greenland Sea</td>
<td>Investigate the seasonal cycle of ice-ocean interaction in the subarctic ocean</td>
<td>Monthly surface heat fluxes, evaporation, precipitation, wind, and surface air temperature</td>
</tr>
<tr>
<td>1984</td>
<td>1-D pycnocline model (mixed layer and pycnocline) [Björk, 1989]</td>
<td>Interior Arctic Ocean and shelf</td>
<td>Simulate the upper ocean thermohaline structure</td>
<td>Monthly wind, cloudiness, precipitation, SAT</td>
</tr>
<tr>
<td>1987</td>
<td>1-D ocean mixed layer model coupled to a thermodynamic sea ice model [Houssais, 1987]</td>
<td>Arctic Ocean, Southern Ocean</td>
<td>Study the effect of variations of the entrained oceanic heat flux on the sea ice cover</td>
<td>Monthly wind, river runoff, ice production and export, Bering Strait inflow</td>
</tr>
<tr>
<td>1989</td>
<td>1-D upper ocean model (mixed layer and halocline) ocean model coupled to an idealized ice model [Walin, 1993]</td>
<td>Greenland Sea</td>
<td>Study the development of deep convection in the ice covered ocean</td>
<td>Heat flux to the atmosphere, precipitation, oceanic heat flux to the ice,</td>
</tr>
<tr>
<td>1990</td>
<td>1-D mixed layer model coupled to a three-layer snow-ice model [Häkkinen and Mellor, 1990]</td>
<td>Arctic Ocean</td>
<td>Simulate Arctic ice mass and mixed layer properties</td>
<td>SAT and wind</td>
</tr>
<tr>
<td>1993</td>
<td>1-D 2 layer (homogeneous mixed and deep layers) ocean model coupled to an idealized ice model [Walin, 1993]</td>
<td>Greenland Sea</td>
<td>Study the evolution of the mixed layer in the convection region</td>
<td>Monthly mean SAT, river runoff, constant inflows at the lateral boundaries</td>
</tr>
<tr>
<td>1995</td>
<td>1-D model of the mixed layer with entrainment parameterized using critical Richardson numbers [Pawlowicz, 1995]</td>
<td>Greenland Sea Gyre</td>
<td>Study interaction between the Arctic and the GIN Sea and its effect on Arctic climate variability</td>
<td>Constant heat, freshwater, wind stress fluxes at the ocean surface</td>
</tr>
<tr>
<td>1995</td>
<td>Simple climate system model consisting of the upper ocean box model and thermodynamic ice model [Robitaille et al., 1995]</td>
<td>Arctic Ocean, Greenland Sea, Norwegian Sea</td>
<td>Investigate the stratification in the Greenland Sea in winter</td>
<td>Monthly incoming shortwave radiation, surface air temperature, wind stress, cloudiness, air humidity, volume fluxes of the Bering Strait inflow, Atlantic water, and river runoff</td>
</tr>
<tr>
<td>1995</td>
<td>1-D mixed layer model coupled to a thermodynamic ice model [Visbeck et al., 1995]</td>
<td>Greenland Sea Gyre</td>
<td>Study seasonal variability in the ocean-ice system</td>
<td>Seasonally varying inflows at Bering Strait and constant inflow from the Kara Sea, monthly river runoff, net precipitation, constant wind stress</td>
</tr>
<tr>
<td>1996</td>
<td>1-D 3-layer (upper mixed layer, the halocline layer, and the deep layer) ocean model coupled with a thermodynamic sea ice model [Alekseev and Ryabchenko, 1996]</td>
<td>Arctic Ocean</td>
<td>Examine the freshwater balance of the Arctic Ocean</td>
<td>Constant air humidity, air temperature, wind speed</td>
</tr>
<tr>
<td>1996</td>
<td>Multibox model of the upper Arctic Ocean coupled to an idealized ice model consisting of ridging and thermodynamics (look-up Tables) and advection [Steele et al., 1996]</td>
<td>Arctic Ocean</td>
<td>Study the relation between evolution of ice cover, heat loss to the atmosphere and ice melt, entrained heat and development of deep convection</td>
<td>Solar radiation, zonally averaged monthly mean SST, wind stress</td>
</tr>
<tr>
<td>1999</td>
<td>1-D pycnocline model coupled to a model of ice volume [Rudels et al., 1999]</td>
<td>Weddell Sea, Greenland Sea</td>
<td></td>
<td></td>
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</tbody>
</table>
Figure 1. Diagram of an idealized Arctic system model. \( Q_f \) - river runoff; \( Q_{(P,E)} \) - net precipitation (precipitation minus evaporation); \( Q_{m_{ao}} \) - flow from the mixed layer to the shelf; \( Q_{sh} \) - shelf water flow to the Arctic Ocean; \( Q_{ice} \) - ice flux to the GIN Sea; \( Q_{PW} \) - Polar water flux; \( F_{adv} \) - sensible heat flux; \( F_{tot} \) - surface heat flux.

2.1.1. Mixed Layer

[16] The Arctic Ocean model is one dimensional. It consists of three interacting layers (Figure 2): a surface mixed layer, an intermediate halocline layer with depth dependent temperature and salinity, and a deep Atlantic layer.

\[ \frac{dh_{ml}}{dt} = \frac{1}{A_{AO}} \left( Q_{sh_{-ml}} - Q_{m_{ao}} + Q_{gML_{-atl}} + \rho_{Ber} \cdot Q_{Ber} + \varepsilon \cdot Q_{ice} \right) - \omega_e, \]  

where \( A_{AO} \) is area of the deep Arctic Ocean \((0.61 \times 10^{12} \text{ m}^2)\), \( Q_{sh_{-ml}} \) is the fraction of the total shelf outflow \( Q_{sh} \) that infloWS the mixed layer. The rest of the outflow contributes highly-saline and cold water to the Arctic halocline (explained in section 2.2). The import of shelf water to the interior Arctic Ocean equals the prescribed outflow from the mixed layer on the shelf \( (Q_{m_{ao}}) \). \( Q_{gML_{-atl}} \) is water volume flux from the mixed layer to the North Atlantic. The Pacific Water inflow through Bering Strait (hereinafter referred to as Bering Water, \( Q_{Ber} \)) has seasonally varying salinity \( (S_{Ber}) \) and temperature \( (T_{Ber}) \). The Bering Water is isopycnally mixed with the Arctic Ocean water flowing either to the mixed layer or halocline depending on the mixed layer density: \( \rho_{Ber} = \begin{cases} 1, & \rho_{Ber} < \rho_{ml} \\ 0, & \rho_{Ber} > \rho_{ml} \end{cases} \), where \( \rho_{ml} \) is density of the mixed layer. The Arctic Ocean exports ice \( (Q_{ice}) \) to the North Atlantic through Fram Strait and channels of the Canadian Archipelago. \( \varepsilon = \rho_{ice}/\rho_{aw} \approx 0.9 \), where \( \rho_{ice} \) is density of the sea ice, and \( \rho_{aw} \) is density of sea water. \( \omega_e \) is the entrainment velocity.

[18] Similar to Björk [1989] and Stigebrandt [1981], the outflow \( Q_{g_{-atl}} \) is assumed to occur as geostrophically balanced coastal currents with the underlying Atlantic water at rest. The outflow is estimated by integrating the thermal wind equation across the flow. There are two major outlets for the Arctic Ocean water: Fram Strait and channels of the Canadian Archipelago. The total number of geostrophic outlets \( (N_{out}) \) is 1.5 assuming one full geostrophic outflow in the Fram Strait and half in the Canadian Archipelago (see Björk [1989] for more discussion on this parameter).

[19] According to Björk [1989], if the \( Y \)-axis is along the coast and the \( X \)-axis is across the flow with the coast located at \( x = 0 \), then the transport per unit depth across the outflow region is:

\[ q_{atl}(z) = \int_{0}^{s_a} v_g dx = -\frac{g}{\rho_{atl}} \int_{z}^{H_a} (\rho_{atl} - \rho(z)) dz, \]  

where \( v_g \) is the coast-parallel velocity of the geostrophic flow, \( s_a \) is an offshore distance where \( \rho (x, z) = \rho_{atl}, \rho_{atl} \) is the Atlantic water density, and \( H_a \) is total thickness of the mixed layer and halocline.

Figure 2. Diagram of the Arctic Ocean model. Three layers in the model: AL – Atlantic layer, HL – halocline layer, and ML – mixed layer. \( Q_{m_{ao}} \) is the outflow from the mixed layer to the shelf. \( Q_{sh} \) is the import of shelf water to the interior Arctic Ocean. \( Q_{sh_{-ml}} \) is the fraction of \( Q_{sh} \) inflowing into the mixed layer. \( Q_{Ber} \) is the Bering Strait inflow with seasonally varying salinity \( (S_{Ber}) \) and temperature \( (T_{Ber}) \). \( Q_{ice} \) is the ice export from the Arctic Ocean and \( Q_{g_{-atl}} \) is the water export from the Arctic Ocean to the North Atlantic through Fram Strait and channels of Canadian Archipelago. \( Q_{gML_{-atl}} \) is the fraction of the Arctic Ocean water outflow from the mixed layer. \( h_{ml} \) is mixed layer thickness, \( h_w \) – mixed layer thickness measured from the base of the ice, \( w_{1o} \) – wind, \( h_{ice} \) – ice thickness, \( w_e \) – entrainment velocity, \( w_a \) – vertical velocity, and \( \rho(z) \) – water density.
[20] Water volume flux from the mixed layer to the North Atlantic is expressed as:

\[ Q_{gel\_ml} = \lambda_{out} \int_0^{h_{ml}} q_{ml}(z) dz, \]  

and the total outflow from the mixed layer and halocline is:

\[ Q_{gel\_all} = \lambda_{out} \int_0^{h_{ml}} q_{ml}(z) dz. \]  

The sign convention is \( Q_{gel\_all} < 0 \), meaning water export from the Arctic basin.

[21] The role of sea ice in the coupling of the Arctic Ocean and Greenland Sea is not our primary focus so the model is designed such that the long-term mean ice thickness is constant. All produced ice is exported from the Arctic model. The ice export \( \left( Q_{ice} \right) \) from the Arctic Ocean is a function of net gain or loss of ice in the Arctic relative to the previous year:

\[ Q_{ice} = P_{rl\_ao} \cdot A_{AO}, \]  

where \( P_{rl\_ao} \) is the annual mean ice production from the previous year.

[22] The entrainment velocity is given by:

\[ w_e = -\frac{1}{g'} \left[ \frac{2m_0(a_0)}{h_w} - \kappa \cdot B_f \right], \]  

where \( g' \) is reduced gravity, \( h_w \) is the mixed layer depth measured from the base of the ice, \( m_0 \) and \( \kappa \) are proportionality coefficients which are discussed later. The friction velocity \( (u*) \), is determined by wind from relations in Thordike and Colony [1984], and the buoyancy flux at the sea surface \( (B_f) \) is defined as:

\[ B_f = g' \left\{ \frac{1}{A_{AO}} [Q_{sh\_ml}(S_{ml} - S_{sh\_ml}) + \nu_{Ber} Q_{Ber}(S_{ml} - S_{Ber})] 
- \epsilon P_{rl\_ao}(S_{ml} - S_{ice}) \right\}, \]  

where \( g \) is gravitational acceleration, \( P_{rl\_ao} \) is ice production in the Arctic Ocean and \( \beta \) is the coefficient of salt contraction.

[23] When \( B_f \) is positive \( (w_e \) may be negative or positive), the depth of the mixed layer is determined from Monin-Obukhov, Ekman length scales or equation (1) depending on values of \( B_f \) and \( w_e \). The detailed description of the length scales used for different dynamical regimes is given by Stigebrandt [1985]. In summary, when \( B_f > 0 \) and \( w_e > 0 \) (retreat of the mixed layer) then the depth of the mixed layer is defined by the minimum value between \( h_{Ekml} \) and \( h_{Ob} \): \( h_{ml} = \min \{ h_{Ob}, h_{Ekml} \} \), where \( h_{Ob} \) is the Monin-Obukhov length \( \left( h_{Ob} = \frac{2m_0(a_0')}{B_f} \right) \) and \( h_{Ekml} \) is the Ekman length \( \left( h_{Ekml} = \frac{k \cdot u^*}{f} \right) \). If \( B_f > 0 \) and \( w_e \leq 0 \) (pure forced convection) then \( h_{ml} = \min \{ h_{Ob}, h_{Ob}, h_{Ekml} \} \).

[24] Salinity changes in the mixed layer are computed from:

\[ \frac{dS_{ml}}{dt} = \frac{1}{h_w} \left\{ \frac{1}{A_{AO}} Q_{sh\_ml}(S_{sh\_ml} - S_{ml}) + \epsilon P_{rl\_ao}(S_{ml} - S_{ice}) 
- \nu_{Ber} Q_{Ber} \right\}, \]  

where \( Q_{sh\_ml} \) is the output (volume transport) from the shelf box model (discussed in the shelf model section), \( S_{sh\_ml} \) is the integrated salinity of the shelf outflow into the mixed layer (discussed later), and \( S(h_{ml\_i}) \) is salinity just below the mixed layer.

[25] The proportionality coefficient, \( m_0 \), that parameterizes dissipation of forced convection in equation (6) is:

\[ m_0 = \log \left[ \frac{100}{Ri_0} \right] + 3.5, \]  

\[ m_0 = \begin{cases} \max(0, m_0) \\ \min(8, m_0) \end{cases}, \]  

where \( R_i_0 = \frac{h_w \cdot g'}{\nu_{Ber} g} \) is bulk Richardson number. Plausible mixed layer dynamics are obtained with \( P \) ranging from 1.2 to 2.0. In the model experiments \( P = 1.2 \) for the Arctic Ocean model and \( P = 1.8 \) for the Greenland Sea model.

[26] The mixed layer deepening rate driven by thermohaline convection is determined by the proportionality factor \( \kappa \). The \( \kappa \) coefficient parameterizes dissipation of the convectively produced turbulence in equation (6). \( \kappa \) is a function of the water column stability: when stability is high the dissipation of convection is high as well. For the Arctic Ocean, which is characterized by year-round strong stratification, the \( \kappa \) coefficient is set to be 0.05 in agreement with Björk [1989].

2.1.2. Halocline Layer

[27] The halocline layer water temperature \( (T) \) and salinity \( (S) \) are calculated from:

\[ \frac{\partial S}{\partial t} = \frac{\partial S}{\partial z} + D_z \frac{\partial T}{\partial z} + \frac{q_{sh}}{A_{AO}} (S(sh) - S(z)) + \nu_{Ber} \frac{Q_{Ber}}{A_{AO}} \cdot (S_{Ber} - S(z)), \]  

\[ \frac{\partial T}{\partial t} = \frac{\partial T}{\partial z} + D_z \frac{\partial T}{\partial z} + \frac{q_{sh}}{A_{AO}} (T(sh) - T(z)) + \nu_{Ber} \frac{Q_{Ber}}{A_{AO}} \cdot (T_{Ber} - T(z)), \]  

where \( w_e \) is the vertical velocity, \( D_z \) is the coefficient of eddy diffusivity, \( S(sh) \) and \( T(sh) \) are salinity and temperature of the shelf outlet at a given vertical grid point where densities of the shelf water and halocline layer are the same (computing \( q_{sh} \) is presented in the shelf model section). The semi-implicit Crank-Nickolson scheme with Thomas algorithm [Fletcher, 1988] has been applied to get a numerical solution of equation (11). The vertical resolution \( (h_l) \) in the model is 1 m.
Figure 3. Diagram of the shelf box model. $S_{sh}$ — shelf water salinity; $T_{sh}$ — shelf water temperature, under ice-free conditions it is a function of surface heat flux ($F_{tot}$); $\phi$ — salt flux to the interior Arctic Ocean; $w_{Q_f}$ — wind; $Q_f$ — river runoff; $Q_{mao}$ — inflow from the Arctic Ocean mixed layer; $Q_{atl\_sh}$ — Atlantic water inflow to the shelf; $Q_{(P-E)}$ — net precipitation; $Q_{sh}$ — ice export from shelf; and $Q_{sh}$ — shelf flux to the Arctic Ocean. Upper part of the shelf flux ($Q_{sh\_ml}$) goes to the Arctic Ocean mixed layer.

[28] The vertical velocity is calculated from the conservation of volume:

$$w_a(z) = \begin{cases} -\frac{1}{A_{40}} \int_0^z q_{Sat}(z')dz' + \frac{S_a}{S_{40}}(z) & \text{for } z < z_{Ber} \\ -\frac{1}{A_{40}} \int_0^z q_{Sat}(z')dz' + \frac{S_a}{S_{40}}(z) & \text{for } z > z_{Ber} \end{cases} \quad (12)$$

where $q_{Sat}(S)$ is the shelf water outflow. This term, $S_{sh}$ and $S_{sh\_out}$ are explained in section 2.2. $z_{Ber}$ is the depth level of the Bering Water injection.

### 2.1.3. Atlantic Layer

[29] The Atlantic layer has constant temperature ($T_{atl} = 0.5^\circ C$) and salinity ($S_{atl} = 34.8$).

### 2.2. Arctic Module: Shelf Box Model

[30] The Arctic shelf model is included in order to better reproduce fresh water and salt fluxes into the Arctic module. The area of the shelf region excluding the Barents Sea ($A_{sh}$) is 0.41 $\times 10^{13}$ m$^2$. A diagram of the shelf box model is shown in Figure 3. The shelf is represented with a box 50 m deep ($h_{sh}$) that exchanges water with the interior Arctic basin. There are four prescribed inflows: mixed layer water ($Q_{mao}$), river runoff ($Q_f$), Atlantic water brought onto the shelf through upwelling and advection from the Barents Sea ($Q_{atl\_sh}$), and net precipitation ($Q_{(P-E)}$) which is precipitation minus evaporation. Arctic mixed layer water enters the box and is transformed by ice production or ice melting and freshwater inflow. The outflowing water is interleaved in the interior Arctic Ocean at the isopycnal levels.

[31] The total volume outflow from the shelf to the interior basin ($Q_{sh}$) is equal to $Q_{mao}$, $Q_{sh}$ is parameterized by means of an outflow function (Figure 4). $q_{sh}(S)$, which is a function of salinity:

$$Q_{sh} = \int_{S_{sh}}^{S_{xx}} q_{sh}(S) dS. \quad (13)$$

The outflow from the shelf box is baroclinic and is constructed such that the intensity of the shelf outflow decreases as its salinity increases with depth (Figure 4a). The largest outflow ($q_{sh}$) corresponds to $S_{sh}$ and then decreases towards higher salinities, becoming zero at some threshold value $S_{xx}$. The area under the $q_{sh}(S)$ line equals the total shelf outflow $Q_{sh}$. It is obvious that $S_{xx}$ depends on the ice production rate on the shelf. The less ice produced on the shelf, the less saline is the water flowing to the halocline. Thus $S_{xx}$ is moving along the $S$ (salinity) axis reaching $S_{max}$ under the highest ice production rate (Figure 4a). When ice production is negative (Figure 4b) $S_{xx}$ is approaching $S_{sh}$ and $q_{sh}(S)$ becomes a Dirac delta function:

$$\begin{cases} q_{sh}(S) = 0, & S \neq S_{sh} \\ \int q_{sh}(S) dS = q_{sh}(S_{sh}) = q_0 = Q_{sh}, & S = S_{sh} \end{cases} \quad (14)$$

[32] An important characteristic of the shelf-interior Arctic basin interaction in the model is the total outflow of salt ($\phi$):

$$\phi = \int_{S_{sh}}^{S_{xx}} q_{sh}(S) \cdot S \ dS. \quad (15)$$

Also, it can be shown that $\phi$ is a function of ice production:

$$\phi = (1 - \xi) \cdot \xi \cdot Q_{sh}(S_{sh} - S_{xx}) + Q_{sh}S_{sh}, \quad (16)$$

where $\xi$ is a fraction of the rejected salt consumed to increase the shelf water salinity, namely, this amount of salt remains on the shelf: $\xi =\begin{cases} 0.2, & \text{if } Pr_{sh} > 0 \\ 1, & \text{if } Pr_{sh} \leq 0 \end{cases}$ and $Pr_{sh}$ is ice production on the shelf.

[33] Parameter $S_{xx}$ determines the flow to the mixed layer and halocline of the Arctic Ocean model. In Björk [1989], $S_{xx}$ is prescribed and constant (no seasonal variation), and...
ice production on the shelf is obtained from the salt balance. In the presented model, the opposite approach is used: ice production is calculated in the thermodynamic sea ice model, which reproduces both seasonal and interannual signals, and $S_{xx}$ is determined from the ice production. The following relationship between the ice production expressed in terms of $\phi$, total shelf outflow $Q_{sh}$ (which equals the prescribed $Q_{mao}$) and $S_{xx}$ has been derived:

$$Q_{sh} = \frac{\phi(S_{sh} - S_{bd})^2}{S_{xx}(S_{xx}^2 + S_{sh}^2) - \frac{1}{2}(S_{xx}^2 - S_{sh}^2)}$$

(17)

The unknown $S_{xx}$ is obtained by using the Newton-Raphson iteration technique [Kreyszig, 1999].

[34] Water temperature in the shelf model ($T_{sh}$) is calculated from different equations depending on whether or not there is ice cover in the model. When there is no ice:

$$\frac{dT_{sh}}{dt} = \frac{1}{h_{sh}} \left( \frac{F_{int}}{\rho_{sh} C_{wp}} + \frac{1}{A_{sh}} \left[ Q_{mao} \cdot (T_{ml} - T_{sh}) + Q_{f} \cdot (T_{f} - T_{sh}) \right] + \left( T_{f} - T_{sh} \right) \right) \cdot Q_{sh}$$

(18)

where $F_{int}$ is the surface heat flux (positive to the ocean), $\rho_{sh}$ is the shelf water density, $C_{wp}$ is the water specific heat, $T_{atm}$ is temperature of the Atlantic, and $T_{f}$ is the river runoff temperature. When ice appears in the model ($h_{ice} > 0$), temperature of the shelf water changes according to:

$$\frac{dT_{sh}}{dt} = \frac{1}{h_{sh}} \left( \frac{F_{int}}{\rho_{sh} C_{wp}} + \frac{1}{A_{sh}} \left[ Q_{mao} \cdot (T_{ml} - T_{sh}) + Q_{f} \cdot (T_{f} - T_{sh}) \right] + \left( T_{f} - T_{sh} \right) \right) \cdot Q_{sh}$$

(19)

where $F_{int}$ is positive to the ocean and $\rho_{sh}$ is the shelf water density.

[35] Salinity changes in the shelf model are described by the equation:

$$\frac{dS_{sh}}{dt} = \frac{1}{h_{sh}} \left\{ \frac{1}{A_{sh}} \left[ Q_{mao} (S_{ml} - S_{sh}) - Q_{r} \cdot (S_{sh} - Q_{PCE}) \cdot S_{sh} \right] + \xi \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot \cdot …
where $H_{\text{atm}} = 8 \times 10^3$ m is the height of the atmospheric box, $\rho_{\text{air}} = 1.37$ kg m$^{-3}$ is the air density, and $C_p = 1012$ J kg$^{-1}$ K$^{-1}$ is the air specific heat. The total energy gain or loss in the box ($F_{\text{tot,a}}$) is estimated through the energy balance:

$$F_{\text{tot,a}} = F_{\text{tot}(0)} + F_{\text{tot}(1)} + F_{\text{adv}},$$

(24)

where $F_{\text{tot}(0)}$ is the energy balance at the bottom atmospheric box boundary (atmosphere – ice), $F_{\text{tot}(1)}$ is the energy balance at the upper atmospheric box boundary (atmosphere – space), and $F_{\text{adv}}$ is sensible heat advected from the Greenland Sea box. The energy balance at the bottom boundary ($F_{\text{tot}(0)}$) is calculated in the ice model of the Arctic Ocean.

[30] The energy balance at the upper boundary ($F_{\text{tot}(1)}$) is computed from the equations in Wallace and Hobbs [1977]. The heat advection to the Arctic ($F_{\text{adv}}$) depends on the SAT gradient between the Greenland Sea and Arctic boxes. This idea is adopted from Marotzke and Stone [1995] and Scott et al. [1999], who investigated atmospheric meridional transports and ocean-atmosphere interactions. Marotzke and Stone [1995] used a simplified parameterization assuming that the transport is proportional to the meridional temperature gradient to describe the meridional atmospheric heat transport between “cold” and “warm” boxes in their model:

$$F_{\text{adv}} = \chi(T_2 - T_1),$$

(25)

where $\chi$ is the coefficient of heat advection, $T_2$ and $T_1$ are SAT in the warm and cold boxes respectively. The physical meaning of $\chi$ is the following. Because transient eddies (traveling highs and lows) are responsible for most of the meridional atmospheric transports in high latitudes [Marotzke and Stone, 1995], the coefficient $\chi$ is used to regulate the intensity of meridional heat transport, allowing higher transport as the temperature gradient increases. This simple relation (equation (25)) captures the salient feature of Stone and Yao’s [1990] parameterization of meridional eddy heat and moisture fluxes: as the meridional temperature gradient increases eddy activity increases as well. Marotzke and Stone [1995] analytically estimated this coefficient to be $1.3$ W m$^{-2}$ K$^{-1}$. It describes the heat diffusion into the “cold” box in their simplified ocean-atmosphere box model. In our model, the estimate of $\chi$ is obtained empirically from a number of model experiments (see section 4 for brief discussion).

2.5. Greenland Sea Module: Ocean Model

[40] The oceanic model describes the development of the thermohaline structure in the Greenland Gyre. The model consists of several layers shown in Figure 6. The appearance and growth of the ice depends on the heat budget of the sea surface and characteristics of the mixed layer.

**Figure 5.** Diagram of the Arctic atmospheric box model. Numbers denote the fluxes: 1 – incoming solar radiation ($F_{\text{ins}}$); 2 - incoming on the ice shortwave radiation ($F_{\text{shw_ins}}$); 3 – back-scattered by air solar radiation ($F_{\text{rfl(1)}}$); 4 – reflected by ice surface ($F_{\text{rfl(2)}}$); 6 - longwave flux at the top of the atmosphere ($F_{\text{lw(1)}}$); 7 – energy balance at the upper boundary ($F_{\text{tot}(1)}$); 8 - absorbed shortwave radiation by the atmosphere ($F_{\text{abs(shw)}}$); 9 - penetrated into ice shortwave radiation ($F_{\text{shw_abs}}$); 10 - surface heat flux at the lower boundary ($F_{\text{tot}(0)}$); and 11 – meridional sensible heat transport from the Greenland Sea domain ($F_{\text{adv}}$).
Evolution of the ice cover is simulated by the ice model and is identical to that in the Arctic Ocean module (section 2.3).

The inflow in the Greenland Sea model is presented in Figure 6. To conserve mass, outflow from each layer equals the inflow. The Polar Water inflow \( Q_{PW} \) is a fraction of the Arctic Ocean outflow \( Q_{g,\text{atl}} \) integrated from the surface to 150 m depth. The temperature \( T_{PW} \) and salinity \( S_{PW} \) of \( Q_{PW} \) are integrated values of \( Q_{g,\text{atl}} \) within the same depth range.

It should be noted that there is not enough observational data to explain how surface water is formed in the central Greenland Sea. It is generally believed that the Greenland Gyre experiences a very small effect from the East Greenland Current (EGC), receiving very little of the Polar Water [Swift, 1986; Aagaard and Carmack, 1989, 1994; Alekseev et al., 1994]. Also it is assumed that the inflow rates of the Polar Water and Atlantic Water have seasonal variability. In winter, when the cyclonic vortex in the Greenland Gyre is intensified, the surface water is forced away from the Gyre center and very little ambient water can penetrate the Arctic front [Swift, 1986]. However in summer, when cyclonic vorticity is significantly weakened, the ambient surface water and sea ice can easily penetrate the front and reach the center of the Gyre [Johannesmessen, 1986]. This idea is well corroborated by oceanographic observations [Pawlowicz, 1995] showing freshening of the central Greenland Sea in summer. During summer, the EGC advects ice into the Greenland Sea model where it melts \( Q_{\text{melt, GS}} \).

The central part of the upper Greenland Sea consists of modified, diluted and cooled Atlantic Water. This water mass (Arctic Surface Water) results from mixing PW and AW and is modified by air-sea interaction. The PW is 150 m thick and includes water from the Arctic Ocean mixed layer and halocline. Thickness of the AW \( H_{AW} \) flowing into the central Greenland Sea is also 150 m. Thickness of the PW and AW jets is based on Swift [1986] and Swift and Aagaard [1981]. Assuming that most interaction between the PW and AW occurs in the very upper part of the central Greenland Sea, the vertical structure of the flows is approximated with linear functions \( q_{AW} \) and \( q_{PW} \), respectively:

\[
q_{AW}(z) = \begin{cases} 
q_{0,AW} - \frac{q_{0,AW}}{H_{AW}} |z|, & |z| \leq H_{AW} \\
0, & |z| > H_{AW},
\end{cases}
\]

\[
q_{PW}(z) = \begin{cases} 
q_{0,PW} - \frac{q_{0,PW}}{H_{PW}} |z|, & |z| \leq H_{PW} \\
0, & |z| > H_{PW},
\end{cases}
\]

where \( q_0 \) is the transport at the surface.

Parameterization of the Greenland Sea mixed layer dynamics is similar to that of the Arctic Ocean. The length scale \( h_{ml} \) in the Greenland Sea is defined as:

\[
\frac{dh_{ml}}{dt} = w_{e} + w_a.
\]

The entrainment velocity is given by equation (6) with \( \kappa \) parameterized by the water column stability and buoyancy flux.

Greenland Sea model sensitivity runs have shown that \( \kappa = 0.05 \) is too low and underestimates the role of convection in the mixed layer deepening (see section 4). Deep mixing in the Greenland Gyre occurs in small-scale “chimneys” driven by intense wintertime convection. We assume that the mixed layer in the Greenland Gyre is deep when “chimney” convection occurs in the region. To permit “chimney” scale convection, the Richardson number is defined by [Chapman, 1997]:

\[
Ri = g' h_{ml} (|B_T| r_0)^{-2/3},
\]

where \( h_{ml} \) is the mixed layer depth, \( B_T \) is the buoyancy flux through the surface, and \( r_0 \) is the radius of a convective region taken to be 50 km. According to Chapman [1997], if the Richardson number \( Ri \) is greater than the transition Richardson number \( Ri_T \) which equals 8, the density gain in the upper layer (due to negative buoyancy flux) cannot overcome the density jump between the upper and lower layers. In this case convection should not penetrate into the lower layer and convection is strongly suppressed leading to a very small \( \kappa \). If \( Ri < Ri_T \) then, according to Chapman [1997], convection penetrates into the lower layer very fast and is not depth limited, assuming large \( \kappa \). From the above theoretical discussions and model experiments, the following expressions for \( \kappa \) have been obtained when \( B_T < 0 \):

\[
\kappa = \begin{cases} 
0.9, & \text{if } Ri_0 < Ri_T \\
0.01 \cdot \log(1000 / h_{ml}), & \text{if } Ri_0 > Ri_T
\end{cases}
\]

When \( B_T > 0 \), \( \kappa = 1 \).
The vertical velocity is estimated through the Ekman pumping – wind stress curl relation [Cushman-Roisin, 1994]:

\[
\frac{dw}{dt} = \frac{1}{\rho_{w} \cdot C_{wp}} \left( \frac{\partial \tau_{x}}{\partial y} - \frac{\partial \tau_{y}}{\partial x} \right) = \frac{1}{\rho_{w} \cdot C_{wp}} \text{curl}(\tau_{xy}).
\]  

(31)

The wind stress curl over the Greenland Sea is adopted from Jonsson [1991].

The buoyancy flux to the mixed layer is calculated using the following equation [Turner, 1973; Stigebrandt, 1981, 1985]:

\[
B_{b} = g \cdot \left\{ \frac{\alpha}{\rho_{ml} \cdot C_{wp} \cdot F_{tot}} \cdot \frac{\beta}{A_{GS}} \cdot \left( \int_{0}^{h_{ml}} q_{aw} \cdot dz + \int_{0}^{h_{ml}} q_{pw} \cdot dz \right) \right. \\
\left. - \varepsilon \cdot \left( S_{ml} - S_{ice} \right) \left( Pr_{i,GS} + \frac{Q_{ml,GS}}{A_{GS}} \right) \right\}
\]  

(32)

where \( \alpha \) and \( \beta \) are the coefficients of heat expansion and salt contraction, respectively; \( \rho_{ml} \) is the mixed layer density calculated through the equation of state, \( C_{wp} = 4184.4 \text{ J m}^{-2} \text{sec}^{-1} \) is the water specific heat, \( F_{tot} \) is the air-sea heat flux (positive for the flux to the ocean), \( A_{GS} \) is the area of the central Greenland Sea, and \( Pr_{i,GS} \) is ice production in the region. \( Q_{ml,GS} \) is the volume of ice advected from the EGC and melted in the central Greenland Sea during summer.

The salinity changes in the mixed layer are described by the salt balance equation:

\[
\frac{dS_{ml}}{dt} = \frac{1}{h_{z}} \left\{ \frac{1}{A_{GS}} \cdot \left( \int_{0}^{h_{ml}} q_{aw} \cdot dz + \int_{0}^{h_{ml}} q_{pw} \cdot dz \right) \right. \\
- \varepsilon \left( S_{ice} - S_{ml} \right) - \frac{Q_{ml,GS}}{A_{GS}} - w_{e} \\
\left. \cdot \left[ S(h_{ml+1}) - S(h_{ml}) \right] \right\}
\]  

(33)

where \( Pr_{i,GS} \) is the ice production/melting rate (if there is any ice) and \( S(h_{ml+1}) \) denotes the salinity just below the mixed layer. Equation (33) is valid only for downward entrainment velocity \( (w_{e} < 0) \), otherwise (during the mixed layer retreat) there is no \( w_{e} \) term in the formula.

Under ice free conditions the temperature changes are given by:

\[
\frac{dT_{ml}}{dt} = \frac{1}{h_{z}} \cdot \left\{ \frac{F_{tot}}{\rho_{w} \cdot C_{wp}} + \frac{1}{A_{GS}} \cdot \left( \int_{0}^{h_{ml}} q_{aw} \cdot dz + \int_{0}^{h_{ml}} q_{pw} \cdot dz \right) \right. \\
- \frac{w_{e} \cdot (T(h_{ml+1}) - T(h_{ml}))}{h_{z}} \right\}
\]  

(34)

where \( T(h_{ml+1}) \) is the water temperature just below the mixed layer. When accretion begins, \( F_{tot} \) in equation (34) is replaced with the heat flux from the water to the ice bottom \( (F_{w}) \) parameterized by the relation from Maykut and Untersteiner [1969].

Below the mixed layer the local rate of change of the salinity and the temperature are:

\[
\frac{dS}{d\tau} = \frac{\partial S}{\partial \tau} = w_{a} + D_{r} \frac{\partial^{2} S}{\partial z^{2}} + \frac{q_{in}(z)}{A_{GS}} \left[ S_{in} - S(z) \right]
\]

\[
\frac{dT}{d\tau} = w_{a} + D_{r} \frac{\partial^{2} T}{\partial z^{2}} + \frac{q_{in}(z)}{A_{GS}} \left[ T_{in} - T(z) \right]
\]  

(35)

where \( q_{in} \) denotes an inflow of water with temperature \( T_{in} \) and salinity \( S_{in} \) at a given depth \( z \).

2.6. Greenland Sea Module: Statistical Model of SAT

In the suggested mechanism of decadal variability, the FW ocean perturbations from the Arctic to the GIN Sea influence the atmosphere through convection, heat flux, and the sea ice freeze/melt cycle. Surface heat flux anomalies induce SAT anomalies in the region [Häkkinen and Cavalieri, 1989; Häkkinen, 1995] and a regional relation between surface heat flux and SAT anomalies is sought. A multiple regression model [Chatfield, 1996] has been developed to predict the next-day SAT anomaly given the previous day’s heat flux and SAT.

A high-order multiple regression model with a backward elimination technique [Jennrich, 1995] has been applied to select the final model:

\[
Y_{i} = \alpha_{i} Y_{i-1} + \alpha_{2} Y_{i-2} + \alpha_{3} Y_{i-3} + \beta_{1} X_{i-1} + \beta_{2} X_{i-2} + \varepsilon
\]  

(36)

where \( Y_{i} \) denotes SAT anomaly on the \( i \)th previous day, \( X_{i} \) is the total net flux anomaly on the \( i \)th previous day, \( \alpha \)'s and \( \beta \)'s are parameters to be estimated, \( \varepsilon \) is an error which is, in the case of a good fit, an independent Gaussian random variable.

The model given by equation (36) differs from usual simple linear regression models in that there are two explanatory variables \( (X \text{ and } Y) \). The first part of the model containing \( Y \)'s is a third-order auto-regressive model. The second part with \( X \)'s describes auto-correlated structure of the residual. Both model selection and parameter estimates have been done using SAS – the statistical programming language. The SAS output for the final model (equation (36)) is given in Tables 3 and 4. The over-all fit test (\( F \)-test) proves a good fit. All \( t \)-tests reveal that the parameters are significantly different from zero at confidence level \( \alpha = 0.05 \) (except for intercept which is zero).

To validate the SAT statistical model, a 1-day forecast of SAT in the central Greenland Sea has been run for 1990–2000 using the National Centers for Environmental Prediction (NCEP) reanalysis data. The statistical model gives very accurate 1-day prediction (Figure 7a). The histogram (Figure 7b) and autocorrelation function (Figure 7c) of the residual show that the error term \( \varepsilon \) is white noise with a standard normal distribution; this also supports the idea that the model
Table 3. Analysis of Variance and F-Test of the Overall Fit

<table>
<thead>
<tr>
<th>Source</th>
<th>DF</th>
<th>Sum of Squares</th>
<th>Mean Square</th>
<th>F value</th>
<th>P value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Corrected total</td>
<td>4011</td>
<td>3648.04</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Error</td>
<td>4006</td>
<td>1441.97</td>
<td>0.35995</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Model</td>
<td>5</td>
<td>2206.07</td>
<td>441.21</td>
<td>1225.76</td>
<td>&lt;0.0001</td>
</tr>
</tbody>
</table>

aDegrees of freedom, which is (number of observations – (number of parameters plus intercept)).
bSum of squared errors; errors are the difference between the model and observations.
cMean Square Error = Sum of Squares/DF.
dTest for the overall model fit, null hypothesis (H0) is that all coefficients in the model are zero (except for the intercept); F = Mean Square Model/ Mean Square Error.
eProbability that a random value from F(5,4006) distribution will exceed the observed F = 1225.76. If P-value is less than the confidence level 0.05, the null hypothesis is rejected.
fRestricted model with all coefficients (parameters) zero except for the intercept.
gComplete (tested) model (equation (36)).
hStatistic is calculated from columns a and b obtained by subtraction of row i from j.

(equation (36)) provides reliable 1-day forecasts of SAT anomalies in the region.

3. Model Calibration and Validation Results

[55] Calibration and validation experiments and sensitivity studies were conducted for each module and for the entire model. The forcing parameters applied in the model run are presented in Figures 8–10.

3.1. Model Initialization

[56] The full model was run to validate the behavior of the simulated system, i.e. to estimate how accurately the model reproduces the annual cycle. During the model run, daily values of the oceanic (T, S, hroz, wroz, Broz), ice (hice, Prroz, Prroz, Prroz, Prroz), and atmospheric (TB) characteristics and fluxes between the sub-models were calculated and compared with observations or output from other modeling studies. For this experiment, the statistical model of the SAT in the Greenland Sea was not employed. Instead, the SAT was assigned the daily mean values calculated from the NCEP reanalysis (Figure 10).

[57] The model was initialized with multiyear mean ocean temperature and salinity data and atmospheric SAT for the central Arctic, the Eurasian shelf, and the Greenland Sea acquired from the Environmental Working Group (EWG) [1998] atlas, the NOAA-CIRES Climate Diagnostics Center (CDC), and the Polar science center Hydrographic Climatology (PHC, http://psc.apl.washington.edu/POLES/PHC2) [Steele et al., 2001]. The outflow from the Arctic Ocean to the Greenland Sea (QAC_ec) was computed in the Arctic Ocean (equation (4)). The heat advection to the Arctic was obtained from equation (25) with the coefficient of heat advection (χh) set to 1.8 W m−2 K−1. Water export from the Arctic Ocean mixed layer to the shelf (Qm) was taken to be 1.1 Sv.

Table 4. Parameter Estimates and Individual t-Tests

<table>
<thead>
<tr>
<th>Variable</th>
<th>Parameter Estimate</th>
<th>Standard Error</th>
<th>t Statistics</th>
<th>P Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intercept</td>
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<td>-0.02</td>
<td>0.9861</td>
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<td>0.02617</td>
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<tr>
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<td>&lt;0.0001</td>
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<tr>
<td>β2</td>
<td>-0.12</td>
<td>0.01976</td>
<td>-6.07</td>
<td>&lt;0.0001</td>
</tr>
</tbody>
</table>

aEstimates for the intercept, α’s and β’s in the model (equation (36)).
bSTD of the estimates.
cIndividual t-tests for testing the hypothesis α = 0, α = 0, α = 0, β = 0, and β = 0.
dP-value, the probabilities that the absolute value of the corresponding t-statistic will exceed that of the t-value given, under the standard normality assumptions and assuming that the true parameter is zero. When P-value is less than the confidence level 0.05, the null hypothesis is rejected.

Figure 7. One day prediction from the statistical model of SAT anomalies for the central Greenland Sea for 1990–2000. (a) NCEP reanalysis SAT data (gray curve) and model predicted SAT (black curve) shown for 1991. (b) and (c) Analysis of the residual (1990–2000). The histogram (Figure 7b) and auto-correlation function (Figure 7c) of the residual (predictions minus observations) unambiguously prove that the error term (e) is white noise with a standard normal distribution. Note scale of the vertical axis in Figure 7c; r(i) is 1 at zero lag (not shown).

3.2. Simulated and Observed Annual Cycles

[58] The model was run for 110 years. The simulated system reaches steady state after approximately 10 years. The results show that the mean annual cycle was reproduced in the sub-models for the Arctic, shelf, and Greenland Sea.

3.2.1. Arctic Ocean

[59] The calculated mixed layer thickness and salinity are presented in Figures 11a and 11b. The Arctic mixed layer is deepest in winter when it reaches 30–60 m and much shallower (up to 5–20 m) in summer [Aagaard et al., 1981; Stigebrandt, 1981; Swift et al., 1997]. The model reproduces a shallow mixed layer in summer (~10 m) which deepens during the cold season and reaches the maximum thickness (~27 m) in April–May (Figure 11a). Note the rapid shallowing of the mixed layer in the early
summer and the slow deepening during the cold season which has been reported by Lenke and Manley [1984].

The mixed layer water in the Arctic Ocean is characterized by low salinities which vary from above 34 psu north of Svalbard to below 32 psu in the Makarov Basin [Coachman and Aagaard, 1974; Rudels, 1998]. Zonally integrated salinity of the upper 10 m obtained from PHC data is plotted in Figure 11b and reveals a noticeable seasonal signal. The simulated salinity (black curve) in the mixed layer reproduces seasonality with minimum in summer and maximum by the end of the cold season. The range of the simulated seasonal salinity is in the range of observed values in the interior Arctic Ocean.

Ice thickness distribution and ice production vary significantly in time and space over the Arctic Ocean [Hibler, 1980; McLaren et al., 1994; Wadhams, 1994]. For the purpose of this study, we reproduced the average ice production in the central Arctic. Figure 11c presents simulated ice thickness in the Arctic model which is reasonably close to results from the model study of Hibler [1979]. The onset of melting is in the middle of May, and freezing starts by the middle of September.

Computed and observed monthly mean SAT in the Arctic are presented in Figure 11d. The observed monthly means and their standard deviations have been estimated from the daily data for the central Arctic (86°/C176N poleward) for the period 1948–2001. The data are NCEP reanalysis acquired from NOAA-CIRES CDC (http://www.cdc.noaa.gov). The simulated SAT has a slight cold bias in spring, being at a lower bound of the 98% confidence interval of the NCEP reanalysis data.

Results from our model of the Arctic Ocean agree with output from the models of Björk [1989] and Alekseev and Ryabchenko [1996]. However, there are few quantitative discrepancies among the model results. For example, from a thermodynamical model of the sea ice – ocean system of Alekseev and Ryabchenko [1996], simulated mixed layer salinity ranges from ~30.8 (August) to 33.9 (June) which is too saline compared to the observations (Figure 11b). Also, the onset of freshening of the mixed layer (July) is delayed in the model of Alekseev and Ryabchenko (from our Figure 11b, observed freshening starts in May) resulting in delay of the mixed layer shallowing in summer (July).

Mixed layer salinity calculated in the Björk [1989] model seems to be underestimated. It changes from ~28.8
in September to 30.6 in May, lower than observed values (Figure 11b). Bjoerk's mixed layer depth variability is similar to our model shown in Figure 11a.

3.2.2. Shelf

Seasonal variability is well pronounced in the Arctic shelf seas (compare gray curves in Figures 12a and 12b of averages for the East-Siberian, Laptev, and Kara Seas). The shelf model reproduces the seasonal signal in shelf water salinity (Figure 12a) and temperature (Figure 12b). The model amplitude of the salinity variations is smaller, and the computed annual mean salinity is lower, than in observations.

Shelf water temperature stays near the freezing point during the winter and rapidly warms to -0.2°C after the ice melts (Figure 12c). The observed seasonal cycle of the water temperature in the upper 50 m layer on the shelf is smoother than simulated. The onset of warming is in May–June when river runoff increases and shelf water gains energy through leads. In the model, leads are not parameterized and hence, the shelf water is isolated from solar radiation until the ice is gone.

The ice model simulates 1.5 m ice on the shelf which melts by August. This agrees with the output from other models [Hibler, 1980; Hibler and Walsh, 1982; Polyakov et al., 1999] that simulate ice < 2 m on the most of the Eurasian shelf and low ice concentration in August and September.

3.2.3. Greenland Sea

The observed mixed layer depth in the GIN Sea varies substantially both in time and space. In summer and early autumn, the mixed layer is shallow (10–30 m) over the whole basin. Maximum thickness of the mixed layer in the GIN Sea is observed in March–May from analysis of NODC NOAA mixed layer depth data. Due to large differences in the upper water characteristics in that region [Swift, 1986], the mixed layer thickness is highly variable over the GIN Sea. The maximum winter mixed layer thickness is observed in the central northern and eastern Norwegian Sea (from 400 m to 800 m), while the mixed layer in the central and western Greenland Sea is shallow (above 200 m). The model reproduces shallowing of the mixed layer in the Greenland Sea in summer (~15 m) and rapid deepening during the cold season (Figure 13a). Maximum thickness of the mixed layer (about 380 m) is simulated in April, at the end of the winter.

Figure 11. Climatology from the Arctic model (black solid curves). (a) Mixed layer thickness. Dashed gray segments show the range of the mixed layer in the Eurasian and Canadian Basins for the winter and summer periods estimated from the thermohaline profiles in the EWG [1998] atlas. (b) Mixed layer salinity. Gray curves denote zonally averaged salinity in the upper 10 m estimated from data acquired from Polar Hydrographic Climatology [Steele et al., 2001]: solid - Eurasian Basin, 82.5N; dashed - Canadian Basin, 80.5N; curve with bullets - Canadian Basin, 76.5N. (c) Ice thickness from the thermodynamic sea ice model. Gray dashed curve shows average Arctic Basin ice thickness simulated by Hibler [1979]. (d) SAT. Solid gray curve denotes climatology obtained from NCEP Reanalysis data (NOAA-CIRES CDC) over the period 1948–2001. Dashed gray curves denote 98% confidence interval for the NCEP Reanalysis monthly means.

Figure 12. Climatology from the shelf model. (a) Shelf water salinity. Gray curves are mean salinities integrated in the upper 50 m from data acquired from Polar Hydrographic Climatology [Steele et al., 2001]: solid - Kara Sea; curve with bullets - Laptev Sea; dashed - East-Siberian Sea. (b) Similar to Figure 12a, but for the shelf water temperature. (c) Ice thickness from thermodynamic sea ice model.
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Climatology from the Greenland Sea model.
which parameterizes dissipation of free convection in the
decreases). Weak water column stability

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it follows that the winter

...In March the midgyre 
...
The simulated winter temperature presented in
Figure 13b is similar to the midgyre temperature at 160–
200 m shown by Pawlowicz [1995, Figure 4b]. From the
figure in Pawlowicz [1995] it follows that the winter
temperature at this depth is colder than –1°C, similar to
our simulated winter temperature and not the PHC results.

The general features of the annual, upper
layer salinity cycle are well reproduced in the model
(Figure 13c). The salinity increases during winter from
entrainment of more saline underlying water and from salt
rejection. With the onset of ice melt, in late May – early
June, the mixed layer freshens and stays diluted until the
start of thermohaline convection and ice freezing. There are
only minor discrepancies between the model and the obser-
vations. The behavior of the observed salinity in the upper
Greenland Sea is more complex (gray curves in Figure 13c)
than in the model. In particular, there are small peaks
imposed on the seasonal variability described earlier, per-
haps due to precipitation which is not taken into account in
the analyzed model. Also advection of fresh and salt water
into the Greenland Sea is not as regular and stationary as in
the model (Figures 9c and 9d).

4. Discussion

The Arctic Ocean model exhibits the best correspon-
dence to the observations compared to the other sub-models
because the dynamics and thermohaline changes of the
mixed layer are mainly controlled by the ice freeze/melt
cycle. This relationship is well reproduced in the model. In
winter, intense ice formation leads to the mixed layer
salinification (equation (8)). The water column becomes
less stable \((g' \text{ decreases})\). Weak water column stability
and strong and negative buoyancy flux at the surface
(equation (7)) promote a high entrainment velocity
(equation (6)) and deepening of the mixed layer
(Figure 11a). Also, the simulated shelf water has higher
salinity in winter (Figure 12a) due to lower river runoff
(Figure 9b) and higher ice production (Figure 12c). That
makes the salinity of the shelf water inflow to the mixed
layer \((S_{\text{shelf}})\) higher and, hence, reduces the positive con-
tribution to the buoyancy flux (equation (7)) promoting
mixed layer deepening. In summer, intense ice melt, first in
the shelf box and then in the Arctic domain, freshens the
upper layer, increases the water column stability, and stops
entrainment. The mixed layer then thins (Figure 11a).

Underestimated SAT in spring may result from the
absence of meridional heat transport in the atmo-
spheric model. As mentioned earlier, only sensible heat
transport is considered in the model (equation (25)). In
the real Arctic climate system, the latent heat transport plays
a significant role. However, the contribution of these two
transports to the meridional heat transport is different during
the year [Jackson and Broccoli, 2003]. It might be that in
spring the latent heat transport to the Arctic is essential and
neglecting it in the atmospheric model has led to the cold
bias.

The shelf model reproduces seasonality, although the
shelf water salinity is lower than the PHC data (Figure 12a).
Note that spatially averaged salinity decreases eastward as
the distance from the North Atlantic increases. The salinity
in the East-Siberian Sea is close to the simulated shelf
salinity. Hence, the most likely reason for the negative
salinity bias in the box model is an underestimation of the
advection of Atlantic water. Another reason for the fresh
bias may be the lack of dynamics in the sea ice model.

A sensitivity analysis of the model parameters has
been conducted (for details, see Dukhovskoy [2003]). The
key parameters controlling the annual cycle in the Arctic
Ocean and shelf models are the mixed layer outflow to the
shelf \((Q_{\text{shelf}})\), and the coefficient of heat advection \((\chi)\); in
the Greenland Sea model the key parameters are the ratio of
the Polar Water \((Q_{\text{PW}})\) and Atlantic Water \((Q_{\text{AW}})\) inflows
(equations (26) and (27)) and the proportionality coefficient
\(k\) which parameterizes dissipation of free convection in the
formula for entrainment velocity (equation (6)).

In the shelf model \((Q_{\text{shelf}} = Q_{\text{shelf}})\) determines the
amount of fresh water advected to the Arctic Ocean model.
The sensitivity experiments have been fulfilled for different
values of $Q_{\text{mix}}$ from 0.1 to 2 Sv. Low values lead to quick salinization of the upper Arctic Ocean, unrealistically deep mixed layer and unstable solutions. Stable and realistic behaviors of the Arctic Ocean model have been obtained for $Q_{\text{mix}} = 1.1$ to 1.8 Sv.

The coefficient of heat advection has been tested for values ranging from 0.5 to 4 W m$^{-2}$ K$^{-1}$. The sensitivity analysis has revealed that the lower values of $\chi$ lead to a colder Arctic as expected. The variability of SAT induced by different $\chi$ is larger in the cold season. However, the summer amplitude of SAT is small because the Arctic gains substantial solar radiation and heat advection is not a primary source of energy. The case $\chi = 0.5$ W m$^{-2}$ K$^{-1}$ results in an unrealistically cold Arctic. The opposite case $\chi = 4.0$ W m$^{-2}$ K$^{-1}$ leads to a very warm Arctic. When $\chi = 1.8$ to 2. W m$^{-2}$ K$^{-1}$ the model reproduces the observed SAT.

The dynamics of the mixed layer in the Greenland Sea model are very sensitive to the temperature and salinity of the upper and underlying layers. Thermohaline characteristics of the upper layer in the model are determined by the ratio of the inflow of Polar Water to Atlantic Water. A number of “tune-up” experiments has been run to adjust the unknown values of $Q_{\text{PW}}$ and $Q_{\text{ATW}}$. If the amount of Polar Water incoming to the Greenland Sea model is too high, the convection in winter is sluggish. In the opposite case, when the Atlantic Water inflow is too high, the convection is fast and reaches the bottom of the model water column in a few time steps. In particular, the annual mean ratio $Q_{\text{PW}}/Q_{\text{ATW}} \approx 3.7 \times 10^{-2}$ leads to low water column stability in the Greenland Sea, with highly possible deep convection. For $Q_{\text{PW}}/Q_{\text{ATW}} \approx 8.0 \times 10^{-2}$, there is strong freshening of the upper Greenland Sea and deep convection ceases. In these experiments the ratio is approximately $6 \times 10^{-2}$.

Another free parameter that determines upper layer dynamics is $\kappa$ (equation (6)). As discussed earlier, this coefficient parameterizes dissipation of the convectively produced turbulence. Different approaches to parameterize $\kappa$ have been tested. The major finding of this sensitivity analysis is that, although it works well for the Arctic Ocean, constant $\kappa$ results in false dynamics of the mixed layer in the Greenland Sea. Having $\kappa$ a function of the water column stability (equation (30)) gives more realistic behavior of the upper layer of the Greenland Sea model.

### 5. Summary

An idealized multibox model of the Arctic Ocean and GIN Sea has been developed, calibrated and validated. The model is used to investigate possible mechanisms of decadal climate variability in the Arctic based on existing concepts of Arctic decadal oscillations. The model presented in this paper realistically describes the major oceanographic and atmospheric characteristics in the Arctic and Greenland Sea over the annual cycle and could be used for investigation of interannual and decadal variability of Arctic climate.

From the validation experiment, we found that the behavior of the Arctic module is in good agreement with observations for the whole annual cycle and has similarities to the dynamics reproduced in the integral models of Björk [1989] and Alekseev and Ryabchenko [1996]. Compared with observations, our model shows better performance for the mixed layer salinity (Figure 11b). The atmospheric model simulates Arctic SAT reasonably close to the observations with a slight negative bias in spring.

The shelf model reproduces seasonality of the oceanic characteristics similar to the observations (Figure 12). The shelf water salinity is lower than the spatially averaged salinity in the upper 50 m of the Laptev, Kara, and East-Siberian seas. Based on the good performance of the Arctic Ocean model, it is believed that the small salinity bias in the shelf model is admissible because it has minimal impact on the solution in the interior domain.

The Greenland Sea model simulates seasonal variability in the ice-ocean component (recall that in the validation experiment, the model was forced with daily climatological mean SAT) in general agreement with observations (Figure 13). Winter mixed layer temperature in the model (Figure 13b) is colder than water temperature from PHC integrated over 10 and 75 meters but is in good agreement with Pawlowicz [1995]. Salinity calculated in the model is similar to the observed values (Figure 13c) which is evidence that the salt balance is well reproduced in the simulated system.

### Notation

- $A_{\text{AO}}$: area of the Arctic Ocean model domain ($0.61 \times 10^{13}$ m$^2$).
- $A_{\text{GS}}$: area of the Greenland Sea model domain ($0.135 \times 10^{13}$ m$^2$).
- $A_{\text{sh}}$: area of the shelf region ($0.41 \times 10^{13}$ m$^2$).
- $B_{\text{f}}$: buoyancy flux.
- $C_{\text{p}}$: specific heat (1012 J kg$^{-1}$ K$^{-1}$).
- $C_{\text{wp}}$: water specific heat (4184.4 J m$^{-3}$ K$^{-1}$).
- $D_{\text{z}}$: coefficient of vertical eddy diffusivity ($1. \times 10^{-6}$ m$^2$ s$^{-1}$).
- $F_{\text{adv}}$: meridional sensible heat advection to the Arctic atmospheric box model.
- $F_{\text{tot}}$: surface heat flux.
- $F_{\text{tot, a}}$: energy balance in the atmospheric box model.
- $F_{\text{s}}$: water heat flux to the ice bottom.
- $g$: gravitational acceleration.
- $g_{\text{r}}$: reduced gravity.
- $H_{\text{atm}}$: height of the atmosphere box (8 km).
- $H_{\text{AW}}$: thickness of the Atlantic Water flowing into the Greenland Sea model.
- $H_{\text{sh}}$: total thickness of the mixed layer and halocline.
- $H_{\text{Ek}}$: Ekman depth scale.
- $h_{\text{Ob}}$: Monin-Obukhov depth scale.
- $h_{\text{ice}}$: ice thickness.
- $h_{\text{ml}}$: mixed layer depth.
- $H_{\text{PW}}$: thickness of the Polar Water flowing into the Greenland Sea model.
- $h_{\text{sh}}$: depth of the shelf box.
- $h_{\text{shw}}$: depth of the shelf box measured from the base of the ice.
- $h_{\text{ml}}$: mixed layer thickness without ice draft.
- $m_{\text{0}}$: entrainment proportionality coefficient for dissipation of forced convection.
- $P_{\text{f, ao}}$: ice production in the Arctic Ocean model.
- $P_{\text{f, GS}}$: ice production in the Greenland Sea model.
\[
\begin{align*}
Pr_{i.sh} & \quad \text{ice production in the shelf model.} \\
Q_{atl.sh} & \quad \text{Atlantic water inflow to the Greenland Sea.} \\
Q_{atl} & \quad \text{Atlantic water inflow to the shelf.} \\
Q_{Ber} & \quad \text{the Bering Strait inflow to the Arctic Ocean.} \\
Q_{G.atl} & \quad \text{total Arctic Ocean outflow to the North Atlantic.} \\
Q_{ML.atl} & \quad \text{Arctic Ocean outflow to the North Atlantic from mixed layer.} \\
Q_{GSDW} & \quad \text{Greenland Sea Deep Water inflow to the Greenland Sea.} \\
Q_{ice} & \quad \text{ice export from the Arctic Ocean and shelf models.} \\
Q_{lwAIW} & \quad \text{lower Arctic Intermediate Water inflow to the Greenland Sea.} \\
Q_{m} & \quad \text{inflow to the shelf box from the Arctic mixed layer (1.1 Sv).} \\
Q_{sh} & \quad \text{volume of ice melted in the Greenland Sea model.} \\
Q_{NSDW} & \quad \text{Norwegian Sea Deep Water inflow to the Greenland Sea.} \\
Q_{P-E} & \quad \text{net precipitation flux.} \\
Q_{PW} & \quad \text{Polar Water inflow to the Greenland Sea.} \\
Q_{sh}(z) & \quad \text{total shelf outflow to the interior Arctic Ocean.} \\
q_{sh}(z) & \quad \text{shelf outflow at a given depth z.} \\
q_{sh,ml} & \quad \text{shelf outflow to the interior Arctic Ocean mixed layer.} \\
Q_{uAIW} & \quad \text{upper Arctic Intermediate Water inflow to the Greenland Sea.} \\
q_{sh} & \quad \text{shelf outflow at salinity equal to shelf water salinity.} \\
S_{atl} & \quad \text{salinity of the Atlantic layer in the Arctic Ocean model.} \\
S_{ml} & \quad \text{mixed layer salinity.} \\
S_{PW} & \quad \text{Polar Water salinity.} \\
S_{sh} & \quad \text{salinity in the shelf box model.} \\
S_{sh,ml} & \quad \text{integrated salinity of the shelf outflow into the mixed layer.} \\
S_{sh, out}(z) & \quad \text{maximum salinity where shelf outflow function,} \ q_{sh}, \text{becomes zero.} \\
T_{0} & \quad \text{surface air temperature.} \\
T_{PW} & \quad \text{Polar Water temperature.} \\
T_{atl} & \quad \text{temperature of the Atlantic water.} \\
T_{ml} & \quad \text{mixed layer temperature.} \\
T_{sh} & \quad \text{water temperature in the shelf box model.} \\
\kappa & \quad \text{friction velocity.} \\
w & \quad \text{vertical velocity.} \\
w_{e} & \quad \text{entrainment velocity.} \\
\phi & \quad \text{total outflow of salt from the shelf box model.} \\
\chi & \quad \text{coefficient of heat advection.}
\end{align*}
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References


Paleoclimatic Analysis


