 Attribution of Projected Changes in Atmospheric Moisture Transport in the Arctic:
A Self-Organizing Map Perspective

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ABSTRACT

Meridional moisture transport into the Arctic derived from one simulation of the National Center for Atmospheric Research Community Climate System Model (CCSM3), spanning the periods of 1960–99, 2010–30, and 2070–89, is analyzed. The twenty-first-century simulation incorporates the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emission Scenarios (SRES) A2 scenario for CO₂ and sulfate emissions. Modeled and observed [from the 40-yr ECMWF Re-Analysis (ERA-40)] sea level pressure (SLP) fields are classified using a neural network technique called self-organizing maps to distill a set of characteristic atmospheric circulation patterns over the region north of 60°N. Model performance is validated for the twentieth century by comparing the frequencies of occurrence of particular circulation regimes in the model to those from the ERA-40. The model successfully captures dominant SLP patterns, but differs from observations in the frequency with which certain patterns occur. The model’s twentieth-century vertical mean moisture transport profile across 70°N compares well in terms of structure but exceeds the observations by about 12% overall. By relating moisture transport to a particular circulation regime, future changes in moisture transport across 70°N are assessed and attributed to changes in frequency with which the atmosphere resides in particular SLP patterns and/or to other factors, such as changes in the meridional moisture gradient. By the late twenty-first century, the transport is projected to increase by about 21% in this model realization, with the largest contribution (32%) to the total change occurring in summer. Only about one-quarter of the annual increase is due to changes in pattern occupancy, suggesting that the majority is related to mainly thermodynamic factors. A larger poleward moisture transport likely constitutes a positive feedback on the system through related increases in latent heat release and the emission of longwave radiation to the surface.

1. Introduction

The Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC; Solomon et al. 2007) underscores previous striking and disturbing findings: the Arctic system appears to be heading toward a new state, and there are no apparent feedbacks within the Arctic that can arrest the cohesive change (Ferguson et al. 2004; Overpeck et al. 2005; Serreze and Francis 2006). Although there is a great deal of uncertainty related to feedbacks in the Arctic system, particularly involving cloud changes, it appears that the overwhelming majority of them are positive, that is, they act to enhance changes (Serreze and Barry 2005). The most often cited of these are the ice/snow albedo feedback and water vapor feedback. The basic process in the former is that as high-latitude temperatures increase, additional sea ice and snow will melt, which will expose dark ocean and land surfaces that are more effective absorbers of solar radiation. This will increase the warming,
leading to further melt of snow and ice, and hence further warming. Because most of the Arctic surface is covered with ice and snow during spring, the effects will be most pronounced in high latitudes (Hartmann 1994). The water vapor feedback involves an increase in precipitable water in the atmosphere as warming raises the saturation vapor pressure. Increased precipitable water enhances the emissivity of the atmosphere, which tends to increase the longwave flux to the surface, particularly in dry atmospheres like the polar regions (Soden and Held 2006). While these feedbacks seem straightforward, the processes by which energy is sequestered in the system until the following year are not well understood. There are only a few known or suspected negative feedbacks within the Arctic system. The aerosol dehydration feedback (Blanchet and Girard 1995) is a possibility, but it is likely too weak to have a discernible effect. Slowing of the thermohaline circulation in response to increased freshwater export to the North Atlantic is likely to have a dampening effect, but the time scales are too slow to have a substantial impact in the near term (Fichefet et al. 2003).

In this study we investigate a potentially influential but poorly understood feedback that extends beyond the Arctic and involves the horizontal transport of moist static energy (sensible heat, latent heat, and geopotential energy) from low to high latitudes by the atmosphere. In the present climate, the moist static energy transport supplies approximately 98% of the energy annually lost to space by the Arctic north of 70°N (Nakamura and Oort 1988). Because the Arctic warms more than lower latitudes, one expects that the lower-tropospheric meridional temperature gradient should relax, poleward advection of sensible heat should decrease, and Arctic warming should weaken. Simulations with global climate models support this reasoning, but they also suggest that increases in moisture transport will more than compensate for the reduction in sensible heat transport (S. Vavrus 2007, personal communication). Recent analyses by Graversen et al. (2008), and earlier work by Alexeev et al. (2005) and Held and Soden (2006), support the notion that meridional energy transport may enhance Arctic warming. Other studies reveal possible linkages between high-latitude atmospheric circulation and tropical surface temperatures (e.g., Cassou and Terray 2001; Hoerling et al. 2004; Hurrell et al. 2004), but connections between tropical variations and poleward transport of moisture are unclear. It is also suggested that changing energy transport, perhaps in magnitude and/or in spatial distribution, may be partly responsible for observed reductions in Arctic sea ice extent (Rigor and Wallace 2004), increases in surface temperature (Comiso 2003), lengthening of the melt season (Belchansky et al. 2004), loss of permafrost (Osterkamp and Romanovsky 1999), and increases in river runoff (Peterson et al. 2002).

This study explores how meridional moisture flux across 70°N changes in a single run of the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3 (CCSM3) over the next 100 yr, forced with continually increasing anthropogenic greenhouse gases. We apply a neural network technique called self-organizing maps (SOMs; Kohonen 2001) for this analysis because it distills voluminous fields of gridded values into representative, fundamental clusters organized in a matrix of 2D fields—geographic maps in this case—that are expressed in a visual and intuitive rendering. The maps are situated in the matrix relative to one another according to their similarity. In this application, the fields of data are sea level pressure (SLP) anomalies north of 60°N from both reanalyses and from the CCSM3. Each pattern in the SOM matrix is readily identifiable as having typical atmospheric features in a region, and inferences can be made about the weather generally associated with those features. The SOM can also be used to analyze other related variables, which is the approach taken in this study to assess patterns of moisture flux and moisture convergence (Skific et al. 2009), and ultimately to ascertain the causes of change in these variables in the future.

The datasets used and the data manipulation that precedes the actual application of the SOM algorithm are described in section 2, while section 3 provides a more in-depth description of the SOM method. Analysis of self-organizing maps of SLP and model validation using the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis is summarized in section 4. The analysis of the corresponding clusters of moisture transport across 70°N, fixed for a particular circulation regime, and its comparison to ERA-40 is given in section 5. Future changes of moisture transport in the twenty-first century and derivation of contributions of dynamic, thermodynamic, and combined fractions of change are provided in section 6, followed by conclusions and future efforts in section 7.

2. Data sources and model output

Six-hourly, multilevel fields of specific humidity and meridional wind for a single run of the NCAR CCSM3 (version 3.0, T85 L26 resolution) were obtained from the Program for Climate Model Diagnostics and Intercomparison (PCMDI), at the Lawrence Livermore National Laboratory. CCSM3 simulations have been shown to reproduce the Arctic atmospheric hydrological cycle reasonably well (e.g., Holland et al. 2007; Finnis et al. 2009a,b). The atmospheric component of the model...
consists of 26 vertical levels, with a top at 2.2 hPa, 13 layers above 200 hPa, and a horizontal resolution of about 1.4°. The atmospheric module is the Community Atmosphere Model (CAM) version 3.0 (Collins et al. 2006). The twentieth-century experiment (20C3M) incorporates the direct effect of sulfates (Smith et al. 2001, 2004), with no indirect aerosol effects. The model is forced by observed concentrations of CO₂, CH₄, N₂O, chlorofluorocarbons (CFCs), ozone (Kiehl et al. 1999), and solar fluxes (Lean et al. 2002). The effects of volcanic eruptions are parameterized (Ammann et al. 2003). The twenty-first-century simulation incorporates the Special Report on Emission Scenarios (SRES) A2 scenario (Nakicenovic and Swart 2000), which assumes a continuously increasing population (15 billion by 2100), increasing greenhouse gases, and slow implementation of new technologies, and appears to be most similar to the trajectory of the real world (Rahmstorf et al. 2007; Pielke et al. 2008).

The original 6-hourly PCMDI fields were interpolated from the hybrid sigma-pressure vertical coordinates to a pressure coordinate system and reduced in size by subsetting from global coverage to the region north of 60°N, from 6-hourly time resolution to daily resolution (1200 UTC only) and from 26 vertical levels to 10 levels (troposphere only). Moisture transport was calculated for five tropospheric layers (1000–850, 850–700, 700–500, 500–400, and 400–300 hPa). The time slices used in this study span periods of 1960–99 from the twentieth-century experiment (20C3M), as well as 2010–30 and 2070–89 from the SRES A2 scenario. The latter two periods were chosen to be consistent with the Arctic Climate Assessment Report (Huntington and Fox 2005; Serreze and Francis 2006) to represent the so-called emerging and mature greenhouse states. The SLP fields are also extracted for the same time periods and interpolated from the original 1.4° × 1.4° grid to a 200 km × 200 km Equal Area Scalable Earth (EASE) grid (Armstrong et al. 1997), covering the area north of 60°N and consisting of 51 × 51 grid points. (The interpolation code was obtained online from http://nsidc.org/data/ease/.) Interpolation to an equal area grid avoids errors that might occur because of unequal weighting of the original latitude–longitude grid boxes in the self-organizing map algorithm, described in the next section.

Daily SLP fields from ERA-40 (Uppala et al. 2005) were used to validate the twentieth-century CCSM3 simulation. The fields from 1958 to 2001 were also interpolated to the same EASE grid prior to applying the SOM algorithm.

3. SOM methodology

SOMs provide a means to visualize the complex distribution of synoptic states (Hewitson and Crane 2002). This technique includes an unsupervised learning algorithm to reduce the dimension of large datasets by grouping similar multidimensional fields together and organizing them into a two-dimensional array (Kohonen 2001). In this study, the high-dimensional data subjected to SOM analysis are fields of daily SLP anomalies from CCSM3 for three time slices and ERA-40 on a 51 × 51 EASE grid over the region north of 60°N.

The SOM consists of a 2D grid of nodes. Each node i corresponds to an n-dimensional weight or reference vector \( \mathbf{m}_i \), where n is the dimension of the input data, treated as a vector created from the grid points in each sample. The initial step of this routine is the creation of a first-guess array, which consists of an arbitrary number of nodes and corresponding reference vectors. In this study we use a grid of 35 nodes, creating a 7 × 5 array. Slightly smaller and larger SOM matrices were tested to determine a suitable number of nodes for this analysis. If the matrix is too small, some characteristic atmospheric patterns may not be represented; if it is too big, adjacent patterns will be too similar and visualization is unwieldy. The 7 × 5 matrix appears to capture and separate the important differences in pressure patterns. Moreover, the results are not affected by small differences in the matrix size. The reference vectors are created at the beginning using linear initialization, which consists of first determining the two eigenvectors with the largest eigenvalues, and then letting these eigenvectors span the two-dimensional linear subspace (Kohonen 2001). We use the covariance matrix of the input SLP dataset to determine the two eigenvectors. In this case the centroid of a rectangular array of initial reference vectors identified with array points corresponds to the mean of the sea level pressure values, and the vectors identified with the corners of the array correspond to the largest eigenvalues. By initiating a SOM in this way, the procedure starts with an already ordered set of weights, and then training begins with the convergence phase. Linear initialization helps achieve faster convergence, which is an advantage of this procedure over other methods, but the SOM results are not sensitive to the selected initialization method. In the process of training, each data sample (i.e., one daily map of SLP) is presented to the SOM in the order that it occurs in the original dataset. The similarity between the data sample and each of the reference vectors is then calculated, usually as a measure of Euclidean distance in space. In this process, the “best match” node is identified as that with the smallest Euclidean distance between its reference vector and the data sample. Only the vectors for the best-matching node and those that are topologically close to it in the two-dimensional array are updated. The updating scheme is shown below:
where \( t \) is a discrete time coordinate, \( \mathbf{m}_i \) is a reference vector, \( \mathbf{x} \) is a data sample, and \( h_{ci} \) is a neighborhood function (Kohonen 2001), usually in the form of the Gaussian function,

\[
h_{ci} = \alpha(t) \exp\left(-\frac{||\mathbf{r}_c - \mathbf{r}_i||^2}{2\sigma^2(t)}\right),
\]

where \( \alpha \) is the training rate function (usually an inverse function of time), \( \mathbf{r} \) is the location vector in the matrix, the distance \( ||\mathbf{r}_c - \mathbf{r}_i|| \) corresponds to the distance between the best-matching node (location \( \mathbf{r}_c \)) and each of the other nodes (location \( \mathbf{r}_i \)) in the two-dimensional matrix, and \( \sigma \) defines the width of the kernel, or a relative distance between nodes, often referred to as the radius of training. The training procedure is controlled by the training rate \( \alpha \), the training radius \( \sigma \), and the duration of training, which is fixed at 20 times the number of data samples. The initial value of \( \sigma \) is 4, and decreases linearly in time. The training scheme is repeated several times, with the training rate reduced by an order of magnitude each time. At the end of each trial the mean quantization error is calculated, defined as

\[
\text{mqe} = \frac{\sum_{i=1}^{M} (\mathbf{x}_i - \mathbf{m}_c)^2}{M},
\]

where \( \mathbf{x}_i \) is a data sample, \( M \) is the number of samples, and \( \mathbf{m}_c \) is its best-matching unit out of 35 reference vectors. A smaller mean quantization error indicates a closer resemblance between \( \mathbf{m}_c \) and the daily SLP anomaly fields. The training is complete once the smallest mean quantization error is identified, because the reference vectors from that training best approximate the data space of interest. The final reference vectors are then mapped onto a 2D grid, with their locations in the matrix corresponding to their matching nodes. The maps in the resulting matrix represent the predominant patterns in which the atmosphere tends to reside, or alternatively the centroid of the particular data cluster.

Although the measure of similarity between the data and the reference vector is linear, it is this iterative training procedure that allows the SOM to account for the nonlinear data distributions (Hewitson and Crane 2002). The nonlinear approximation of the data space is therefore a great advantage of the method compared to some other approaches, such as empirical orthogonal functions (EOFs; Reusch et al. 2005).

4. SOM of CCSM3 and ERA-40 Arctic sea level pressure fields

a. Twentieth-century analysis

Daily sea level pressure fields from ERA-40 (1958–2001) and CCSM3 for periods of 1960–99, 2010–30, and 2070–89 for the region north of 60°N are used to create the master SOM. Daily SLP anomalies are derived by subtracting the gridpoint SLP from the domain-averaged SLP for each daily field (Cassano et al. 2007). The spatial distribution of the daily SLP anomalies represent the SLP gradient, and thus the circulation, but are not influenced by the absolute SLP values. Areas with elevation higher than 500 m are removed from the fields because pressure reduction to sea level can lead to unrealistic singularities emerging in the SOM training, which then obscure the realistic patterns.

Once a SOM of sea level pressure has been created from the combined sets of both ERA-40 and CCSM3 SLP anomalies (hereafter the master SOM), all daily SLP anomaly fields may be mapped to the best-matching pattern in the master SOM to form clusters of daily maps. This is achieved by finding the trained reference vector associated with a node that minimizes the Euclidean distance, or the squared difference, between itself and the data sample. Once all of the samples have been assigned to a node in the SOM, the frequencies of occurrence can be determined, that is, the fraction of daily fields that reside in each cluster.

Ascribing a particular daily SLP sample to a specific circulation pattern in the SOM can also be useful for analyzing associated variables for the same days as those in each cluster. By mapping the new variable onto a particular SLP pattern, the new SOM representation can be used to describe the conditions associated with a specific circulation regime. In section 5 we apply this approach to fields of moisture flux across 70°N.

Figure 1 presents the master SOM for SLP anomalies north of 60°N, which is derived using the combined ERA-40 and CCSM3 SLP daily fields. These are the dominant circulation patterns in which the atmosphere tends to reside, according to these datasets. In the bottom-right are patterns with a strong Icelandic low and a moderate-to-strong Aleutian low, with high pressure over the northern Eurasian continent. The upper-right part of the map is dominated by pronounced low pressure in the Atlantic sector extending into Barents Sea, while the western central Arctic, continental regions, and the Pacific sector are dominated by high pressure. These patterns represent a moderate or strong Beaufort high in the winter. The bottom-left corner of the SOM is characterized by a pronounced low pressure area in the central Arctic with high pressure over northwestern
Eurasia. Toward the upper-left corner, a center of low pressure is located in the Kara and Laptev Seas, while high pressure is located over northeast North America. The dominant feature in patterns near the middle of the SOM is high pressure in the central Arctic.

The frequency of occurrence of winter [December–February (DJF)] patterns in CCSM3 is presented in Fig. 2a. Frequencies are expressed as a percent of days out of the total number of days in the dataset that belong to a particular cluster. The highest winter frequencies occur in patterns on the right-hand side of the master SOM. Circulation patterns on the left-hand side are more characteristic of summer [June–August (JJA)] conditions (Fig. 2b).

The frequencies of occurrence of twentieth-century SLP patterns for the CCSM3 and ERA-40 are compared in Fig. 3. The bordering clusters characterized by more pronounced SLP gradients occur more frequently in the model than in the ERA-40 fields (Fig. 3a), while those positioned in the middle of the SOM occur less frequently. The distribution of frequencies of occurrence for the model fields does not reveal a preponderance of any particular group of clusters, while the real atmosphere (Fig. 3b) exhibits a preference for the regimes in the upper-middle and middle patterns in the SOM. These patterns tend to occur more often in summer, but not exclusively so.

The SLP anomalies averaged for an area north of 75°N for each individual cluster are presented in Fig. 4. The patterns in the middle, characterized by more pronounced high pressure over the central Arctic, also have higher SLP anomalies. A comparison of the mean SLP climatologies (Figs. 5a,b) confirms the conclusions resulting from the differences in the ERA-40 and CCSM3 frequencies of occurrence shown in Fig. 3. Compared to the ERA-40, mean model SLP anomalies (relative to the Arctic-mean SLP for each day) are lower in the central Arctic and in the Atlantic sector, along with higher pressure in northern Eurasia and the northeastern American continent.

The differences in the distribution frequencies and SLP climatologies stem from the fact that the comparison is made between two single realizations of the twentieth-century climate. DeWeaver and Bitz (2006)
analyzed the simulated Arctic atmospheric circulation in CCSM3 and found biases similar to those identified here, that is, SLP in the Arctic is too low. They also found that the modeled wintertime Beaufort high is too weak. Using a single model run clearly introduces uncertainty in future climate projections and attributions owing to natural variability, model physics and numerics, and uncertainty in emission projections. Nevertheless, as

![Fig. 2. Frequency of occurrence of (a) winter (DJF) days and (b) summer (JJA) days in CCSM3. Frequencies are presented as percent of total days of the 1960–99 period that map into each node of the master SOM.](image)

![Fig. 3. Frequency of occurrence of sea level pressure anomaly patterns in the period from (a) CCSM3 during 1960–99 and (b) ERA-40 during 1958–2001. Frequencies show percent of days out of the total that map to each node in the master SOM.](image)
demonstrated in section 5, despite the differences between the modeled and observed SLP, this model run realistically simulates twentieth-century moisture flux across 70°N. In addition, Cassano et al. (2007) analyzed 15 different IPCC AR4 models, including several realizations from CCSM, and they found that the CCSM3 was one of the most realistic models in terms of adequately reproducing the observed Arctic hydrologic cycle. An assessment of the IPCC AR4 set of models by Chapman and Walsh (2007) also identifies the CCSM3 model as one of the most realistic in simulating Arctic atmospheric behavior. Because the primary focus of this study is the changes projected from the twentieth to the twenty-first century, the absolute accuracy of the simulated circulation patterns are not of central importance.

b. Twenty-first-century analysis

In this section we compare frequencies of occurrence for the distant future (2070–89) model simulations with those of the twentieth century to investigate how circulation patterns are projected to change in the future (Fig. 6). Black solid (dashed) contours show areas of significantly (>95% confidence) higher (lower) difference in frequency of occurrence. The range in a 95% confidence interval is

\[
\pm 1.96 \sqrt{\frac{p_1(1-p_1)}{n_1} + \frac{p_2(1-p_2)}{n_2}},
\]

where \(p_1(1-p_1)/n_1\) and \(p_2(1-p_2)/n_2\) are the variances of two independent, random, binomial processes, \(p_1\) and \(p_2\) are the expected frequencies of occurrence for the two time periods \(p = 1/35\), \(n_1\) is the number of samples in the first dataset, and \(n_2\) is the number of samples in the second dataset (for more details see Cassano et al. 2007). Because this statistical test does not account for the effects of serial correlation in the daily SLP fields, and thus likely overestimates the degrees of freedom, we determine an approximation for the effective degrees of freedom by dividing the number of samples of the two datasets by 7. This value is determined from the

![Figure 4](image1.png)

**Fig. 4.** Mean sea level pressure anomalies (hPa) averaged over the area north of 75°N for all days mapped to each SOM node. All data used to create the SOM are included.

![Figure 5](image2.png)

**Fig. 5.** Mean climatology of daily sea level pressure anomalies (hPa): (a) CCSM3 during 1960–99 and (b) ERA-40 during 1958–2001.
serial correlation of the SLP time series (not shown), which indicates that the atmosphere tends to reside in a circulation regime for about 1 week. This procedure decreases the degrees of freedom, thus establishing a higher threshold for determination of a significance level.

A pronounced, statistically significant increase is apparent in patterns with low pressure over the central Arctic (left of the master SOM; see Figs. 1 and 4), as well as those with strong high pressure across the western Arctic region and strong low pressure in the Atlantic sector and eastern Arctic (upper right of Fig. 1). The clusters in the middle, which are mostly dominated by a weak or a moderate high pressure over the central Arctic, decrease in frequency. Taken together these changes represent a decrease in pressure over the central Arctic in this greenhouse gas–forced model projection. Indeed, differences in the mean SLP anomalies between 2070–89 and 1960–99 (Fig. 7) indicate reductions in SLP in the central Arctic, along with increases in SLP in the North Atlantic and Pacific Oceans in the late twenty-first century. This tendency suggests that the Arctic Oscillation (Thompson and Wallace 1998) may reside in a positive phase more frequently in the future.

Cassano et al. (2007) formulated an equation that separates the factors contributing to a temporal change in a variable of interest into a portion caused by a change in the frequency of occurrence of daily maps in a cluster, a portion resulting from a change in the cluster-mean physical variable, and a third resulting from a combination of the two effects. The equation is given as follows:

$$\Delta x = \sum_{i=1}^{N} [(x_i + \Delta x_i)(f_i + \Delta f_i) - x_i f_i],$$  \hspace{1cm} (1)$$

where $\Delta x$ is the total change in a variable between two different time periods, $x_i$ is the cluster-averaged variable in the initial time period, $f_i$ is the frequency of occurrence of the daily maps in cluster $i$ during the initial period, $\Delta f_i$ is the change in cluster frequency between the two periods of interest, $\Delta x_i$ is the change in the cluster-averaged variable between the two periods of interest, and $N$ is the total number of clusters ($N = 35$ in this study). Expanding (1):

$$\Delta x = \sum_{i=1}^{N} (x_i \Delta f_i + f_i \Delta x_i + \Delta x_i \Delta f_i).$$  \hspace{1cm} (2)$$

Previously we have observed and discussed the first term, $x_i \Delta f_i$, which relates changes in the pressure field to changes in the frequency of occurrence of circulation patterns. The second term, $f_i \Delta x_i$, relates to temporal changes in the variable of interest averaged over all days that belong to a cluster. In the case of cluster-averaged SLP anomalies, the values are nonzero because they are calculated for only a portion of the entire analysis domain, the area north of 75°N. Physically, a change in the cluster-averaged SLP may result from a general intensification or weakening of high/low pressure centers without a significant change in their spatial distribution. These changes are necessarily smaller than the differences in cluster-mean values between adjacent nodes. This assertion is supported by the nearly constant quantization error in time, indicating that the atmospheric
circulation patterns represented by each cluster do not change significantly in the future, and that the change in frequency distribution for each cluster captures the important changes in atmospheric circulation.

The changes in frequency of occurrence from the twentieth to twenty-first centuries are presented in Fig. 6. For this run of the CCSM3, patterns on the far left and right of the SOM that are dominated by low pressure will increase, while those characterized by relatively high pressure over the central Arctic will decrease in frequency. Figure 7 shows the manifestation of this change in the difference in mean SLP between the end of the twenty-first century and the twentieth century. The SLP is projected to decrease over most of the Arctic Ocean and increase over Arctic lands.

The three plots in Fig. 8 show the changes in the cluster-averaged SLP anomalies [i.e., $\Delta x_i$ in Eq. (2)] north of 75°N between 2010–30 and 1960–99, 2070–89 and 2010–30, and 2070–89 and 1960–99. These changes in SLP anomalies occur for the fixed circulation regimes of the SOM. The differences in cluster-averaged SLP anomalies generally reveal decreased pressure over the central Arctic, consistent with results from other GCMs (Chapman and Walsh 2007), which contributes significantly to the total change.

5. Moisture fluxes across 70°N: Twentieth-century analysis

The flux of moisture across the imaginary wall along 70°N latitude is calculated for each daily field. The atmosphere is divided into the following five tropospheric layers: 300–400, 400–500, 500–700, 700–850, and 850–1000 hPa. The flux in each layer is determined as

$$ Q = \frac{1}{g} \int_{p_1}^{p_2} \bar{v} q \, dp. $$

Here, $Q$ is the precipitable water in the layer (kg m$^{-2}$), $q$ is specific humidity, $\bar{v}$ is mean layer meridional wind, $g$ is gravity (9.8 m s$^{-2}$), and $dp$ is the pressure differential in a layer between pressure levels $p_1$ and $p_2$. The moisture flux expressed in this way has the units of kilogram per meter–second. The total flux across the wall at 70°N is obtained by integrating around the latitude circle at
each level and multiplying by the latent heat of evaporation ($L = 2.5 \times 10^3 \text{ J kg}^{-1}$). The values at each level are then summed vertically to obtain the total moisture transport into the Arctic $F_q$, which has units of watts. It is common to express this value as a flux per unit area of the Arctic region, so the flux is divided by the area of the polar cap north of 70°N to obtain units of watts per square meter.

The twentieth-century mean, zonally averaged moisture flux profile $L(tq)$ across 70°N for CCSM3 (blue line) and ERA-40 (green line) is presented in Fig. 9. Greenland has been excluded from the calculations. The model reproduces the observed moisture flux remarkably well. The largest differences occur in the lower levels, which results in the model overestimating the twentieth-century moisture flux across 70°N by about 12%. The larger modeled fluxes in the lower levels are likely related to the generally lower Arctic SLP in the model than in ERA-40. Above 850 hPa, however, the model profile is nearly indistinguishable from the observed values.

Because each day of ERA-40 and model output can be ascribed to one of the SOM clusters, we can analyze poleward moisture fluxes corresponding to each atmospheric pattern in the SOM matrix. The cluster-averaged values of the flux across 70°N are presented in Fig. 10 (Greenland has been excluded). Red areas in Fig. 10 correspond to patterns in the master SOM characterized by strong Icelandic lows and by pronounced low pressure over the central Arctic, both of which tend to advect large quantities of moisture poleward. Figure 11 presents a height–longitude view of moisture flux across 70°N mapped onto the circulation patterns in the master SOM [kg (m s)$^{-1}$]. The strongest moisture transport occurs in the lower and the middle troposphere, where the specific humidity is typically largest. Strong positive (poleward) moisture fluxes (red shading) are located east of the low pressure centers and west of the high pressure regions in the corresponding clusters in the master SOM. Correspondingly, negative equatorward moisture fluxes (blue areas) are found west of the low pressure systems and east of the high pressure features. Features in the lower-right portion of the SOM generate the highest moisture transport. These patterns correspond to a positive North Atlantic Oscillation (NAO) index (Hurrell et al. 2003), with a pronounced Icelandic low in the Atlantic sector and high pressure over the Eurasian continent. The circulation patterns in the center of the SOM, related to weak or moderate high pressure over the central Arctic, generally show the weakest northward moisture transport. High pressure over the central Arctic, usually centered in the Beaufort Sea, generally indicates divergent flow and a stronger southward branch of moisture transport across the Canadian Arctic archipelago. Strong, low pressure over the central Arctic favors convergence and increased northward moisture transport (patterns in the lower-left corner).

Seasonal-mean moisture fluxes across 70°N for the model’s twentieth century are presented in Fig. 12. Fluxes are strongest in summer primarily because of the increased depth of the moist layer and generally more poleward wind vectors (Groves and Francis 2002). A deeper moist layer also allows the stronger upper-level winds to advect more moisture during summer. In winter fluxes are the weakest, because precipitable water values are lower, and the moist layer is shallow under a strong surface-based inversion. The strongest fluxes in all seasons
FIG. 11. Longitude–height representation of the moisture flux in five layers across 70°N (kg m⁻¹ s⁻¹) for CCSM3 mapped onto the corresponding SOM circulation patterns shown in Fig. 1. Red (blue) shaded areas denote northward (southward) fluxes. Vertical layers are bounded by 1000, 850, 700, 500, 400, and 300 hPa.
occur in the Atlantic sector, because this is the main pathway of moisture entering the Arctic. Temperature gradients are typically strongest in this area, and transports are driven by the primary storm track, that is, the Icelandic low. Baroclinicity in this sector is associated with strong horizontal temperature gradients sharpened by coastal orography (Serreze and Barry 2005). Katabatic winds associated with the high Greenland plateau increase the temperature contrast in the region and help sustain a strong Atlantic baroclinic zone (Serreze and Barry 2005). In summer the meridional temperature gradient relaxes, but cyclonic activity is still high in the Arctic. Cyclones penetrate farther northward, with baroclinicity sustained by differential heating between the Arctic Ocean and snow-free land, and intensified by the coastal orography. Weak or moderate southward moisture flux occurs between 120° and 60°W. This area west of Greenland and across the Canadian Archipelago represents the main exiting branch of moisture out of the Arctic. The area of strong southward moisture flux just east of Greenland is most likely the result of high orography channeling the return flow of the Atlantic low pressure systems.

6. Attribution of changing moisture flux across 70°N in the twenty-first century

We now apply the same principle described in section 4b to provide insight into the causes for changes in the moisture flux across 70°N from the twentieth century to the late twenty-first century. The first term in Eq. (2) represents the portion of the total change owing to shifts in the frequencies with which daily SLP fields reside in the patterns depicted in the SOM. A change in this distribution represents a change in the surface circulation, and thus we loosely refer to this contribution as the dynamic factor. The second term in Eq. (2) captures the fraction of the total change that is due to a change in the cluster-averaged value of the parameter of interest for each fixed SLP pattern. In the case of moisture flux, changes of this type are likely caused mainly by thermodynamic effects, such as varying moisture gradients and the vertical distribution of water vapor; thus, we refer to this contribution as the thermodynamic factor. Clearly there could be dynamic influences affecting this term as well, but they are expected to be of secondary importance. The third term in Eq. (2) represents the contribution from the interaction of both changing pattern frequency and the cluster-averaged variable. This term tends to be small.

The attribution of change is evaluated annually and seasonally between the three time slices (between 2010–30 and 1960–99, 2070–89 and 2010–30, and 2070–89 and 1960–99). It should be noted that these results are meant to provide insight into the causes of future change, not a quantitative accounting. The dynamic factor $x_i \Delta f_i$ is obtained by multiplying the initial cluster-mean moisture flux values (shown in Fig. 10) by the change in the...
pattern frequency between each time slice, an example of which is shown in Fig. 6. (The results for this contribution are shown in the top panels of Figs. 14–16.) The thermodynamic factor $f_i \Delta x_i$ is obtained by combining the initial distribution of frequencies of occurrence shown in Fig. 3a with the changes in cluster-mean moisture flux $\Delta x_i$, which is presented in Fig. 13 for each time slice. The thermodynamic factor is shown in the middle panels of Figs. 14–16. Finally, the combined term is $\Delta x_i$ multiplied by $\Delta f_i$ and is plotted in the bottom panels. To obtain the annual-mean contribution to the total change, the values for each contribution are summed over all of the clusters. The numerical results are summarized in Table 1 along with a breakdown by season. It should be noted that these results are not sensitive to the size of the SOM used in this analysis. As shown in Table 2, the annual-mean contributions to the total change by the three terms are virtually unchanged for varying sizes of the SOM matrix. The $7 \times 5$ array chosen for this study represents a balance between having enough nodes to capture the representative patterns in the Arctic atmosphere and the ability to present those patterns graphically.

In all three time comparisons, the thermodynamic factor clearly plays the most important role in the changing moisture transport in this model simulation, which is in turn driven by the change in the cluster-mean flux (Fig. 13). This quantity exhibits a positive temporal change in every node. Because the model is forced by a realistic estimate of increasing greenhouse gas composition through the twenty-first century, the most logical explanation for this result is the forcing. From the late

![Fig. 13. Temporal differences in the cluster-averaged moisture flux (W m$^{-2}$) from CCSM3 across 70°N: (a) 2010–30 minus 1960–99, (b) 2070–89 minus 2010–30, and (c) 2070–89 minus 1960–99.](image-url)
twentieth century to the near future, we find that over 80% of the increase in total northward moisture flux is due to the thermodynamic factor. A larger increase in the moisture transport of about 2.5 W m\(^{-2}\) (16%) occurs later in the twenty-first century, from 2010–30 to 2070–89, compared to approximately 0.9 W m\(^{-2}\) (5%) from 1960–99 to 2010–30. This results from the substantial warming, and the consequential increased water vapor content, by the end of the century. The increase in the later twenty-first century is also dominated by thermodynamic effects.

The dynamic term plays only a secondary role in driving projected moisture flux changes. As discussed in section 4, low pressure anomalies over the central Arctic occur more frequently in the future in this simulation, while features showing weak or moderate high pressure occur less often. This also favors an increase in poleward moisture flux, but only explains about 16% of the total projected change. The combined term contributes least to the total change, but its influence increases in the far future (2070–89), accounting for about 8% of the total change.

The seasonal analysis also reveals the dominant contribution by the thermodynamic factor in governing increases in moisture flux by the twenty-first-century CCSM3 (Table 1). The largest increases in moisture flux occur in summer: 0.3 W m\(^{-2}\) (7%) between the near future and the twentieth century and 0.8 W m\(^{-2}\) (19%) between the near future and the far future, with a total of 1.1 W m\(^{-2}\) (26%) from the present to the far future. Summer increase in moisture transport accounts for about 32% of the total change by the late twenty-first century.
In summer the atmospheric moisture content, as well as the saturation vapor pressure, achieve their highest values. The increased meridional gradient in specific humidity can be explained by the exponential relationship between saturation vapor pressure and temperature. At the warmer low-latitude temperatures, a given increase in temperature will lead to a relatively larger increase in saturation vapor pressure than that at higher latitudes for the same warming, which likely results in an increased specific humidity gradient. High-latitude meridional temperature gradients are also sustained by land–ocean differential heating. While Arctic land areas warm as the snow and ice cover melts away, the Arctic Ocean’s temperature is confined near the melting point. These factors contribute to the largest seasonal increase in poleward moisture fluxes. The thermodynamic factor plays a dominant role in the summer, explaining about 79% of the moisture increase between the near future and the twentieth century in this simulation, while the other effects account for a larger portion of the change between the near and far future. This can be explained by the increased frequency of occurrence in summer circulation patterns that favor convergence and thus an increase in the poleward moisture flux.

In spring the thermodynamic term dominates in all time intervals. Between the twentieth century and the near future the flux increases about 0.3 W m\(^{-2}\) (8%), and between the near and far future the increase is 0.4 W m\(^{-2}\) (11%), resulting in a total change of 0.7 W m\(^{-2}\) (19%) between the twentieth century and the far future. In the later period the role of the dynamic term decreases dramatically, resulting in the thermodynamic factor accounting for virtually all of the change in the moisture flux occurring between the near and far future. The spring increase in moisture transport accounts for about 23% of the total change by the late twenty-first century.

**FIG. 15.** As in Fig. 14, but for the period from 2010–30 to 2070–89.
The change in the moisture flux in winter is small between the twentieth century and the near future (approximately 0.1 W m$^{-2}$; 4%), but increases between the near and far future (0.7 W m$^{-2}$; 20%), resulting in a total change of 0.8 W m$^{-2}$ (24%) between the twentieth century and the late twenty-first century. The winter increase in moisture transport by the late twenty-first century accounts for 24% of the total change. While the thermodynamic term plays the leading role between the near and far future (accounting for 91% of the total change), between the twentieth century and the near future the dynamic term dominates. In the early period dynamics explains about 46% of the total change in moisture flux, with the combined term playing a secondary role, accounting for 36%. The thermodynamic term accounts for only about 18% of the total change. This may be explained by a modest increase in frequency of occurrence of the winter patterns in the right of the SOM. Thermodynamic influences, mainly driven by the increased moisture content of the atmosphere, may be hampered by weakened temperature and humidity gradients that are expected in the near future (Serreze and Francis 2006). Delayed ice growth and the resulting thinner ice cover reduce the insulating effect of the sea ice. Because the important source of atmospheric heating of the polar atmosphere in cold seasons is the ocean, its effect may, in the emerging warming state, weaken the poleward moisture transport. Between the near- and far-future projections, the contributions by the dynamic and combined terms decrease, and the thermodynamic factor dominates. Weakening of winter dynamics in the late future can be explained by a decrease in the frequency of occurrence of the winter patterns in the lower-right corner, corresponding to positive NAO-like patterns that generally bring more moisture into the Arctic (Fig. 6).
The changes in autumn appear to be similar to those for winter, albeit smaller. Moisture flux increases by approximately 0.1 W m\(^{-2}\) (2\%) between the twentieth century and the near future and 0.6 W m\(^{-2}\) (14\%) between the near and far future, resulting in a total change of 0.7 W m\(^{-2}\) (16\%) between the twentieth century and the late twenty-first century. The autumn change represents about 21\% of the annual-mean change. In the early period the largest contribution is from the dynamic term, which accounts for about 59\% of the total change, followed by the combined term at 45\%. Interestingly, the thermodynamic term is negative, acting to weaken the poleward fluxes, which may be explained by the strong influence of decreased temperature and moisture gradients across 70°N as the sea ice declines. This explanation is consistent with Serreze and Francis (2006), who found maximum autumn warming over the Arctic Ocean in four of five climate models analyzed, which reduced the meridional gradients. Later in the twenty-first century, thermodynamic effects dominate, accounting for 54\% of the total change, followed by dynamics (28\%) and, finally, the combined influence (18\%). The general increase in the depth of the moist layer in the late twenty-first century during autumn allows stronger upper-level winds to play a greater role in moisture transport.

### 7. Discussion and conclusions

Changes in poleward moisture fluxes across 70°N in the twenty-first century and their relationships to surface circulation patterns have been analyzed using daily output from the NCAR CCSM3 and a neural network technique called self-organizing maps (SOMs). The study focuses on changes in the moisture flux that are projected for the early (2010–30) and late (2070–89) twenty-first century, compared to the 1960–99 period, based on a single model realization forced by the A2 SRES scenario. Comparison of the modeled moisture fluxes for the twentieth century with those derived from the ERA-40 indicates that the model realistically simulates the moisture flux across a wall at 70°N within about 12\%. Based on a partitioning of the change into its components for fixed and varying circulation regimes, we find that approximately 75\% of the change in moisture transport during the twenty-first century is attributed to factors related primarily to thermodynamic factors rather than changing circulation patterns. Poleward moisture flux across 70°N is projected to increase by approximately 0.9 W m\(^{-2}\) (5\%) between 1960–99 and 2010–30, and by about 2.5 W m\(^{-2}\) (16\%) between 2010–30 and 2070–89. These values are of similar magnitude to the increase in radiative forcing by anthropogenic greenhouse gases since the 1800s. Thermodynamic factors driving the enhanced moisture flux include increases in precipitable water and the meridional moisture gradient. The depth of the near-surface moist layer will also increase, allowing the typically stronger winds aloft to play a greater role in the transport. The modeled increase in northward moisture flux implicates further

### Table 1. Contributions of dynamic, thermodynamic, and combined terms to the total change in moisture flux across 70°N (W m\(^{-2}\)), for various time frames of the twenty-first century. Changes are observed annually and seasonally. Changes between the twentieth and the late twenty-first century are bold.

<table>
<thead>
<tr>
<th>Period observed</th>
<th>Dynamic term (W m(^{-2}))</th>
<th>Thermodynamic term (W m(^{-2}))</th>
<th>Combined term (W m(^{-2}))</th>
<th>Total (W m(^{-2}))</th>
<th>Dynamic term (%)</th>
<th>Thermodynamic term (%)</th>
<th>Combined term (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2010–30 minus 1960–99</td>
<td>0.2</td>
<td>0.7</td>
<td>0.0</td>
<td>0.9</td>
<td>17.4</td>
<td>81.8</td>
<td>0.8</td>
</tr>
<tr>
<td>2070–89 minus 1960–99</td>
<td>0.5</td>
<td>2.6</td>
<td>0.3</td>
<td>3.4</td>
<td>15.5</td>
<td>76.8</td>
<td>7.7</td>
</tr>
<tr>
<td>2010–30 minus 1960–99</td>
<td>0.1</td>
<td>0.0</td>
<td>0.0</td>
<td>0.1</td>
<td>45.5</td>
<td>18.2</td>
<td>36.4</td>
</tr>
<tr>
<td>2070–89 minus 1960–99</td>
<td>0.0</td>
<td>0.7</td>
<td>0.1</td>
<td>0.8</td>
<td>4.8</td>
<td>86.8</td>
<td>8.4</td>
</tr>
<tr>
<td>2010–30 minus 1960–99</td>
<td>0.1</td>
<td>0.3</td>
<td>−0.0</td>
<td>0.3</td>
<td>22.3</td>
<td>78.6</td>
<td>−0.9</td>
</tr>
<tr>
<td>2070–89 minus 1960–99</td>
<td>0.1</td>
<td>0.7</td>
<td>−0.0</td>
<td>0.8</td>
<td>11.1</td>
<td>92.4</td>
<td>−3.5</td>
</tr>
<tr>
<td>2010–30 minus 1960–99</td>
<td>0.1</td>
<td>0.2</td>
<td>−0.0</td>
<td>0.3</td>
<td>23.7</td>
<td>79.6</td>
<td>−3.3</td>
</tr>
<tr>
<td>2070–89 minus 1960–99</td>
<td>0.3</td>
<td>0.8</td>
<td>0.1</td>
<td>1.1</td>
<td>23.6</td>
<td>68.4</td>
<td>8.0</td>
</tr>
<tr>
<td>2010–30 minus 1960–99</td>
<td>0.1</td>
<td>−0.0</td>
<td>0.0</td>
<td>0.1</td>
<td>58.8</td>
<td>−3.6</td>
<td>44.8</td>
</tr>
</tbody>
</table>

### Table 2. Contributions of dynamic, thermodynamic, and combined terms to the annual change in moisture flux across 70°N (W m\(^{-2}\)), between the late twenty-first century and the twentieth century, for various matrix sizes.

<table>
<thead>
<tr>
<th>Size of matrix</th>
<th>Dynamic</th>
<th>Thermodynamic</th>
<th>Combined</th>
</tr>
</thead>
<tbody>
<tr>
<td>20 (4 × 5)</td>
<td>0.34</td>
<td>2.68</td>
<td>0.40</td>
</tr>
<tr>
<td>30 (5 × 6)</td>
<td>0.36</td>
<td>2.66</td>
<td>0.40</td>
</tr>
<tr>
<td>35 (5 × 7)</td>
<td>0.43</td>
<td>2.63</td>
<td>0.36</td>
</tr>
<tr>
<td>40 (5 × 8)</td>
<td>0.38</td>
<td>2.67</td>
<td>0.37</td>
</tr>
<tr>
<td>48 (6 × 8)</td>
<td>0.45</td>
<td>2.65</td>
<td>0.32</td>
</tr>
<tr>
<td>56 (7 × 8)</td>
<td>0.46</td>
<td>2.65</td>
<td>0.34</td>
</tr>
<tr>
<td>80 (8 × 10)</td>
<td>0.42</td>
<td>2.65</td>
<td>0.35</td>
</tr>
<tr>
<td>120 (10 × 12)</td>
<td>0.39</td>
<td>2.68</td>
<td>0.35</td>
</tr>
<tr>
<td>240 (10 × 24)</td>
<td>0.41</td>
<td>2.63</td>
<td>0.38</td>
</tr>
</tbody>
</table>
amplification of Arctic warming, because of the release of additional latent heat through condensation of water vapor and the increased emissivity of the atmosphere. Increased warming may augment the already rapidly declining permanent Arctic ice stores. The increased moisture content of the high-latitude atmosphere, particularly in summer, may also lead to an increase in cloud cover and precipitation, thus intensifying the hydrologic cycle. Skific et al. (2009) finds that for the same model simulation, Arctic net precipitation increases by about 20% and mean cloud fraction by about 10% by the late twenty-first century. They also find that the largest increase in net precipitation over the Arctic Ocean occurs in summer, with thermodynamic factors accounting for over 70% of the increase. The summer maximum in net precipitation is related to a seasonal maximum in moisture flux convergence (Serreze and Barry 2005), and thus an increase in moisture flux in the future will likely lead to increased precipitation in the Arctic. This eventuality was demonstrated by Cassano et al. (2007), who find increases in high-latitude precipitation during the twenty-first century in ensemble output from 15 GCMs used for the IPCC AR4. They also calculate that more than 75% of that change is due to atmospheric thermodynamics. According to Liu et al. (2007) the increase in winter moisture transport during the past few decades appears to be linked with more frequent cyclones and increased cloud amount in the Arctic.

Our results indicate that moisture fluxes are projected to increase in this model simulation. The future behavior of the other two components of the moist static energy flux—geopotential and sensible heat advection—are yet to be explored in terms of their behavior relative to moisture transport, their drivers and feedbacks within the Arctic system, and their roles in Arctic amplification. Linkages between the dramatic changes within the Arctic and the global system remain poorly understood in present conditions, and thus the uncertainty regarding future changes will remain an important focus for global change research.

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