Spatial and Temporal Structure of Atmospheric Water Vapor Transport in the Mackenzie River Basin

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ABSTRACT

The transport of water vapor through the Mackenzie River basin, a typical high-latitude river basin, is examined for the period from August to October 1994. The spatial and temporal variability in the transport is considered with both objectively analyzed fields and radiosonde data.

Previous studies of the high-latitude water vapor have made use of radiosonde data and have been able to document some features of annual cycle of water vapor transport. These studies have left unresolved many important aspects of moisture transport processes. In particular, detailed information as to the spatial and temporal variation of the transport has not been fully documented or understood. In order to address these important issues, the authors make use of the objectively analyzed fields from the European Centre for Medium-Range Weather Forecasts to study the high-latitude transport of water vapor. This paper presents findings regarding the transport of water vapor over northern parts of North America. It is shown that the transport is highly variable in time with transient synoptic-scale disturbances being responsible for much of the transport. The prospect of using the objectively analyzed fields to determine the spatial structure of humidity fluxes and the evaporation–precipitation field in data-sparse high-latitude regions is discussed. The results obtained with the objectively analyzed fields are compared with those obtained directly from radiosonde data for stations in and around the basin. The influence that the local land topography has on the regional water vapor balance is also discussed.

1. Introduction

The analysis of water vapor transport and the hydrological cycle in high latitudes is a topic of interest for many researchers. Water vapor plays an important role in the energetics of the atmosphere through the release of latent heat and radiational interactions (Webster 1994; Serreze et al. 1995). It is one of the major greenhouse gases, and the Arctic appears to be the area most sensitive to greenhouse warming (Manabe et al. 1991; Hinzman and Kane 1992). In addition, the high-latitude water vapor balance has a significant influence on the composition and circulation in the Arctic Ocean (Barry et al. 1993). Through the freshwater export to the North Atlantic and its influence on the thermohaline circulation of the world ocean, the high-latitude water vapor balance also has a direct impact on the entire climate system (Aagaard and Carmack 1989).

One of the most important components in the freshwater budget of the Arctic Ocean is the discharge from the north-flowing rivers. It influences ice cover formation, ice draft to the North Atlantic, global changes of oceanic salinity, and other processes that determine the global climate. According to Aagaard and Carmack (1989), the river runoff has the greatest impact on the freshwater budget of the Arctic and exceeds the net balance by an order of magnitude. Accordingly, small (order 10%) changes in the runoff would drastically change the net fresh water balance of the Arctic Ocean. Manak and Mysak (1989) and Mysak et al. (1990) state that the fluctuations in salinity and sea ice concentration in the Arctic and the North Atlantic may have their origin in the increased discharge from the Mackenzie River. This, in its turn, may induce long-term (decades to centuries) climatic fluctuations in midlatitudes (Aagaard and Carmack 1989; Weaver and Hughes 1992; Bjornsson et al. 1995). From this point of view, the water balance of the rivers flowing into the Arctic Ocean, and particularly, the Mackenzie River basin, is of great importance.

The Mackenzie is one of the largest rivers flowing into the Arctic Ocean. It has the fourth largest runoff among the north-flowing rivers, after the three Siberian rivers, the Ob, the Yenisei, and the Lena (Aagaard and Carmack 1989). The information on the magnitude of the Mackenzie’s runoff varies. Based on Water Survey of Canada data for the period 1973–93 at Arctic Red, a station near the mouth of the Mackenzie River, the mean annual discharge is 288 km³ yr⁻¹. In contrast, the

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``widely quoted'' (Aagaard and Carmack 1989) estimate by UNESCO (1978) is 340 km$^3$ yr$^{-1}$. The reasons for this discrepancy are unclear but serve to highlight the uncertainty present in our knowledge of some of the most fundamental characteristics of the basin's water cycle.

The Mackenzie basin (Fig. 1) covers an area of 1.787 x 10$^6$ km$^2$, which is almost 20% of Canadian land mass, and includes three large lakes, the eastern slopes of the Canadian Rockies, and extensive wetlands. The typical yearly mean temperature is around 0°C; most of the ground is underlain by permafrost, and the river itself is frozen from November to June. From the viewpoint of topographic and climatic characteristics, such as the basin area/runoff ratio, the presence of mountains as well as flatlands, permanent frost in most areas, etc.; the Mackenzie basin is a typical high-latitude major river basin, similar to the basins of the Lena, the Yenisei, the Kolyma, and the Yukon. The Ob basin appears to be different, since it has almost no mountains and relatively little permafrost. The mean annual precipitation in the Mackenzie basin is approximately 410 mm and includes three large lakes, the eastern slopes of the Rocky Mountains. Evaporation within the basin is estimated to be 100–200 mm yr$^{-1}$ from land and about 400 mm yr$^{-1}$ from water surfaces.

Based on climate model simulations, the Arctic is predicted to be the area where the signature of global warming is amplified (i.e., Manabe et al. 1991). Miller and Russell (1992) predict a 15%–30% increase in the discharge from the Mackenzie would occur as a result of global warming. In addition, Jones (1988) reports that the surface air temperature in the Mackenzie basin is increasing at a rate of 0.5°C–0.75°C per decade—the second highest rate in the Arctic Ocean drainage basin after the Yenisei basin.

One of the interesting features of the Mackenzie basin is that it is bordered by the Rocky Mountains to the west, which are an obstacle to storms coming from the Pacific. The perturbations in the moisture flow that are the result of the presence of the Rocky Mountains are also of interest from the viewpoint of the Mackenzie basin water budget and are discussed in this paper.

The results by Bjornsson et al. (1995) show a high correlation between cyclogenesis over the North Pacific and the precipitation in the Mackenzie basin. It suggests that the moisture is brought to the Mackenzie basin by storms generated over the Pacific Ocean. The Beaufort and Arctic Storm Experiment (BASE) project was carried out to study these storms, and included a set of field measurements during September–October 1994. This determined the choice of time interval for the study presented in this paper.

The studies of atmospheric moisture transport are often associated with the problem of the applicability of one or another source of data (e.g., Mo and Higgins 1996; Schmitz and Mullen 1996). The question is whether to use the interpolated radiosonde data, or the objective analysis data generated by the assimilation of a variety of data into a numerical weather prediction model. Traditionally, the interpolation of radiosonde data has been considered a better way, due to imperfection of models and assimilation procedures. Recent improvements in the objective analysis techniques provide better estimates of the quantities important in hydrological balance studies (Mo and Higgins 1996). However, the accuracy of these estimates may depend on the properties of the region of interest: land or ocean, mountains or plains, etc., and this dependence is not always documented well enough.

The choice of datasets to use in such studies depends on the region of interest and the objectives of the study. In regions that have a dense station coverage or when one is interested in mean values over long time periods, it is probably preferable to use raw radiosonde data directly. For example, Serreze et al. (1995) used interpolated radiosonde data to calculate various mean characteristics of water vapor transport across the 70°N parallel, and Walsh et al. (1994) obtained monthly mean characteristics of water vapor balance in the Mackenzie basin from radiosonde data. However, for studies where radiosonde stations are sparse such as is the case of the Mackenzie basin, or when one is interested in spatial and temporal variability on short timescales, it seems appropriate to use results of objective analysis data assimilation rather than of plain interpolation of radiosonde data. Trenberth and Guillemot (1995) investigated the applicability of the analyses from the National Centers for Environmental Prediction (formerly the National
Meteorological Center) and ECMWF to the atmospheric moisture budget studies, but made no final conclusions, stating that “there is no acknowledged source of truth” for water vapor characteristics. Schmitz and Mullen (1996) used objectively analyzed fields from ECMWF to study water vapor transport into the Sonoran Desert during the monsoon period. In situations where moisture is transported not with a continuous flow (like a monsoon) but by transient weather phenomena with life terms of several days, the usage of objective analysis data seems even more justified. As we shall see, this is indeed the case for the Mackenzie basin.

In this paper, the following problems are addressed. First, the hypothesis on the major role that transient weather phenomena play in water balance of the Mackenzie basin area will be verified. As it was pointed out by Walsh and Chapman (1990), “short-term climatic fluctuations in the Arctic have received surprisingly little attention in recent years . . . .” This paper will hopefully begin to address this oversight. Second, the internal structure of moisture transport through the Mackenzie basin area from the Pacific Ocean during August–October period is studied, as well as its variation throughout this period and its relation to storm tracks in the region. The results of Bjornsson et al. (1995) concerning the Pacific Ocean being the source of moisture in the Mackenzie basin is thus demonstrated. Third, the usefulness of ECMWF data for studies similar to ours is investigated by comparison of moisture fluxes obtained from the analysis data with those computed directly from 10 stations soundings data, 5 at the Pacific coast, and 5 to the east of the Rocky Mountains.

2. Data

To analyze the water vapor fluxes, we use the ECMWF advanced surface and upper air objective analysis data (ECMWF 1992). The data are available every 6 h (0000, 0006, 1200, and 1800 UTC). We use the T106 upper air datasets for the Northern Hemisphere interpolated from 31 sigma levels to 11 standard pressure levels (1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, 100 mb) and to the horizontal grid with a uniform 1.125° spacing. Standard surface and upper air fields are included in the analysis. Various derived fields, such as precipitable water and the integrated water vapor flux, are calculated.

One of the purposes of our study is to find out how well the ECMWF analysis describes the process of moisture transport, what factors determine the accuracy of the analysis and what variables are most vulnerable to distortion. According to Trenberth and Olson (1988), wind and humidity fields are the most affected by errors in the analysis scheme, and these are exactly the quantities necessary to calculate moisture fluxes. Another possible source of error in the results obtained with the objectively analyzed data is the presence of mountains. In our case, they have heights up to the 700-mb level and have highly variable topography. The problem is the interpolation from sigma to pressure coordinates and the resultant presence of “false” data underground. With limited vertical and horizontal resolution, the surface pressure field may not always represent the details of topography correctly, and it may be impossible to totally eliminate this source of error. Another source of error arises from the data assimilation procedures used to generate the analysis. In order to determine how exactly the analysis data differ from reality, the radiosonde data and the derived functions (precipitable water, moisture flux) for six stations in Canada and four stations in Alaska are compared to the ECMWF data. To investigate the influence of the Rocky Mountains on the ECMWF data accuracy, we chose five of these stations to the west of the mountains (1–5 in Fig. 1) and five to the east (6–10 in Fig. 1).

3. Analysis procedures

The equation describing the conservation of water vapor in an air column above a particular point of the earth’s surface, if the flux across the upper boundary, diffusion and liquid and solid phase transports are neglected, is as follows:

$$P - E = -\frac{\partial W}{\partial t} - \text{div} \mathbf{Q}. \quad (1)$$

Here \(P\) is the precipitation rate, \(E\) is the evaporation rate (both at the surface), \(W\) is the precipitable water defined by

$$W = \frac{1}{g} \int_{0}^{p_{0}} q \, dp, \quad (2)$$

and \(Q\) is the vertically integrated moisture flux:

$$Q = \frac{1}{g} \int_{0}^{p_{0}} q \mathbf{v} \, dp. \quad (3)$$

In these formulas \(q\) is specific humidity, \(\mathbf{v}\) is the horizontal wind velocity, \(p\) is the pressure, \(p_{0}\) is the surface pressure. The lower limit of integration can be set to zero or to any other level in upper atmosphere above which the specific humidity can be neglected. In our calculations, the value of 100 mb was chosen.

For long time intervals (one month or more), the contribution of the \(\text{div} \mathbf{Q}\) term is more significant than that of the precipitable water tendency (Walsh et al. 1994; Serreze et al. 1995). Indeed, averaging Eq. (1) over a long period of time yields

$$\overline{P - E} = -\frac{\Delta W}{\Delta t} - \overline{\text{div} \mathbf{Q}}. \quad (4)$$

where the first term in the right-hand side obviously converges to 0 as \(\Delta t\) grows; for \(\Delta t = 1\) month, it does not exceed 25% of the mean flux divergence in Arctic regions (Serreze et al. 1995). However, at smaller \(\Delta t\), the precipitable water tendency in Eq. (4) does not need
Fig. 2. Components of water vapor balance of the Mackenzie basin [defined by Eq. (7)], given in kg s$^{-1}$, as a function of time, from 1 August. The bold solid line is $\langle P - E \rangle$, the thin solid line is $\langle \text{div} \mathbf{Q} \rangle$, and the dotted line is $\langle -\partial W/\partial t \rangle$. The bold horizontal line shows the mean value of $\langle P - E \rangle$, equal to $0.015 \times 10^9$ kg s$^{-1}$. The maxima in $\langle \text{div} \mathbf{Q} \rangle$ indicate inflows of moisture, the minima, outflows. Also indicated by arrows are the two events that are shown in Fig. 3. 

In this paper, the vector $\mathbf{Q}$, calculated every 6 h, is used as the primary source of diagnostics (vector maps, Hovmoeller diagrams, time series, etc.) To study the effect that topography to the west of the basin has on the moisture transport, we also define vectors of the moisture transport in the upper and lower level, $\mathbf{Q}_\text{hi}$ and $\mathbf{Q}_\text{lo}$, as follows:

$$\mathbf{Q}_\text{hi} = \frac{1}{g} \int_0^{700\text{mb}} qv \, dp, \quad \mathbf{Q}_\text{lo} = \frac{1}{g} \int_{700\text{mb}}^{p_0} qv \, dp. \quad (6)$$

Using the 700-mb surface as a cutoff was motivated by the fact that this is the approximate height of the barrier to the west of the basin. As with Eqs. (2) and (3), we use 100 mb as the lower limit of integration in Eq. (6).

4. Results and discussion

a. The “atmospheric river” over the Mackenzie basin

Figure 2 displays the time series of all three terms of Eq. (5) over the Mackenzie basin area, as obtained from the ECMWF data for the period 1 August–31 October.
Fig. 3. Weather events corresponding to two of the maxima in Fig. 2: 27 Aug (day 27) and 4 Oct (day 65). Contours show the geopotential height at 700 mb, arrows show vertically integrated moisture transport $Q$. Here, "H" and "L" indicate centers of high and low pressure.
FIG. 4. Biweekly means of \( \mathbf{Q} \) (arrows) and \( z_{700} \) (contours): (a) 1–15 Aug, (b) 16–31 Aug, (c) 1–15 Sep, (d) 16–30 Sep 1994. The maximum vector length is 1000 kg (m s\(^{-1}\)).
1994. The components of moisture balance are given in kg s$^{-1}$ (a mass flux convergence of $10^9$ kg s$^{-1}$ is equivalent to a basin-averaged precipitation of 2.08 mm h$^{-1}$). From Fig. 2, one can see that the moisture transport into the basin is highly variable. Strong inflows and outflows follow each other, and the mean value of $\langle P - E \rangle$ is an order of magnitude smaller than its peak values. This suggests that the water vapor balance in the Mackenzie basin is highly dependent on transient weather phenomena. One can also see that the other two terms, $\langle -\delta W / \rangle$.
shown in Fig. 3b. The biweekly mean fields for October brought into the basin by the cyclonic flow around this pulses of moisture over the Mackenzie basin are now quasi-stationary low, often observed over the Gulf of Fig. 4b along the Pacific coast line, develops into the flow also changes. The quasi-stationary trough, seen in moves southward (Figs. 4b±4d). The structure of the sure zone disappears, the filament of moisture flow created by small low pressure centers moving into the basin and results in the maximum in Q indicated in Fig. 2. In this instance, the increased moisture flow takes place between the transient low and a quasi-stationary high pressure ridge that resides over the North Pacific. Another type of event, typical for late September and October, is shown in Fig. 3b (4 October or day 65). Here the transport is associated with a synoptic-scale low over the Bering Strait. The low-level moisture is blocked by the mountains. However, a pulse of moisture associated with the cyclonic flow around the system propagates into the basin at upper levels, even though the system itself does not. The complete set of maps shows that each peak on Fig. 2 is associated with a weather event transporting moisture from the Pacific Ocean through the basin. The mean interval between consecutive events is 4–5 days.

To localize the changes in the pathways by which water vapor is transported through the basin, it is convenient to consider biweekly averaged moisture transport maps. Figures 4a–4d show the maps for August–September 1994. We can see that in early August, a filament of increased Q is located as shown on Fig. 4a; moisture moves from North Pacific, over the Bering Strait and northeastward in the northern parts of Alaska, from where it propagates through the Mackenzie basin. One such event that occurred during this period is shown in Fig. 3a. The detailed analysis of the geopotential height at 700-mb surface (z_{700}) maps shows that the filament is created by small low pressure centers moving northeastward and then southeastward into the Mackenzie basin along the boundary of a massive and stable high pressure zone in the Pacific that exists throughout the early part of August 1994. That is, the transport of water vapor into the Mackenzie basin occurs in the direction of anticyclonic flow.

By the end of August, as the North Pacific high pressure zone disappears, the filament of moisture flow moves southward (Figs. 4b–4d). The structure of the flow also changes. The quasi-stationary trough, seen in Fig. 4b along the Pacific coast line, develops into the quasi-stationary low, often observed over the Gulf of Alaska in autumn and known as the Aleutian low. The pulses of moisture over the Mackenzie basin are now brought into the basin by the cyclonic flow around this quasi-stationary low. An example of such event is shown in Fig. 3b. The biweekly mean fields for October are not displayed, as the situation in October is similar to that in the second half of September.

These observations support the results of Bjornsson et al. (1995) on high correlation between the cyclogenesis in North Pacific and the Mackenzie River runoff. Indeed, during the season under consideration, most of moisture that the basin receives from the atmosphere is brought by cyclones that originate over the North Pacific.

Such filaments of high moisture transport in the atmosphere are sometimes called atmospheric rivers (Zhu and Newell 1994), or “tropospheric rivers” (Newell et al. 1992). Newell and Zhu (1994) associated these “rivers” with storm tracks and performed the harmonic analysis of the Q fields, calculated from 1-yr global data, to localize the areas with highly variable moisture transport. They found the highest amplitude oscillations in a domain of frequencies shorter than 3 days. However, low-frequency harmonics (with periods longer than 4 days) are harder to detect as a result of changes in the position of the river axis.

A question of the relation between the atmospheric river and storm tracks in the area is of interest. In Fig. 5, tracks of lows, represented by geopotential height at 700-mb minima, and mean moisture transport fields for 2 biweekly periods have been plotted (Fig. 5a displays the second half of August, Fig. 5b, the first half of October). Also shown are fields of standard deviation of geopotential height at 700 mb [\sigma(z_{700})]. Storm tracks are often associated with zones of high standard deviation of geopotential height at a fixed pressure level as a function of time (Blackmon 1977; Lau 1988), since regions in which there is a higher occurrence of storms are characterized by highly variable pressure. In August (Fig. 5a), the zone of enhanced moisture transport is collocated with the tracks of storms for this period. This is in agreement with Figs. 3a and 4a,b, which indicated that the transport was associated with mesoscale and synoptic-scale lows propagating north of a ridge axis over the Gulf of Alaska. In October (Fig. 5b), as well as in September, which is not displayed, the zone of enhanced transport propagates far to the east of the region of high occurrence of lows (the latter, in most cases, do not propagate farther than the Rocky Mountains). This is again in agreement with Fig. 3b and 4d that indicated that the transport occurred ahead of the center of synoptic-scale lows in the Gulf of Alaska. We can also observe that in our case, the maps of biweekly \sigma(z_{700}) do not always represent the storm tracks adequately.

The southward motion of the main stream of the river and the transition between transport regimes can be better seen from the Hovmoeller diagram (see appendix A) of the normal component of Q across the 130°W meridian, that intersects the Mackenzie basin to the east of the Rocky Mountains (Fig. 6). The diagram displays the continuous image of the atmospheric river: one can discern individual weather events, estimate their size,
Fig. 5. Tracks of $z_{700}$ minima, biweekly means of $Q$ (arrows) and the standard deviation of $z_{700}$ over the same time periods: (a) 15–31 Aug, 1–15 Oct 1994. The maximum vector length is 1000 kg (m s$^{-1}$). Tracks are numbered in order of appearance.
Fig. 6. Normal Hovmoeller diagram of $Q$ across 130°W long. The vertical axis, lat from 30°N to 85°N, the horizontal axis, time (days from 1 to 92). Red color indicates positive (eastward) flow, blue, westward flow.

intensity, duration, periodicity. In early August, the eastward flow (red) at 70°–80°N and westward (blue) at 55°N are parts of the same anticyclonic flow around the quasi-stationary ridge seen at Fig. 4a. Then, in early September, we observe the same transition in the flow structure that we have observed in Fig. 4b–4d, with the river axis now at 55°N. The transition occurs around day 42, when the quasi-stationary low over the Gulf of Alaska is established.

In Fig. 7, a Hovmoeller diagram of moisture transport into the Mackenzie basin is displayed. The vertical axis is the smoothed contour $s$ of the basin (see appendix A), $s = 0$ and 1 being the easternmost point of the basin. Here, red color means inflow, blue color, outflow. The

Fig. 7. Normal Hovmoeller diagram of $Q$ across the Mackenzie basin contour. The vertical axis is the isometric parameter along the contour, from 0 to 1. $s = 0$ and 1 correspond to the easternmost point of the basin; $s = 0.3$—to the river mouth; $s = 0.75$—to the southmost point of the contour. The direction around the basin is counterclockwise. Red color indicates inflow, blue, outflow.
horizontal stripes of white indicate intervals on the contour where there is strong curvature (e.g., the southern point of the contour at $s = 0.75$). Since the size of the basin is comparable with the size of weather phenomena bringing moisture into or out of the basin, we cannot discern the structure of these events. However, this picture shows even more clearly the discrete structure of water vapor transport through this area. The change in the direction of the transport can also be observed. In August, the main inflow of moisture comes from the north and northwest through the northern part of the contour ($0.25 < s < 0.75$), while in September and October (after day 42 or so), it arrives from the west and southwest ($0.25 < s < 0.75$). We have already seen this transition in Fig. 6. At the eastern side of the contour, we can see weakened moisture pulses exiting the basin. In August, this occurs along the southeastern part ($s > 0.75$) of the contour, while after the transition, they shift to the northeast ($s < 0.2$). In addition, the seasonal decrease of the intensity of moisture pulses passing through the basin is best seen from day 42 to 92.

It therefore seems appropriate to find out if there are any quantitative differences in the flow before and after the transition that has been identified. Let us return to Fig. 2 that shows each component of Eq. (5) over the Mackenzie basin as a time series, the dotted line being the $(-\partial W/\partial t)$ series. Walsh and Chapman (1990) showed that the precipitable water component makes only $15\%\,-20\%$ of the total balance for monthly means in high latitudes. This is also true for the Mackenzie basin, but only for the mean values $\langle P - E \rangle = 1.5 \times 10^7$ kg s$^{-1}$, $\langle -\text{div} \mathbf{Q} \rangle = 1.2 \times 10^7$ kg s$^{-1}$, $\langle -\partial W/\partial t \rangle = 0.3 \times 10^7$ kg s$^{-1}$, not for any instant in time. In Fig. 2, we can see that the absolute values of $\langle -\partial W/\partial t \rangle$ and $\langle -\text{div} \mathbf{Q} \rangle$ are of the same order of magnitude. Moreover, we observe that $\langle -\partial W/\partial t \rangle$ and $\langle -\text{div} \mathbf{Q} \rangle$ are strongly anticorrelated, while the value of $\langle P - E \rangle$ is significantly smaller than each component and is positively correlated with $\langle -\text{div} \mathbf{Q} \rangle$. These correlation effects can be seen most clearly after day 42, when the change in the flow structure, observed in Figs. 4–7, occurs. The cross correlations between $\langle -\partial W/\partial t \rangle$, $\langle -\text{div} \mathbf{Q} \rangle$, and $\langle P - E \rangle$ are shown in Table 1. Indeed, after the transition, the correlation between $\langle -\partial W/\partial t \rangle$ and $\langle -\text{div} \mathbf{Q} \rangle$ becomes closer to $-1$ than it was before, while the correlation between $\langle -\partial W/\partial t \rangle$ and $\langle P - E \rangle$ becomes even closer to 0. This means that most incoming moisture that enters the basin after the transition does not precipitate immediately (in case of $\langle P - E \rangle = 0$ we would have $100\%$ anticorrelated $\langle -\partial W/\partial t \rangle$ and $\langle -\text{div} \mathbf{Q} \rangle$), but either stays in the atmosphere as precipitable water, to precipitate later, or is transported out of the basin.

Table 1. Pairwise cross correlations between $\langle -\partial W/\partial t \rangle$, $\langle -\text{div} \mathbf{Q} \rangle$, and $\langle P - E \rangle$. Here, $C(x, y)$ means the correlation of $x$ and $y$. Correlations are computed for the entire period and two “half” periods that correspond to the two transport regimes identified in the paper.

<table>
<thead>
<tr>
<th>Time interval</th>
<th>Days 1–92</th>
<th>Days 1–42</th>
<th>Days 43–92</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C(\langle -\partial W/\partial t \rangle, \langle -\text{div} \mathbf{Q} \rangle)$</td>
<td>$-0.77$</td>
<td>$-0.59$</td>
<td>$-0.84$</td>
</tr>
<tr>
<td>$C(\langle -\partial W/\partial t \rangle, \langle P - E \rangle)$</td>
<td>$0.14$</td>
<td>$0.38$</td>
<td>$-0.03$</td>
</tr>
<tr>
<td>$C(\langle P - E \rangle, \langle -\text{div} \mathbf{Q} \rangle)$</td>
<td>$0.53$</td>
<td>$0.52$</td>
<td>$0.57$</td>
</tr>
</tbody>
</table>

At the moment, we are unable to produce reliable maps of moisture flux divergence for individual moments of time, even though most of the “noise” related to sparse data seems to vanish after averaging over long periods of time [see also, for instance, Schmitt and Mullen (1996) who also concluded that the ECMWF data with $1.125^\circ$ resolution do not allow to estimate moisture flux divergence accurately]. However, even if a smaller grid spacing was used, there are but a few radiosonde stations in the area, and the reliability of $P - E$ estimates for small parts of the basin would be questionable anyway, since the lack of observation datasets a limit for the analysis validity.

b. ECMWF analyses versus radiosonde data

This section will discuss the ability of the ECMWF analyses to represent various diagnostic fields, other than moisture flux divergence. The most important topic is, of course, how well $\mathbf{Q}$ itself is estimated by the analysis.

The values of five basic meteorological functions (wind speed vector, relative humidity, temperature, geopotential height) as well as the values of moisture transport vector given by ECMWF analysis and by radiosonde soundings have been compared for 10 radiosonde stations (Fig. 1) inside and in the vicinity of the Mackenzie basin.

The choice of stations was based on several factors; namely, it included stations inside and outside the basin, at the Pacific coast as well as inland, near the mountains and far from those, in Canada as well as in the United States. Also, most selected stations are located on the way of the atmospheric river that is detected by our analysis.

It is known that relative humidity is the parameter least accurately described by objective analyses (e.g., Trenberth and Olson 1988). Our results again demonstrate this fact. Figures 8 (Annette Island, Alaska) and 9 (Fort Nelson, Northwest Territory) show the vertical profiles of October 1994 monthly mean zonal wind (Figs. 8a, 9a), relative humidity (Figs. 8b, 9b) and the
**Fig. 8.** Vertical profiles of October mean zonal wind magnitude (a), relative humidity (b), and the product of specific humidity and zonal wind (c) for 55°N, 132°W, the radiosonde station at Annette Island, AK. The analyses are solid lines, the soundings are dotted lines.

**Fig. 9.** Same as Fig. 8, but for the radiosonde station at Fort Nelson, NWT (59°N, 123°W).
product of specific humidity and zonal wind ($qu$) (Figs. 8c, 9c). We can see that while analyzed and radiosonde wind speed profiles are almost identical, the relative humidity is considerably overestimated by the analysis at 500–400-mb levels. In any case, this discrepancy has no significant effect on moisture fluxes that turn out to be reproduced by the analysis with much better accuracy than the relative humidity fields (Figs. 8c, 9c). This is, of course, because the region in which there is a large error in relative humidity is the one in which the specific humidity is low due to cold temperatures. Still, the analysis tends to overestimate the value of moisture flux (e.g., Fig. 9c), with the magnitude of the difference up to 10% for monthly means of the integrated transport.

To compare temporal structures of observed and analyzed moisture fluxes, values of $Q$, as well as $Q_{hi}$ and $Q_{lo}$, have been calculated from the soundings (snd) and from the analyses (ana) every 12 h for the 10 locations shown in Fig. 1. The results of this comparison for two stations (Annette Island, Alaska, and Fort Nelson, Northwest Territory) are shown in Figs. 10, 11, and 12. Figure 10 shows the values of $Q$ for the two stations, Fig. 11, the values of $Q_{hi}$. Fig. 12, the values of $Q_{lo}$. Each figure displays graphs of three functions of time: the solid line is the analysis-based value of zonal mois-
ture flux, the dotted line is its observed value, and the thin solid line in the lower part of the plot (corresponding to the right scale) is the relative error defined as

\[ r = \frac{|Q_{\text{ana}} - Q_{\text{snd}}|}{\sigma(Q)} \]  \hspace{1cm} (7)

where \( \sigma(Q) \) is the standard deviation of \( Q_{\text{ana}} \) or \( Q_{\text{snd}} \) \( [\sigma(Q_{\text{ana}}) = \sigma(Q_{\text{snd}})] \), which is a good estimate of mean magnitude of \( Q_{\text{ana}} \) or \( Q_{\text{snd}} \) when their mean values are very small, as is the case.

We can see that despite these discrepancies, the analyzed and observed curves are fairly close to each other, and the relative errors defined by (7) keep at the average level of 0.2. The discrepancies may be caused by as flaws in analyzed data, as by wrong observations. There are instances when errors are as large as 0.7–0.8 (e.g., Fig. 12, day 42–46, the period when the transition in flow regime was occurring, or Fig. 12b, day 53, where radiosonde data look suspicious). Another source of discrepancies is the fact that the radiosonde data are available only twice a day, while the analyses are available every 6 h.

The mean discrepancy of 20% between the analysis and radiosonde data means at least 35% uncertainty in \( \langle \text{div}Q \rangle \) (see appendix B). However, it is not clear how much of it is due to errors in the analysis and how much to those in the radiosonde data.

Comparing the quality of analyses for upper-and lower-level moisture fluxes, we observe that lower-lever fluxes are described by the analysis less accurately, especially at Canadian inland stations: Fort Nelson, Fort Smith, Norman Wells. Mean values of \( r \) for \( Q_{\text{lo}} \) are 0.15–0.18, while for \( Q_{\text{hi}} \) they are closer to 0.25 (compare Figs. 11 and 12). In Fig. 12b (days 1–30), we observe the discrepancies between the analysis and the soundings that cannot be related to such trivial sources of error as bad or missing radiosonde data or the nonstandard time of radiosonde launch (e.g., 0900 UTC instead of 1200 UTC). A possible source of error is the “phasing error,” that is, the analysis predicts the event correctly but at a wrong time (e.g., Fig. 11a, day 35, Fig. 11b, day 48). It is also noticed that discrepancies depend on height: \( Q_{\text{lo}} \) is described by the analysis less accurately than \( Q_{\text{hi}} \) (compare Fig. 11a with Fig. 12a and Fig. 11b with Fig. 12b). The correlation between analyzed and observed \( Q \) and \( Q_{\text{lo}} \) is about 0.9 for all 10 stations, while for \( Q_{\text{hi}} \) it is closer to 0.75–0.8. A possible explanation is that the analysis does not capture the features of topography in the region, such as the roughness of surface in mountains.

The following conclusions can be made on the applicability of ECMWF analysis to moisture transport studies: 1) the relative humidity is predicted by the analysis least accurately among the other meteorological parameters. However, this does not have much effect on the values of vertically integrated moisture fluxes; 2) the estimate of low-level (below 700 mb) moisture transport is of lower quality than that of upper-level transport; 3) the time series of analyzed and observed moisture fluxes are highly correlated, 4) the “relative error” [defined by (7)] keeps mostly at 20% level, and the difference between the analysis-based and observed monthly mean values of \( Q \) never exceeds 10%. However, the error in the integrated moisture flux convergence can be as large as 35%. This all suggests that the ECMWF analyses have limited applicability as a source of data for water vapor transport studies in the region.

5. Summary

This paper describes the first results obtained regarding the spatial and temporal variability of the atmospheric water vapor transport through the Mackenzie River basin. The data for August, September, and October 1994 have been analyzed in detail. In future, all the reanalyzed data since 1979 will be considered, so that the annual cycle as well as the interannual variability can be documented.

The following important aspects of humidity transport have been revealed and will be further studied in the project.

1) A structure described by Zhu and Newell (1994) as an atmospheric river has been identified over the Mackenzie River basin during the season under consideration. This structure consists of a chain of moisture pulses coming into the Mackenzie basin from the Pacific Ocean, carried either by low pressure centers, or by the external parts of cyclonic eddies passing outside the basin. The position of the river changes with time. In August, its axis passes from the North Pacific to the northern parts of Alaska and the Beaufort Sea and comes to the Mackenzie basin from the northwest, while in late September and October, it passes over the Rocky Mountains, with its axis located near the 55°N latitude. The transition between the two regimes occurred in early September. In both cases, the weather systems carrying moisture have their origin over the North Pacific, and this supports the results by Bjornsson et al. (1995). The comparison between maps of moisture transport and the storm tracks suggests that the concept of an atmospheric river is wider than one of a storm track: spatial boundaries of increased moisture transport can extend far beyond the zone of tracks of the storms that generate these moisture flows. The periodicity of moisture pulses passing through the basin was found to be approximately 4 days (25 for 92 days).

2) As expected, precipitation minus evaporation integrated over the basin, \( \langle P - E \rangle \), is found to be correlated with the convergence of \( Q \) across the basin’s boundary. This implies the dependence of the water balance on transients. Meanwhile, \( \langle P - E \rangle \) is usually smaller than each of its components \( \langle -\partial W/\partial t \rangle \) and
The applicability of the ECMWF objective analysis data to the moisture transport studies has been investigated by comparison of values of moisture transport computed from the analysis data and from radiosonde data. Even though relative humidity fields often tend to be overestimated by the analysis, the spatial (vertical) and temporal structure of moisture fluxes is depicted by the analysis with sufficient accuracy. However, the uncertainty in the convergence of moisture flux integrated over the Mackenzie basin is estimated as high as 35%, and the field of moisture flux convergence cannot be obtained with sufficient accuracy from the analyses. We conclude that the ECMWF analyses provide fair estimates for a broad but limited set of hydrological parameters.

The results presented in this paper, as well as the methods we have developed (such as the extension of Hovmoeller diagrams to curvilinear and variable contours and vector functions), may be productive in studies of the spatial and temporal structure of atmospheric rivers and of other climatological and hydrological phenomena in the atmosphere and ocean.

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APPENDIX A

Hovmoeller Diagrams as a Method of Diagnostics Visualization

One of the problems encountered in the analysis of water vapor flux as a function of space and time is the standard problem of visualizing three- and four-dimensional functions. A common technique for visualization of time-dependent fields is known as Hovmoeller diagrams.

This technique can be generalized to display any time-dependent fields in time-dependent coordinates. For example, if a scalar function \( F(x, y, t) \) is defined on an \((x, y)\) plane, and a certain curvilinear contour is given as \( x(s), y(s), s \) being an isometric parameter such that \( x^2 + y^2 = \text{const} \), then a function

\[
F(s, t) = F[x(s), y(s), t]
\]

can be plotted in \((s, t)\) coordinates.

Similarly, if \( \mathbf{Q}(x, y, t) \) is a vector function of space and time, and \( x(s), y(s) \) a contour in \((x, y)\) plane, as above, scalar functions

\[
Q_n(s, t) = \{Q[x(s), y(s), t]n\}, Q_t(s, t) = \{Q[x(s), y(s), t]\tau\}
\]

\([n(s) \text{ and } \tau(s) \text{ being the normal and tangential vectors to the contour}]\) can be plotted. In this case, \( Q_n \) is the normal flux of \( \mathbf{Q} \) across the contour, and \( Q_t \) is the tangential flux. Contour plots of \( Q_n \) and \( Q_t \) in \((s, t)\) coordinates (normal and tangential Hovmoeller diagrams) provide a lot of information about temporal structure of the fluxes of \( \mathbf{Q} \). For instance, when \( \mathbf{Q} \) is the moisture transport vector and \( x(s), y(s) \) is the contour of the Mackenzie basin, the normal Hovmoeller diagram shows in detail where, when, and how much moisture is transported through the basin.

APPENDIX B

Error Estimate for \( \langle \text{div} \mathbf{Q} \rangle \)

We found that there is a 20% discrepancy between values of \( \mathbf{Q} \) as given by the analyses and by the radiosonde data. Now the estimate of error in

\[
\langle \text{div} \mathbf{Q} \rangle = \int_A \int_A \text{div} \mathbf{Q} \, dA = \int_{\partial A} \mathbf{Q}_n \, dt
\]

[according to (5)] is needed. \( A \) being the Mackenzie basin area and \( \mathbf{Q}_n \) a normal component of \( \mathbf{Q} \) to the basin contour.

If \( \mathbf{Q} \) is given on a lat–long grid, then any estimate of \( \langle \text{div} \mathbf{Q} \rangle \) is a linear combination of \( \mathbf{Q}_n \) at boundary grid points

\[
\int_{\partial A} \mathbf{Q}_n \, dt = \sum_{i=1,N} C_i q_i,
\]

where \( q_i \) is a value of \( \mathbf{Q}_n \) at \( i \)th boundary grid point. For a rectangular area, \( C_i \) would be all equal to the grid length \( D = 1.125^6 \times 1.1 \times 10^5 \text{ m} \); let us suppose that it is approximately so in our case. Thereby, we estimate \( \langle \text{div} \mathbf{Q} \rangle \) as

\[
\langle \text{div} \mathbf{Q} \rangle = D \sum_{i=1,N} q_i.
\]

We assume that there is no systematic error in \( \mathbf{Q} \) represented by the objective analysis, that is, the errors in different grid points are independent from each other. In this case, the uncertainty in \( \langle \text{div} \mathbf{Q} \rangle \) will be
\[ \Delta(\text{div} Q) = D \sqrt{\sum_{i=1}^{N} (\Delta q_i)^2}. \]

A typical rms value of \( Q \) is 100 kg m\(^{-1}\) s\(^{-1}\), and all \( \Delta q \) can be assumed equal to \( \Delta q_1 \), which is 20% of the rms value of \( Q \) that is, 20 kg m\(^{-1}\) s\(^{-1}\). Therefore,
\[
\Delta(\text{div} Q) = D \sqrt{N} (20 \text{ kg m}^{-1} \text{ s}^{-1}) = 1.2 \times 10^3 \text{ m} (20 \text{ kg m}^{-1} \text{ s}^{-1}) \sqrt{N}.
\]

In our case, \( N = 58 \), so \( \Delta(\text{div} Q) = 1.6 \times 10^3 \text{ kg s}^{-1} \). The rms value of \( \text{div}(Q) \) is \( 4.5 \times 10^7 \text{ kg s}^{-1} \). It means that the error estimate for \( \text{div}(Q) \) is 1.6/4.5 = 35%.

If the errors in \( q_i \) have a systematic component, then the error in \( \text{div}(Q) \) will be even larger. However, it is not clear, how much of the 35% are due to objective analysis and how much to the flaws in radiosonde data.

This estimate does not depend on a numerical method employed to calculate \( \text{div}(Q) \). The accuracy of the \( \text{div}(Q) \) estimate therefore can only be improved by increasing the resolution of the objective analysis, or other improvements of the quality of radiosonde data and the methods and resolution of the objective analyses.

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