Atmospheric water vapor transport and continental hydrology over the Americas

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Abstract

The advective transport of atmospheric water vapor, its role in global hydrology, and the water balance of continental regions are discussed and explored. A data set consisting of 10 years of global wind and humidity observations that has been interpolated onto a regular grid is used to estimate atmospheric water vapor fluxes across the boundaries of selected continental regions. The flux of water vapor from the waters surrounding the Americas is computed and graphically represented as integrated normal flux across meridional and zonal segments. The total flux is decomposed into transport by mean motion and transport by correlated transient anomalies in the wind and humidity fields. Seasonal maps of these observations for the Americas reveal the underlying mechanisms and illustrate the characterizing features of the hydrologic cycle and continental water balance for this large land mass.

1. Introduction

Water vapor, as a variable constituent of the atmosphere, accounts for only 0–4% by volume of tropospheric air. Because saturation vapor pressure increases exponentially with temperature, the vapor content of the atmosphere decreases with height, with the result that atmospheric water vapor is concentrated in the lower troposphere. The atmosphere generally contains about $12 \times 10^3$ km$^3$ equivalent of liquid water, enough to cover the Earth's surface to a depth of about 24 mm (precipitable depth). Given this storage and the global average precipitation rate of 1000 mm year$^{-1}$, the average residence time of a water molecule in the atmospheric reservoir is about 9 days.

Water vapor is transported in the atmosphere by molecular and turbulent diffusion, convection, and advection. The vertical flux of evaporation from the oceans and the
land surface into the atmospheric boundary layer is accomplished by diffusion, and
the vapor is vertically mixed by diffusion and convection. On the temporal and spatial
scales of lateral transport, advection is the dominant mechanism. On these scales,
water vapor can be treated as a scalar admixture advected by the horizontal wind.

In this paper the transport of water vapor within the general circulation of the
atmosphere is considered. A graphical presentation of the large-scale land–ocean
atmospheric exchange of water vapor is made to illustrate the coupling of the atmos-
pheric branch of the hydrologic cycle to its well-known land component. In this way,
the various transport mechanisms and their seasonal and geographical variations are
identified.

The global- and continental-scale transport of water vapor has important impli-
cations for climate variability and hydrology owing to the following factors: (1)
enormous amounts of energy are consumed or released upon phase changes of
H$_2$O; (2) the distribution of the suspended liquid phase (clouds) and the ice phase
of water (snow or sea ice) significantly reduce the efficiency with which the Earth
system converts the incident solar radiation to energy available and useful for driving
dynamic atmospheric motion; (3) water vapor is the principal greenhouse gas respon-
sible for re-radiating the longwave flux emitted by the heated Earth’s surface; (4)
water is a limiting factor in the functioning of the biosphere, which exerts significant
control over various hydrologic and atmospheric processes. Furthermore, these
factors are linked and have influential feedbacks on each other.

After a brief review of recent work on the atmospheric branch of the hydrologic
cycle, the mathematical framework for representing the transport of water vapor in
the atmosphere is outlined. Specifically, the linkages and coupling between the state-
ments of water balance for a surface–subsurface soil control volume and that for the
overlying atmosphere (from the surface boundary to a height above which a negligible
amount of water vapor exists) are considered. This consideration has implications for
interpreting the sources and nature of natural fluctuations in surface hydrologic
processes as well as for understanding the modulated response of variations in
climate where land–atmosphere interaction is significant. Finally, a gridded global
observed data set on general circulation and humidity is used to derive and represent
the land–ocean atmospheric exchange at continental scale. The transport processes
are decomposed into their constitutive components — transport by mean airmass
general circulation and transport by transient eddies in the atmosphere. For the case
of the Americas, the seasonal climatology of the continental-scale atmospheric
branch of the hydrological cycle is presented.

2. Background

The atmospheric water vapor transport vector has been an object of study, on both
the global and the regional scale, since the availability (after World War II) of the
aerological data required for vapor flux computations. A decade of balloon and
radiosonde observations provided the data for Benton et al. (1950) to clarify the
time- and distance-scales of vapor transport in the atmosphere, and to demonstrate
the important role of large-scale atmospheric motion in the hydrologic cycle. Benton and Estoque (1954) used the vapor flux vector divergence field in computations of evapotranspiration over the North American continent for the calendar year 1949. In his Presidential Address to the Royal Meteorological Society, Sutcliffe (1956) called for increased scientific inquiry into the problem of water balance and the general circulation of the atmosphere, and recommended wider application of the techniques introduced by Benton and Estoque. Starr and Peixoto (1958) found centers of divergence of atmospheric water vapor flux over many of the world's arid zones, and inferred from mass balance considerations the existence of net subsurface inflow to these regions. Lufkin (1959) made the first estimate of water vapor transport from the oceans to the continents as a function of latitude. Hastenrath (1966) analyzed water vapor flux over the Central American seas as part of a more comprehensive study concerned with the general circulation and energetics in that area. Rasmusson (1967, 1968) investigated water vapor flux over North America and the Central American seas as part of a continental water-balance study; this work was followed by a regional study of the hydrology of eastern North America (Rasmusson, 1971). Rasmusson (1977) recommended the application of vapor flux data in the routine computation of regional water balances over areas large enough to control the relative error of the resulting flux divergence estimates (he recommended $10^6$ km$^2$ or more). He also provided guidelines for operational vapor flux computations within the context of the World Weather Watch. Peixoto and Oort (1983) used the divergence field to study connections between the atmospheric and surface branches of the hydrologic cycle. Chen (1985) analyzed water vapor transport and maintenance during the solstice seasons to explore how the high water vapor content of certain tropical areas is maintained by the large-scale atmospheric circulation. Salstein et al. (1983), Trenberth et al. (1987) and Chen and Pfauendnter (1993), among others, have studied the interannual variability of the vapor flux regimes at the global scale.

Most relevant to the study here are the investigations of the hydrologic and atmospheric branches of the hydrological cycle that have estimated the exchange of water vapor across coastlines (e.g. Horton, 1943; Benton et al., 1950; Hutchings, 1961; Drozdov and Grigorieva, 1965; Rosen and Omolayo, 1981; Shiklomanov, 1989). Rosen and Omolayo (1981) made estimates of the seasonal Northern Hemisphere cross-coastal flux of water vapor. Their study of the total vapor flux (by all transport mechanisms combined) reveals important features of the water vapor exchange imbedded in the general circulation. Hutchings' (1961) study of the water vapor transport over the Australian continent is also revealing. Bavadekar and Mooley (1978) performed land water-balance component estimates using knowledge of the regional vapor transport regime.

3. Basic concepts

3.1. The vertically integrated vapor flux vector

The atmosphere's specific humidity and the zonal and meridional components of
the wind are variable in both space and time. In the geophysical coordinate system, the four-dimensional domain of these variables is defined by \( \phi \) (latitude), \( \lambda \) (longitude), \( p \) (pressure), and \( t \) (time). Following Peixoto (1973) and Peixoto and Oort (1983), the dimensionality of the domain is reduced to two by defining a vertically integrated, time-averaged vapor flux vector.

At a given point in space and time, a vector field of the advective transport of water vapor by the horizontal wind is defined as follows:

\[
q V(\lambda, \phi, p, t) = (qu)i + (qv)j
\]

in which \( q \) is the specific humidity (mass \( H_2O \) vapor per mass air), \( V \) is the horizontal wind vector (LT\(^{-1}\)), \( u \) is the zonal wind component (positive eastward), \( v \) is the meridional wind component (positive northward), and \( i \) and \( j \) are the unit vectors in the zonal and meridional directions, respectively. The horizontal transport vector \( q V \) has dimension L T\(^{-1}\). As a product of terms, \( q V \) increases with either velocity or specific humidity; for example, slow-moving moist air may transport water vapor at the same rate as less moist but fast-moving air.

For a vertical atmospheric column with a unit base, the vertically integrated horizontal vapor flux, \( Q \), is obtained by taking the mass-weighted vertical integral of Eq. (1):

\[
P_0 Q(\lambda, \phi, t) = \int_0^{P_0} q V \frac{dp}{g} = Q_\lambda i + Q_\phi j
\]

in which \( P_0 \) is the surface pressure, \( g \) is the acceleration of gravity, and \( Q_\lambda \) and \( Q_\phi \) are, respectively, the zonal (positive eastward) and meridional (positive northward) components of \( Q \). The vapor flux vector \( Q \) has dimension M T\(^{-1}\) L\(^{-1}\). As a vertically integrated quantity, \( Q \) expresses the magnitude and direction of the net transport of water vapor through the depth of the atmosphere above a point on the Earth's surface; a small value of either component does not necessarily imply negligible transport at all levels in the atmosphere.

Eq. (2) may be averaged over a time period \( \tau \) to compute the corresponding mean values \( \bar{Q}_\lambda \) and \( \bar{Q}_\phi \), where the overbar denotes the time-average operator. The time-averaged, vertically integrated moisture flux vector \( \bar{Q} \) has dimension M T\(^{-1}\) L\(^{-1}\); it is a function of two horizontal space variables, latitude and longitude.

### 3.2. Atmospheric water balance

The mass conservation equation for water vapor in the atmosphere may be expressed as

\[
\frac{\partial W}{\partial t} + \nabla \cdot Q = E - P
\]

in which \( W \) is the water vapor storage within the atmospheric column, \( E \) is the rate of evapotranspiration into the base of the column, \( P \) is the rate of precipitation from the column, also measured at the surface, and \( \nabla \cdot Q \) is the divergence of the lateral vapor
flux, with $Q$ as defined above. $W$, also known as the precipitable water in the column, may be expressed as either water vapor mass per unit surface area ($M \cdot L^{-2}$) or equivalent depth of liquid water (L). The latter unit is more common in hydrologic practice; however, the former is used in this paper.

For an atmospheric control volume, bounded by a conceptual closed vertical wall and overlying a region of the Earth's surface having area $A$, Eq. (3) takes the form

$$\left( \frac{\partial W}{\partial t} \right) + (\nabla \cdot \mathbf{Q}) = (E - P)$$

(4)

where the angled brackets denote a space average. Applying Green's theorem allows Eq. (4) to be written

$$\left( \frac{\partial W}{\partial t} \right) = (E - P) - \frac{1}{A} \oint Q \cdot \mathbf{n}_\gamma \, d\gamma$$

(5)

in which $\mathbf{n}_\gamma$ is the outward unit vector normal to the horizontal boundary ($\gamma$) of the region. Eq. (5) states that the time rate of change in water vapor storage within the atmospheric volume is equal to evapotranspiration minus precipitation less the net lateral outflow through the vertical boundaries of the control volume.

Eqs. (4) and (5) express the mass balance of water in the atmospheric branch of the hydrological cycle. An excess of evapotranspiration over precipitation is balanced by the change in water content of the atmosphere and the net outflow of vapor across the boundaries of the region. A region in which the mean divergence of $\mathbf{Q}$, $\langle \nabla \cdot \mathbf{Q} \rangle$, is positive is a source of water vapor to the rest of the atmosphere, whereas a region with negative $\langle \nabla \cdot \mathbf{Q} \rangle$, a region of convergence, is a sink of atmospheric water vapor. The mean net transport of water into a region by horizontal atmospheric motion is equal to $-\langle \nabla \cdot \mathbf{Q} \rangle$ ($MT^{-1} L^{-2}$) and the total transport into the region is the closed-path integral $\oint Q \cdot \mathbf{n}_\gamma \, d\gamma$, ($MT^{-1}$) with the normal to the boundary defined outward as in Eq. (5). In analyzing and modeling the atmospheric water balance, the expression 'div $Q$' (omitting the negative sign) is frequently used in reference to either of these quantities.

3.3. Soil water balance

In the land surface branch of the hydrologic cycle, the mass balance for a soil column control volume can be written in the simplified form

$$\left( \frac{\partial S}{\partial t} \right) = (P - E) + \langle R_{in} - R_{out} \rangle$$

(6)

in which $S$ is the mass of water stored in the surface and subsurface reservoirs of the region, $R_{in}$ is the rate of water influx across the horizontal boundaries of the region by surface and subsurface lateral transport, and $R_{out}$ is the rate of water efflux by these lateral processes. If the region's boundaries are surface drainage divides, and if the subsurface drainage corresponds to that on the surface, then $\langle R_{in} \rangle = 0$ and the term $\langle R_{in} - R_{out} \rangle$ is simply the rate of runoff per unit area; such a simplification is also the
case when the study region is an entire continent. For smaller land regions that are not
delineated along natural drainage divides, surface and subsurface lateral transport
into a region may be significant.

Eq. (6) may be rearranged as follows:

\[ (P - E) = -\langle E - P \rangle = \langle R_{\text{out}} - R_{\text{in}} \rangle + \frac{\partial S}{\partial t} \]  

which states that an excess of precipitation over evapotranspiration is balanced by net
surface and subsurface runoff and/or increased terrestrial storage in the region.

The term \( \langle E - P \rangle \) is common to Eqs. (5) and (7), and thus provides the link between
the terrestrial and atmospheric branches of the hydrologic cycle:

\[ \langle R_{\text{out}} - R_{\text{in}} \rangle + \frac{\partial S}{\partial t} = -\langle \nabla \cdot \mathbf{Q} \rangle - \frac{\partial W}{\partial t}. \]  

Therefore, if the aerological data are known, as well as runoff and precipitation,
Eqs. (5) and (8) may be used to estimate the average rate of change of groundwater
storage and the average evaporation rate, two quantities that are difficult to measure.
Eq. (8) also reveals the identity of atmospheric convergence and land runoff. Typically,
the storage terms \( \partial S/\partial t \) and \( \partial W/\partial t \) are smaller than the other terms by at least
one order of magnitude. Over an annual cycle, when \( \partial S/\partial t \) and \( \partial W/\partial t \) can be
neglected, the convergence of atmospheric water vapor is necessarily equal to the
land liquid water runoff. Clearly, large amounts of heat release are associated with
this equality of water mass in different phase. In the tropics, where there is abundant
river outflow from the continents, the atmosphere thus receives a large deposit of
latent heat; this heating is a major source of tropical dynamic air motion regimes.

The accuracy of atmospheric runoff estimates is highly sensitive to the quality of
the measurements. Accuracy is improved by enlarging the path of the boundary
integral (5), which leads to the recommendation of Rasmusson (1977) that the region
should have an area of at least \( 10^6 \) km\(^2\).

3.4. Modes of transport

Any atmospheric variable, \( f \), may be expressed in terms of a time mean and a
perturbation from that mean:

\[ f(\lambda, \phi, p, t) = \bar{f}(\lambda, \phi, p) + f'(\lambda, \phi, p, t) \]  

in which the overbar denotes the mean value over a specified time interval. By
definition, the mean value of the perturbations is zero (\( \bar{f}' = 0 \)).

For example, expressing the zonal component of the horizontal wind and the
specific humidity in this manner, the time mean of the zonal component of the
horizontal vapor transport vector, \( qV \), contains cross-product terms:

\[ \overline{q\bar{u}}(\lambda, \phi, p) = \overline{(q + q')(\bar{u} + u')} \]
\[ = \overline{q\bar{u}} + \overline{q'u} + \overline{\bar{q}u} + \overline{q'u} \]
\[ = \overline{q\bar{u}} + \overline{q'u} \]  

\[ (10) \]
The first term on the right-hand side of Eq. (10) represents the advection of mean specific humidity by mean zonal atmospheric motion. The second term is a correlation term; for example, \( q'u' \) is positive for a given location if winds that are more westerly (or less easterly) than the mean wind \( (u' > 0) \) are more moist than average \( (q' > 0) \), negative if winds that are more easterly (or less westerly) than the mean wind \( (u' < 0) \) tend to be more moist than average, and zero if no correlation exists between the zonal wind and the specific humidity. Likewise, for the meridional component of vapor transport

\[
\bar{q}v(\lambda, \phi, p) = \bar{q}v + q'v'
\]

By convention, the terms on the right-hand sides of Eqs. (10) and (11) are called, respectively, transport by mean motion and transport by transient eddies. The distinction between mean and eddy transport depends upon the averaging period.

The time-averaged vertically integrated water vapor flux vector, \( \vec{Q} \), contains a mean and a transient eddy term in each of its components, i.e.

\[
\bar{Q}_\lambda \equiv \int_0^{p_i} \frac{\bar{q}u}{g} dp = \int_0^{p_i} \frac{\bar{q} \bar{u}}{g} dp + \int_0^{p_i} \frac{q'u'}{g} dp
\]

and

\[
\bar{Q}_\phi \equiv \int_0^{p_i} \frac{\bar{q}v}{g} dp = \int_0^{p_i} \frac{\bar{q} \bar{v}}{g} dp + \int_0^{p_i} \frac{q'v'}{g} dp
\]

Globally, the zonal component of \( \vec{Q} \) tends to exceed the meridional component by an order of magnitude. The mean motion dominates in \( Q_\lambda \), whereas in \( Q_\phi \) the mean and transient eddy terms are of comparable magnitude. Transient eddy transport is particularly important at mid-latitudes and during the active winter season, where the eddies play a critical role in the poleward transport of sensible as well as latent heat (Peixoto and Oort, 1983).

A similar decomposition can be performed in the spatial dimension by averaging with respect to longitude around a belt of constant latitude. The resulting zonal mean and standing eddy terms are of great interest and value in studies of vapor transport over the globe as a whole. The present study focuses on the water balances of continental regions on a spatial scale that is too small to justify such spatial decomposition. However, the distinction between the time mean and transient eddy modes of transport will figure in the computation of \( \vec{Q} \) and the analysis of vapor flux into and out of regions.

4. Observational data and computational method

4.1. Aerological data

The humidity and wind data used in this study are from a data set provided by the
Geophysical Fluid Dynamics Laboratory (GFDL) of the National Oceanic and Atmospheric Administration (NOAA) at Princeton University through the courtesy of Dr. Abraham Oort. The GFDL data set consists of 10 years of observations, transformed into a gridded form. The measurements were taken during the period May 1963–April 1973. For the years 1963–1968, the observations were once daily, at 00:00 GMT, although a small number of 12:00 GMT observations were included. For the years 1968–1973, both 00:00 and 12:00 GMT data were used (Oort, 1983, pp. 5-8). The aerological data were interpolated onto the regular grid by means of an objective analysis scheme, using the zonal average of data in a latitudinal belt as a first approximation. The objective analysis scheme, known as CRAM (Conditional Relaxation Analysis Method) has been described by Oort (1983). Values of $\vec{q}, \vec{u}, \vec{v}, \vec{q'u'}$, and $\vec{q'v'}$ are given at each node of a 2.5° latitude by 5° longitude grid, and at each of 11 pressure levels; the overbars denote monthly averages, with $\overline{\vec{q'u'}}$ and $\overline{\vec{q'v'}}$ being the horizontal transient eddy fluxes in the zonal and meridional directions, respectively.

There are sources of uncertainty and error in both the observations and the analysis technique. Ideally, error bounds and confidence limits should be quantitatively derived for any estimate of a measure based on such data. This is a difficult task; at the least, the sources of uncertainty and error must be recognized.

The major sources of observational error are as follows: (1) the measurements were taken at most twice daily, and the diurnal cycle is not sufficiently captured; (2) the

CASE A: EQUATOR

Fig. 1. A simple example illustrating the meaning of the bars in Figs. 2–13. For constant $Q$, the magnitude of $f_1$ depends upon the normal component of $Q$ and the length of the segment. The right-hand side shows how these results would appear in figures such as Figs. 2–13.
planetary boundary layer, which contains a large fraction of the water vapor in the atmospheric column, is not well resolved in the soundings; (3) the sensors and instrument packages lose calibration and are often of different manufacture. Elliott and Gaffen (1991), among others, discussed the problems associated with radiosonde humidity archives.

As for the analysis technique, the major sources of uncertainty and error are as follows: (1) the analyzed fields are based upon spatially sparse point observations; (2) the technique is sensitive to the first-approximation field, and the zonal averages are likely to be poor estimates for wind and humidity; (3) the analyzed fields are not required to satisfy certain dynamical constraints. Of these, only (2) is specific to the GFDL data; (1) and (3) are problems for all analyzed data sets. For example, the archived analyses at the European Centre for Medium Range Weather Forecasting do not preserve total mass balance (Trenberth, 1991), although they are derived from a dynamically constrained, data-assimilating model.

4.2. Computational method

The components of the vertically averaged water vapor flux vector, as defined in Eq. (12), are evaluated by trapezoidal rule integration. For each node, the surface

![Graph](image_url)
pressure, $p_s$, is set equal to its mean annual value, with the beginning pressure level for integration selected according to a global topography data set (Oort, 1983, p. 27). For the coastal nodes used in this study, vertical integration generally begins at 1000 mbar.

5. Boundary flux estimation

5.1. Method of flux computation

The closed-path line integral in Eq. (5) may be expressed as a sum of sub-integrals over segments of the boundary:

$$
\oint Q \cdot n \, d\gamma = \sum_{k=1}^{m} \int_{x_k}^{x_{k+1}} Q \cdot n \, d\gamma
$$

in which $x_k$ and $x_{k+1}$ are longitude–latitude points in sequence along the boundary, and $m$ is the number of segments. (For a closed path, $x_{m+1}$ must be the same point as $x_1$.) Each sub-integral has dimension $MT^{-1}$ and represents the time rate of moisture mass flux across that segment of the boundary.

Fig. 3. The total moisture flux (mean motion plus transient eddies) across segments of horizontal regional boundaries, for North and South America. Values shown are the monthly average values for March, April, and May (MAM), 1963–1973.
Fig. 4. The total moisture flux (mean motion plus transient eddies) across segments of horizontal regional boundaries, for North and South America. Values shown are the monthly average values for June, July, and August (JJA), 1963–1973.

The vertically integrated moisture flux vector is decomposed into its mean motion and transient eddy terms,

\[ Q = Q_M + Q_{TE} \]  

where

\[ Q_{TE}(\lambda, \phi) \equiv \int_{0}^{p_2} \frac{q' V'}{g} \, dp \]  

(15a)

and

\[ Q_M(\lambda, \phi) \equiv \int_{0}^{p_1} \frac{q' \tilde{V}}{g} \, dp \]  

(15b)

The boundary sub-integrals in (19) are abbreviated as

\[ f_k \equiv \int_{x_k}^{x_{k+1}} Q \cdot n_\gamma \, d\gamma \]  

(16)
By definition,

$$f_k = f_{M,k} + f_{TE,k}$$  \hfill (17)

Consistent with the definition of $\mathbf{n}_r$ as the outward unit vector, $f_k$ is positive for moisture flux out of the region, and negative for flux into the region.

5.2. Results for North and South America

The coastlines of North and South America were approximated as rectilinear boundaries defined by connecting nodes on the GFDL data grid (spaced at 2.5° latitude and 5° longitude). The sub-integrals in Eq. (16) were evaluated by trapezoidal rule integration over each boundary segment, for each month of the year. Each sub-integral $f_k$ represents the moisture flux across either an east–west or a south–north line segment on the Earth’s surface. The results are shown in the set of Figs. 2–13. In these figures, the magnitude of the moisture flux, $f_k$, across segment $k$ of the boundary is shown by a bar perpendicular to segment $k$, with its base at the segment. The length of the bar is proportional to $f_k$ by the scale indicated, and the bar points into or out of the region according to the direction of net moisture flux across segment $k$. In addition, those bars representing flux into the region ($f_k < 0$) are
cross-hatched to distinguish them further from bars representing flux out of the region ($f_k > 0$).

It is important to note that the bars do not represent vector components; each represents the integral of the normal component of the flux vector over the length of the segment. The graphical scale (in units of kg s$^{-1}$) refers to the bars and has no relevance to the continental outlines, which are not drawn to scale and are, in fact, highly distorted by the cylindrical equidistant projection. Fig. 1 shows a simple example, illustrating the meaning of the bars in the set of figures to follow.

Figs. 2–5 show the total moisture flux ($f$) across the segments of the rectilinear continental regions for the months December, January, and February; March, April, and May; June, July, and August; September, October, and November. Figs. 6–9 present the transient eddy moisture flux ($f_{TE}$) and Figs. 10–13, the mean motion flux ($f_{M}$) for the same seasonal sequence. For Figs. 10–13, values of $f_{TE,k}$ and $f_{M,k}$ were computed, then the values for 3 months were simply averaged. Therefore, the transient eddy values plotted in Figs. 6–9 represent moisture transport by motions on time-scales less than 1 month during the 3-month period, and the mean motion values plotted in Figs. 10–13 represent transport on the time-scale of 1 month.

The total flux of water vapor onto these continents reflects the circulation of the
Fig. 7. The moisture flux by transient eddies across segments of horizontal regional boundaries, for North and South America. Values shown are the monthly average values for March, April, and May (MAM), 1963–1973.

atmosphere and the distribution of temperature at the Earth's surface and in the lower troposphere. The pattern of the equatorial easterly trade winds and the mid-latitude westerlies is clearly reflected in the zonal transport. The tropical warm air moving off the equatorial Atlantic Ocean delivers moisture at an extremely high rate over the northeast coast of South America. The large magnitudes of the segment fluxes both onto and off Patagonia (the southern tip of South America) are attributable to the high velocities in the mid-latitude westerly winds; at these dynamically active latitudes the winds circle the globe almost entirely over ocean, which provides much less resistance to air flow than does land, and at the same time allows development of high relative humidity. These bars are longer by half in the Southern Hemisphere summer (December–February) than in the winter (June–August), owing to the higher saturation vapor pressure of the warmer air. On the other hand, the very low landward values of $f$ along the west coast of South America between 35°S and 20°S are due to low velocity and low vapor content of the air originating from the adjacent waters. The upwelling of cold deep-ocean waters along the subtropical (western) coast of South America leads to a cooling of the overlying air, a generally stable profile and sinking motion. With low values of $q$ and $u$, the zonal flux is minimal. This explanation may be used to hypothesize the mechanism by which El Niños may affect the regional hydrology (Rao and Marques,
Fig. 8. The moisture flux by transient eddies across segments of horizontal regional boundaries, for North and South America. Values shown are the monthly average values for June, July, and August (JJA), 1963–1973.

1984). For this problem, a detailed study of the vapor transport during El Niño and La Niña years is necessary; furthermore, large-scale dynamical motion features of the tropics change during El Niño years and their influence has to be separated.

Two major streams of atmospheric water vapor onto North America are evident: flow from the Pacific Ocean and from the Gulf of Mexico. The latter source is strongest in the Northern Hemisphere spring (March–May). The transport in the mid-latitude zone is westerly, as expected. It is nevertheless surprising that the Gulf of Mexico supply to North America hydrology seasonally either exceeds or is comparable with the Pacific Ocean supply.

The meridional transient eddy flux of moisture onto these continents is poleward in all seasons (Figs. 6–9). In North America, $f_{TE}$ is most prominent off the Gulf of Mexico, where it is greatest in the Northern Hemisphere winter, accounting for more than half the total flux across the boundary between 100°W and 80°W. The increased eddy transport in winter reflects the strong meridional specific humidity gradient between the tropical seas and the cold continental interior, as observed by Hastenrath (1966), who also found that eddy transport accounts for up to 50% of the total flux over the Gulf of Mexico in winter. Rasmusson (1966, 1967, 1968, 1971) also found that transient eddies transport (onshore) large amounts of water vapor from the Gulf; he furthermore demonstrated that the eddy transport is characterized by a
diurnal cycle and that fluctuations with a few days time-scale may be responsible for much of the flux. The use of monthly-average data (and sparse data, especially in South America) may thus provide an underestimate of this transport process. The Global Energy and Water Cycle Experiment (GEWEX) Continental Scale International Project (GCIP) is focused on the Mississippi basin, which is affected by this large transient eddy water vapor transport from the Gulf of Mexico (Kinter and Shukla, 1990).

The transport by mean motion \( f_M \) in Figs. 10–13 can be obtained by subtracting \( f_{TE} \) from \( f \). In particular, \( f_M \) from the Gulf of Mexico is rather large in March–May and June–August, reflecting the intensification of the Atlantic subtropical high in summer and the resulting strong persistent northward flow around its western end. Recalling that \( f_{TE} \) is fairly small across that boundary in June–August, it is interesting to note that although the year-round supply of moisture across the Gulf Coast has its evaporative source in the tropical Atlantic and the Gulf of Mexico, the supply is accomplished by different modes in different seasons. The zonal mean vapor transport off the Gulf of Mexico reverses direction in the course of the year, from westerly in December–February to easterly in June–August. Across a number of boundary segments along the east coast of North America, \( f_M \) and \( f_{TE} \) are of opposite sign; this opposition holds mostly for meridional transport, although in December–February
Fig. 10. The moisture flux by monthly mean motion across segments of horizontal regional boundaries, for North and South America. Values shown are the monthly average values for December, January, and February (DJF), 1963–1973.

(Figs. 6 and 10), a small zonal $f_{TE}$ is opposite $f_M$ between 4°S and 60°N. In South America, meridional values of $f_{TE}$ and $f_M$ tend to be of opposite sign along the southeast coast.

5.3. Estimating annual continental runoff

In Figs. 2–5, the total seasonal influx of atmospheric moisture onto the continent is proportional to the total length of cross-hatched bars minus the total length of white bars. These sums may be used to estimate the annual runoff from each continent by Eq. (8), neglecting the change in storage as discussed above.

For South America, this technique gives an annual runoff of 5000 km$^3$. For comparison, UNESCO (1978) estimated South America's river runoff at 11 760 km$^3$, and Baumgartner and Reichel (1975) estimated this value at 11 100 km$^3$. Owing to the factors discussed under 'Observational data and computational method', particularly the poor radiosonde network density in this region and the underestimation of boundary-layer flow (which is significant in the tropics, where high humidity coincides with low-level easterly surface flow) the present estimate is likely to be low. However, the other researchers' estimates may be high. Both UNESCO and Baumgartner and Reichel gave 6000–7000 km$^3$ as flow from the Amazon River,
whose drainage basin covers 40% of the continent. The flow rates of the other major South American rivers (Orinoco, La Plata, and Parana) are each at least one order of magnitude smaller than that of the Amazon. Furthermore, outside the Amazon Basin, the continent is rather dry. It is unlikely that the rest of South America contributes nearly the equivalent of a second Amazon in runoff (5000 km$^3$). The true value of South America’s runoff probably lies between these extremes, possibly nearer Sellers’ (1965) estimate of 8800 km$^3$.

For North America, more accuracy might be expected, owing to the richer observational network. This study gives an annual continental runoff of 2200 km$^3$, compared with 5800 km$^3$ river runoff (Baumgartner and Reichel, 1975) and 6500 km$^3$ (Sellers, 1965). Again, this study underestimates considerably the annual runoff compared with the other workers’ estimates. Errors related to the analysis scheme (interpolation and the zonal-mean first approximation) may be responsible in this case; for example, analyzed fields that are excessively smoothed in the zonal direction could result in underestimated convergence, because the differences between wind and humidity at the west and east coast at a given latitude might be artificially reduced. The area boundaries of the North American continent are also difficult to define, especially at high latitudes; this factor may result in large discrepancies between atmospheric and land-based estimates of total runoff.
Fig. 12. The moisture flux by monthly mean motion across segments of horizontal regional boundaries, for North and South America. Values shown are the monthly average values for June, July, and August (JJA), 1963–1973.

6. Discussion and conclusions

The vapor transport in the atmospheric branch of the hydrological cycle, its cross-coastal component, and its net convergence over land masses, may be a source of additional constraint and closure for the water balance of continental regions. In this paper, atmospheric water vapor and wind observations are used to estimate the seasonal flux of moisture from the oceans surrounding the Americas to the land masses. The total flux is decomposed into transport by mean motion and by transient eddies.

The vapor flux climatology in the zonal plane follows the mean motion field; the hydrologic cycle is imbedded in the general circulation of the atmosphere. Over North America, including the region of the proposed GEWEX GCIP study area, the Gulf of Mexico is a major source of water for the continental water balance. Transient eddies are a major mode of water vapor transport to this area.

Bar graphs of the boundary flux of $Q$ are a useful technique for the presentation of moisture flux data. Although it would entail substantial interpolation between nodes, the ability to define more realistic continental boundaries would improve the method.

The hydrological cycle is recognized as a central element in studies of climatic fluctuations (Chahine, 1992). The atmospheric water vapor transport and its cross-
coastal components may provide valuable insight and even closure for hydrologic cycle and water-balance investigations over continental land masses. In a pioneering study, Starr and Peixóto (1958) used atmospheric observations to develop important and fascinating hypotheses on the surface hydrology of deserts.

Atmospheric moisture transport calculations can also provide estimates of continental precipitation recycling (Brubaker et al., 1993). Recycling refers to the contribution of land evapotranspiration to the precipitating clouds over the same region; in this manner regional hydrology controls its own forcing — precipitation. Rodriguez-Iturbe et al. (1991) and Entekhabi et al. (1992) demonstrated that such land–atmosphere interaction significantly modifies the nature of fluctuations in the climatic system.

Atmospheric sounding observations provide valuable information on the climatology and nature of the moisture transport regimes. Investigations on the physical mechanisms governing the moisture delivery systems to continental regions will reduce the uncertainty associated with specifying components of the surface hydrologic balance (e.g. precipitation as an unconditional random process). It may also provide closure if the data quality (e.g. sparsity) problems are solved.
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