Estimation of Continental Precipitation Recycling

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ABSTRACT

The total amount of water that precipitates on large continental regions is supplied by two mechanisms: 1) advection from the surrounding areas external to the region and 2) evaporation and transpiration from the land surface within the region. The latter supply mechanism is tantamount to the recycling of precipitation over the continental area. The degree to which regional precipitation is supplied by recycled moisture is a potentially significant climate feedback mechanism and land surface–atmosphere interaction, which may contribute to the persistence and intensification of droughts.

Gridded data on observed wind and humidity in the global atmosphere are used to determine the convergence of atmospheric water vapor over continental regions. A simplified model of the atmospheric moisture over continents and simultaneous estimates of regional precipitation are employed to estimate, for several large continental regions, the fraction of precipitation that is locally derived.

The results indicate that the contribution of regional evaporation to regional precipitation varies substantially with location and season. For the regions studied, the ratio of locally contributed to total monthly precipitation generally lies between 0.10 and 0.30 but is as high as 0.40 in several cases.

1. Introduction

Continental climates exhibit persistence in several distinct regimes, with occasional transition between the regimes. Droughts may persist for prolonged periods followed by abrupt transitions to long periods of moisture abundance without any clear statistical equilibrium in the climate. Physically based models that include feedbacks and interaction in hydrologic and climatologic processes, when forced by relatively small short-term fluctuations, can capture climatic insensitivity and runs of drought periods with persistent anomalous precipitation (Rodriguez-Iturbe et al. 1991a,b; Entekhabi et al. 1992).

Precipitation over a land region is derived from two sources: 1) water vapor adverted into the region by airmass motion, and 2) water vapor supplied by evapotranspiration from the land surface of the region. The extent to which a region's precipitation is supplied by this second mechanism is an indicator of the importance of land-surface processes in the water balance of the region and may also be an indicator of general climatic sensitivity to land-surface change. The recycling process is a potentially significant climate feedback mechanism and land surface–atmosphere interaction, which may contribute to the persistence and intensification of droughts. The objective of the work described here is to obtain quantitative estimates of the degree to which land–atmosphere moisture recycling is active over several large continental regions.

In this study, recycled precipitation is defined as water that evaporates from the land surface within a specified control volume and falls again as precipitation within the same control volume. The remainder of the precipitation falling inside the control volume is considered to be of advective origin, without regard to whether it most recently evaporated from an ocean surface or from a land source outside the control volume.

With recycling defined in this manner, the distinction between recycled and advected precipitation is not necessarily the same as the distinction between precipitation water of terrestrial origin (i.e., most recently evaporated from the land surface) and oceanic origin (i.e., most recently evaporated from the ocean surface). Advected precipitation water is equivalent to oceanic precipitation water if and only if the inflow boundary of the control volume lies along the coastline. For an inland region, it is important to assure compatibility of definitions before comparing different estimates of the sources of the region's precipitation.

Estimates of the ratio of locally derived and advected precipitation water are conditional on the size of the territory that is considered. The ratio is small for regions of limited extent and increases for larger regions. Areas may be isolated based on hydrologic characteristics and response to large-scale climatic forcing. Given the boundaries and extent of a region, the precipitation

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recycling ratio is a measure of the degree of control exerted by land surface hydrology (through evapotranspiration) on the climatic regime of the region. Because the climatic regime is the forcing factor for land-surface hydrology, the recycling ratio is also a measure of land surface-atmosphere feedback.

2. Background

Until the late 1930s, according to Benton et al. (1950), it was generally believed that the water precipitated over the continents was directly derived from moisture evaporated from the continents. Under this view, the flux of water in the atmosphere from the oceans to the continents was assumed to be only that required to replenish the loss of water from the continents by runoff. Apparently, there was no conception that vast quantities of incoming water vapor from the oceans could pass over continental regions without precipitating or that water vapor that evaporated from the continent could be carried away by the wind as atmospheric runoff.

Benton et al. critiqued the theory that precipitation is largely land derived and described developments leading to a more correct description of the role of the atmosphere in the hydrologic cycle. They emphasized two concepts. First, vertical motions in the atmosphere are necessary in order to produce significant precipitation; therefore, simply increasing the water vapor in the air over a region does not necessarily increase precipitation. Second, the atmosphere is in continuous motion, carrying very large quantities of water vapor across the continents from the oceans; moisture added to the atmosphere by evaporation and transpiration may travel hundreds or thousands of miles before being reprecipitated.

In the same paper, Benton et al. estimated the relative contributions of the sources of precipitation over the Mississippi watershed. They distinguished between precipitation events from, and evaporation into, maritime and continental air masses and estimated that not more than 10% of the total precipitation of the Mississippi watershed is moisture having a land source within the watershed, while at least 90% is external in origin.

McDonald (1962) argued against what he called “the evaporation-precipitation fallacy,” that is, the idea that local water shortages might be alleviated by creating open areas from which water could evaporate and enhance local precipitation. In arguments parallel to those of Benton et al., he attributed the fallacy to misconceptions concerning the magnitude of and distance scales involved in atmospheric water vapor transport.

Budyko (1974, pp. 239–243) used a simple one-dimensional model (which is described in section 3) to estimate, for the European part of the former Soviet Union, the contributions of locally evaporated and advected moisture to regional precipitation. Budyko defined $\beta$ as the ratio of total precipitation to precipitation due to moisture advected from outside the region. On an annual basis, only about 10% of the precipitation in the former European Soviet Union was of local origin, according to Budyko’s estimate; on a monthly basis, the estimated contribution of locally evaporated water ranged from 4% in October to 18% in April and May. Budyko stated, “Even on the most extensive continents, where the relative role of local evaporation is the greatest, as calculations show, the main portion of precipitation is formed from water vapor of external origin, not local.” Shiklomanov (1989) used aerological data to enhance Budyko’s model and obtained monthly estimates of the local fraction of precipitation in the former European Soviet Union ranging from 0.1% in January and February to 20% in June.

The significance of the oceans as sources of precipitation to the continents is no longer disputed. This understanding of the atmospheric branch of the hydrologic cycle has been strengthened by analysis and observation (Sutcliffe 1956). During the last half of this century, however, studies have shown that land regions can also be significant sources of water vapor to the atmosphere (Rasmussen 1967, 1968, 1971; Starr and Peixoto 1958; Peixoto and Oort 1983).

Thornthwaite (1946) observed that the average July evapotranspiration in the eastern United States (between 5.5 and 6.0 inches) exceeded the July evaporation from the water surface in the Gulf of Mexico and the Caribbean (about 4.0 inches). Benton and Estoque (1954) showed that the North American continent, as a whole, is a source of moisture to the atmosphere in summer (May–August) and a sink during the rest of the year. Starr and Peixoto (1958) observed strong maxima of vapor flux divergence over three separate arid regions, implying a large excess of evaporation over precipitation in these areas. Stidd (1967) stated:

If we consider that in most areas the soil will be drier in the fall than in the spring and that the difference will represent a net loss of moisture to the atmosphere (in the absence of runoff), it is not hard to understand that land areas in general furnish a source of moisture to the atmosphere in summer and that the oceans provide a sink, so that, in effect, the hydrologic cycle is reversed.

If the land surface acts as a significant source of moisture to the atmosphere during certain seasons, then it is reasonable to expect that the balance of locally evaporated versus advected precipitation might be shifted during those seasons (assuming the presence of the dynamical processes required to produce precipitation at all). Furthermore, it is reasonable to expect that alterations in the storage properties or to the recharge-discharge cycle of the land surface and subsurface might affect the precipitation regime.

Stidd (1967) studied local modifications in climate following large-scale irrigation development in the Co-
lumbia River Basin. His analysis showed an increase in July and August precipitation, extending to several thousand square kilometers around the project. Fowler and Helvey (1974) reexamined the topic by alternate procedures and concluded that a large increase in precipitation due to nearby irrigation was improbable and that Stidd’s claim of a significant increase in precipitation did not appear sound. Stidd (1975) replied with statistical analyses supporting his earlier claim. He argued that Fowler and Helvey examined the problem on too small a scale, citing the fact that the irrigation project lies near the center of the drainage basin and that the additional moisture added to the air by irrigation would be expected to precipitate not immediately adjacent to the irrigation project but in the foothills downwind as the moist surface air is carried over the surrounding mountains. This explanation is in agreement with the statements of Benton et al. (1950) and McDonald (1962) that increasing the water vapor content of the air will increase precipitation only if an uplift mechanism exists.

Natural or anthropogenic changes that enhance (or inhibit) convection could alter the contribution of local moisture to precipitation. Dettwiller and Changnon (1976) found upward trends in the warm season rainfall at Paris, St. Louis, and Chicago, and suggested that the urban heat island contributes to larger and more intense shower clouds. Schickedanz (1976, pp. 99–100) found evidence for irrigation-related increases in summer rainfall in several areas of the United States Great Plains; he hypothesized that

. . . any increased rainfall does not come directly from the increased atmospheric moisture alone, but by thermodynamic and physical side effects produced by the presence of a cool, moist dome over the irrigated area.

Anthes (1984) speculated that if human activities can inadvertently affect precipitation, perhaps humans could make intentional changes in surface characteristics so as to modify precipitation in a useful way. He proposed planting bands of vegetation in semiarid regions, hypothesizing that the bands could increase convective precipitation through three mechanisms: increased low-level moist static energy, the generation of mesoscale circulations, and increased atmospheric water vapor.

In a study of the water vapor budget over central North America in summer 1979, Zangvil et al. (1993) concluded that local evapotranspiration is the main source of moisture for daytime precipitation. Portis et al. (1991) studied the 1976 and 1979 growing seasons for the same study region and found that local evaporation is a major source of moisture for small rainfall events, whereas larger events draw from both local evaporation and horizontal atmospheric water vapor flux convergence. Zangvil et al. (1992) similarly establish the relative importance of locally evaporated moisture for precipitation over the eastern Mediterranean region.

Lettau et al. (1979) used climatonomic methods (Lettau 1969) to quantify precipitation recycling in the Amazon River Basin in an effort to assess the possible climatic consequences of large-scale deforestation in the region. Among their results were estimates of the ratio of total regional precipitation to the regional flushing of precipitable water of direct oceanic origin (γ). They analyzed six regions of 5° longitude in width; over the six regions, γ increased with distance downwind of the coast, reaching a maximum value of 1.884. This value of γ indicates that 47% of the precipitation falling on the subregion centered at 75°W has been most recently evaporated from the continent, both upwind of and within that particular subregion. (It should be noted that Lettau et al.’s comparison of γ with Budyko’s β for this inland region is not valid; β compares total precipitation to that of advective origin, where the advective term includes moisture evaporated from the land lying between the coast and the inland region’s boundary in addition to water of oceanic origin, whereas γ compares total precipitation to that of oceanic origin only.)

Salati et al. (1979) studied the inland gradient of the oxygen-18 content of precipitation in the Amazon Basin. For yearly averaged station data, the gradient d(δ18O)/dx is much smaller than in other continental areas. The small inland decrease in isotopic content of precipitation is an indication that significant amounts of moisture are added to the air mass as it passes over the region, and that this evaporated moisture is important in precipitation falling on the region; this effect varies with the season and the location. Salati et al. did not give a numerical estimate of precipitation recycling, but they cited a study by Marques et al. (1977) that found that inflowing moisture accounted for only 52% of the precipitation between Belém and Manaus in 1972.

In atmospheric general circulation models (GCMs), water can be tagged according to its evaporation site and traced in order to determine the relative contributions of different evaporative sources to a region’s precipitation. Jousame et al. (1986) conducted such an experiment for the month of July using the Laboratoire de Météorologie Dynamique GCM; they produced global charts showing the influence of ocean- and continental source regions in continental precipitation. Koster et al. (1986) used the NASA/Goddard Institute for Space Studies’ GCM in a similar experiment for all four seasons. Koster et al. (1988) also studied the inverse problem, that is, the characteristic distances and directions traveled by water evaporating from a source region. Shukla and Mintz (1982) conducted a sensitivity experiment with the NASA/Goddard Laboratory for Atmospheric Sciences’ GCM, in which two different constraints were placed upon the land surface evapotranspiration: in the first case, no
evapotranspiration was allowed, and in the second case, evapotranspiration was set equal to the model-calculated potential evapotranspiration. They found that “land-surface evapotranspiration has a large influence on the precipitation, temperature, and motion fields of the atmosphere.”

The importance of land regions in supplying moisture to the atmosphere has been well demonstrated. There are indications that land-evaporated moisture supplies a large fraction of the precipitation for some continental regions. Questions remain concerning the contribution of local evaporation to local precipitation. Several quantitative estimates have been made, and there is general agreement on the physical factors necessary for such a recycling process to occur.

Recent research by Rodríguez-Iturbe et al. (1991a, 1991b) and by Entekhabi et al. (1992) shows that the close coupling between the land surface and the atmosphere in continental-type climates helps to explain the statistical structures exhibited by climatic variables. In their work, the coupling is explicitly represented by including the precipitation recycling mechanism in the land-surface water balance equation. The resulting functional dependence of precipitation on soil moisture, together with the stochastic nature of soil and climate parameters, results in a bimodal probability distribution of soil moisture states corresponding to drought and pluvial conditions. Further, the dynamics of the soil moisture equation can exhibit fixed point, limit cycle, and chaotic behavior.

The importance of precipitation recycling as an index of land surface–atmosphere coupling is a compelling motivation for further study of this process and for quantification of the degree to which it is active over a variety of continental regions.

3. Recycling model

The sources of precipitation water are indicated conceptually in Fig. 1, which shows a simplified model of atmospheric moisture fluxes over a land region. The horizontal arrows indicate the advective flux of water vapor into and out of the atmospheric control volume; $W$ is the amount of water vapor contained in the air as it moves through the control volume; $E$ is the net evapotranspiration from the underlying land surface; and $P$ is the net precipitation onto the land surface. The two sources of precipitation are indicated by the two small branches that join to form the larger arrow labeled $P$: $P_m$ is precipitation of local (evaporative) origin and $P_a$ is that of advective origin. The arrow labeled $E$ splits into two branches, indicating that a certain fraction of the locally evaporated or transpired water is not returned to the land surface as precipitation but joins the atmospheric vapor reservoir and is advected out of the control volume.

Budyko (1974) considered a land region traversed by a length scale $l$ (an air streamtube of streamline length $l$), average precipitation $P$, and average evapotranspiration $E$. The average precipitation (Fig. 1) is composed of an advective portion ($P_a$) and a local evaporative portion ($P_m$), that is,

$$P = P_a + P_m.$$  \hspace{1cm} (1)

The water vapor content of the air moving across the region is also composed of an advected and an evaporated portion. Air enters the region at velocity $u$ normal to the intersection of the region boundary and streamtube and with moisture content $w$ (precipitable water). The velocity $u$ is vertically averaged with specific humidity weighting so that the product $w u$ is the rate of water vapor flux into the region. The vertical flux quantities $P_a$, $P_m$, and $E$ are treated as constants equal to their average values; therefore, the locally evaporated moisture content of the air increases linearly and the advected moisture content decreases linearly as the air moves across the region. It follows that the average horizontal flux of advected moisture over the region is

$$Q_a = w u - \frac{1}{2} P_a$$  \hspace{1cm} (2)

and the average horizontal flux of locally evaporated moisture is

$$Q_m = \frac{1}{2} (E - P_m).$$  \hspace{1cm} (3)

The atmosphere is assumed to be fully mixed so that the ratio of advected to locally evaporated water falling as precipitation is equal to the ratio of advected to evaporated moisture present in the air. Precipitation draws from each reservoir of moisture in proportion to the abundance of moisture due to each source. This is modeled as

$$\frac{P_a}{P_m} = \frac{Q_a}{Q_m} = \frac{wu - \frac{1}{2} P_a}{\frac{1}{2} (E - P_m)}.$$  \hspace{1cm} (4)

\hspace{1.5cm} FIG. 1. Conceptual model of the atmospheric moisture fluxes over a land region. Terms $P_m$ and $P_a$ are precipitation of local evaporative and advective origin, respectively.
Equations (4) and (1) are solved for the ratio of total to advected precipitation, which is Budyko's recycling coefficient ($\beta$):

$$\beta = \frac{P}{P_a} = 1 + \frac{El}{2wu}. \quad (5)$$

The second term on the right-hand side of (5) is the ratio of locally evaporated to advected precipitation. In this model, the relative contribution of locally evaporated water is directly proportional to the evaporation rate and the length scale of the streamline traversing the region and is inversely proportional to the rate at which external moisture enters the region. The inverse of $\beta$ is the fraction of precipitation due to advective origin, that is,

$$\frac{P_a}{P} = \frac{1}{\beta}, \quad (6)$$

and the fraction of precipitation due to local origin is

$$\frac{P_m}{P} = 1 - \frac{1}{\beta} = \left[1 + \frac{wu}{El}\right]^{-1}. \quad (7)$$

In this study, Budyko's arguments are extended to a two-dimensional land region, with moisture influx and efflux through the sides of an atmospheric control volume. The boundary of the region consists of a segment or set of segments ($\lambda_{in}$) across which the atmospheric moisture flux is inward and a segment or set of segments ($\lambda_{out}$) across which the flux is outward.

The integrals of the vertically integrated water vapor flux vector, $\vec{Q}$ (Peixoto and Oort 1983), over $\lambda_{in}$ and $\lambda_{out}$ are, respectively, the inflow and outflow of atmospheric moisture through the sides of the control volume:

$$F^+ = -\int_{\lambda_{in}} \vec{Q} \cdot \hat{n}_s d\lambda$$

and

$$F^- = \int_{\lambda_{out}} \vec{Q} \cdot \hat{n}_s d\lambda, \quad (8b)$$

in which $\hat{n}_s$ is the outward unit normal vector.

By definition, $F^+$ contains only advective moisture, and $F^-$ contains the advected moisture that remains after $P_a$ is removed, as well as moisture of local origin:

$$F^- = (F^+ - P_a A) + (E - P_m) A, \quad (9)$$

in which $A$ is the area of the region and $P_a$, $P_m$, and $E$ are the rates of advective and local precipitation and evaporation per unit area. If the average horizontal flux of advected moisture over the region is taken as the arithmetic mean of the incoming and outgoing advective moisture, then

$$Q_a = \frac{F^+ + (F^+ - P_a A)}{2} = \frac{F^+ - P_a A}{2}. \quad (10)$$

Likewise, the average horizontal flux of locally evaporated moisture is the mean of the incoming and outgoing local portions,

$$Q_m = \frac{0 + (E - P_m) A}{2} = \frac{(E - P_m) A}{2}. \quad (11)$$

Invoking, as before, the assumption of a well-mixed atmosphere, the recycling coefficient is

$$\frac{P}{P_a} = \beta = 1 + \frac{EA}{2F^+}. \quad (12)$$

Equation (12) is identical to (5), with $F^+$ replacing $wu$ and $A$ replacing $l$. Brubaker et al. (1991) also propose an alternate approach to using vapor flux data in estimating recycling by (5). In that approach, effective values of $l$ and $wu$ are computed for an area based on the average magnitude and direction of the vapor flux over the region.

In Budyko's model, it is assumed that the mean evaporation and mean total precipitation rates are characteristic of all points within the region. Equations (10) and (11) further assume that the concentrations of advected and locally derived moisture change linearly as the air mass traverses the territory, and $P_m/P$ is computed from the mean values of these linearly varying quantities. Even with spatially homogeneous precipitation and evaporation rates over the region, the linear model is only a first-order approximation to the changing composition of the moisture reservoir. As the averaging distance in the linear model is reduced, a differential equation governing the spatially varying recycled precipitation fraction emerges. This more general model is derived by Drozdov and Grigor'eva (1965); the final expression appears below. In this model, the concentration of water vapor of each origin is allowed to vary nonlinearly as the reservoir is depleted by precipitation and replenished by evaporation. The region's recycled precipitation fraction is computed as the mean value of the spatially varying $P_m/P$, rather than as a ratio of mean values of the respective moisture concentrations, as in the Budyko model. The resulting regional recycling ratio is (Drozdov and Grigor'eva 1965)

$$\frac{\bar{P}_m}{P} = 1 - A^* + \Lambda^* \left[\frac{A^*}{\Lambda^* + T - 1}\right]^{1/\tau}, \quad (13)$$

where the overbar denotes a spatial average and the following dimensionless variables have been defined: the ratio of moisture influx to total regional precipitation,

$$\Lambda^* = \frac{wu}{Pl}, \quad (14)$$

and the ratio of average evaporation to average precipitation,

$$T = \frac{E}{P}. \quad (15)$$
In terms of these dimensionless variables, (7) is

\[ \frac{P_m}{P} = \left[ 1 + 2 A \lambda^2 \right]^{-1}. \tag{16} \]

While (13) is more rigorously developed, the simpler linear model in (7) is more easily incorporated into analytical studies of large-scale water balance (Rodríguez-Iturbe et al. 1991a; Entekhabi et al. 1992). For this study, the simpler model is relied on, but in section 6, the estimates of recycling are compared for the study regions based on both (7) and (13) in order to assess the magnitude of error.

4. Observational data and estimation techniques

a. Aerological data

The humidity and wind data used in this study are from a dataset provided by the Geophysical Fluid Dynamics Laboratory (GFDL) of NOAA through the courtesy of Dr. Abraham Oort. The GFDL dataset consists of 10 years of observations transformed into a gridded form (Oort 1983). The measurements were taken during the period May 1963 through April 1973. For the years 1963–68, the station-based, upper-air observations were once daily, at 0000 UTC, although a small number of 1200 UTC observations were included. For the years 1968–73, both 0000 and 1200 UTC 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 is described by Oort (1983).

Values of specific humidity \( \bar{q} \), zonal and meridional wind \( \bar{u}, \bar{v} \), and zonal and meridional transient eddy fluxes \( \bar{q}u', \bar{q}v' \) are given at each node of a 2.5° latitude by 5° longitude grid and at each of 11 pressure levels; the overbar denotes monthly averages.

b. Precipitation data

The precipitation data used in this study are from the World Monthly Surface Station Climatology (and associated datasets) distributed by the National Center for Atmospheric Research (NCAR) in Boulder, Colorado. Monthly station precipitation data are given for over 3900 different World Meteorological Organization stations through 1987, with data for some stations going as far back as the mid-1700s. From this dataset, the precipitation records for May 1963 through April 1973 were selected in order to obtain precipitation data that are temporally compatible with the GFDL aerological data.

c. Estimation techniques

The zonal and meridional components of the vertically integrated water vapor flux vector, \( \bar{Q} \), were evaluated from the GFDL data by trapezoidal rule integration, summing the contributions of mean flow and transient eddys (Brubaker et al. 1991). For each node, the surface pressure was set equal to its mean annual value, with the beginning pressure level for integration selected according to the surface pressure climatology provided by Oort (1983).

Evapotranspiration was computed from mass balance by neglecting changes in atmospheric moisture, that is,

\[ E \approx P + \frac{1}{A} \int \bar{Q} \cdot \bar{n} d\lambda = P + \nabla \cdot \bar{Q}, \tag{17} \]

in which \( E, P, \) and \( \nabla \cdot \bar{Q} \) refer to monthly regional mean values. The rationale for estimating evapotranspiration as a residual is that the atmospheric water balance reflects actual evapotranspiration, whereas pan evaporation data estimate potential evapotranspiration, and model equations for actual evapotranspiration generally require estimates of soil and vegetation parameters that would be difficult to obtain reliably for all regions and on continental scales.

The divergence term, \( \nabla \cdot \bar{Q} \), was computed by trapezoidal rule integration of the flux integral in (17). Regional average precipitation, \( P \), was computed by the Thiessen polygon method (Bras 1990) from the NCAR station data as follows: First, \( P \) was computed for each month from May 1963 through April 1973, and then the ten January values, ten February values, and so on, were averaged to give a set of monthly average precipitation figures representative of the 10-yr period.

5. Results

Regions were selected for analysis on the basis of hydrological interest and the existence of previous recycling estimates in the literature. The four regions selected are shown in Fig. 2. The North American region lies east of the Continental Divide and within the Mississippi Basin (except for the northeast corner, which lies in the Great Lakes drainage). Several estimates of local recycling or of the land-evaporated fraction of precipitation have been made for this region, which receives atmospheric moisture from the Pacific Ocean and the Gulf of Mexico. The Eurasian region corresponds roughly to the European part of the former Soviet Union, a region for which monthly water budgets and recycling ratios have been reported by Budyko (1974) and Shiklomanov (1989). The South American region contains most of the Amazon river system and rain forest. This region is currently the focus of intense hydroclimatological interest due to the threat of massive deforestation, with possible ramifications for the region’s climate and the climate system of the earth as a whole (Nobre et al. 1991). The African region lies south of the Sahara desert and includes most of the drainage basin of the Niger River. This region is threatened by drought and desertification, and its annual
cycle of atmospheric moisture supply exhibits unique characteristics (Nicholson 1989).

a. Eurasian region

The results for the Eurasian region are presented in Fig. 3. The largest $P_{m}/P$ occurs in June when the recycling rate is 0.31, and the lowest is 0.0 in February. Table 1 lists the various terms used in calculating $P_{m}/P$. The annual sequence is as expected for atmospheric moisture convergence, precipitation, and evaporation: the summer months exhibit the period of greatest precipitation, evaporation, and atmospheric convergence rates.

In Fig. 3, our $P_{m}/P$ results are compared to the work of Budyko (1974) and Shiklomanov (1989). This study is in general agreement with their results in all but the summer months, lending further support to the claim

![Graph](image)

**Fig. 3.** Eurasian region: the ratio of locally evaporated to total precipitation, as estimated in this study using (16), and as reported by Budyko (1974) and Shiklomanov (1989).

<table>
<thead>
<tr>
<th>Month</th>
<th>$-\text{Div}Q$ (mm/mo)</th>
<th>$P$ (mm/mo)</th>
<th>$E$ (mm/mo)</th>
<th>$\frac{F_{+}}{A}$ (mm/mo)</th>
<th>$P_{m}/P$</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>24</td>
<td>35</td>
<td>11</td>
<td>79</td>
<td>0.07</td>
</tr>
<tr>
<td>February</td>
<td>36</td>
<td>36</td>
<td>-1</td>
<td>106</td>
<td>-0.00</td>
</tr>
<tr>
<td>March</td>
<td>24</td>
<td>32</td>
<td>8</td>
<td>95</td>
<td>0.04</td>
</tr>
<tr>
<td>April</td>
<td>17</td>
<td>36</td>
<td>19</td>
<td>130</td>
<td>0.07</td>
</tr>
<tr>
<td>May</td>
<td>2</td>
<td>45</td>
<td>42</td>
<td>118</td>
<td>0.15</td>
</tr>
<tr>
<td>June</td>
<td>-23</td>
<td>56</td>
<td>79</td>
<td>88</td>
<td>0.31</td>
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<tr>
<td>July</td>
<td>-18</td>
<td>63</td>
<td>81</td>
<td>115</td>
<td>0.26</td>
</tr>
<tr>
<td>August</td>
<td>-18</td>
<td>56</td>
<td>73</td>
<td>126</td>
<td>0.23</td>
</tr>
<tr>
<td>September</td>
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<td>44</td>
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<td>169</td>
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<tr>
<td>October</td>
<td>18</td>
<td>43</td>
<td>24</td>
<td>188</td>
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<tr>
<td>November</td>
<td>32</td>
<td>50</td>
<td>18</td>
<td>212</td>
<td>0.04</td>
</tr>
<tr>
<td>December</td>
<td>38</td>
<td>46</td>
<td>8</td>
<td>123</td>
<td>0.03</td>
</tr>
</tbody>
</table>

**Table 1.** Eurasian region hydroclimatology. Components in the estimation of $P_{m}/P$. 

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that external moisture sources dominate in supplying precipitation to this region for much of the year. The differences in the summer months may simply be due to use of different study regions (the region considered in this study does not include the entire former European Soviet Union, which was the study region for the former work) and different datasets.

b. North American region

The results for the North American region are presented in Fig. 4. The estimates are based on the values listed in Table 2. Referring to Fig. 4, $P_m/P$ lies between 0.15 and 0.34, with maxima in July and October.

For two GCM grid squares roughly corresponding to the region considered here, Koster et al.'s (1986) GCM tracer test showed that the evapotranspiration from land regions contributes significantly to the region's precipitation. Because Koster et al. (1986) consider the entire North American continent as the source area for the study region, their estimate of continental moisture contributions to the region's precipitation is considerably higher than those estimated here based on contributions from within the study region alone. Koster et al.'s (1986) study showed that the percentage contributions of the North American source region to local precipitation were 36.4% and 41.6% in winter and 80.1% and 61.9% in summer for the western and eastern grid squares, respectively. The magnitude of the estimates in this study are roughly one-half due to the definition of the source area, but the seasonal patterns are similar.

The recycled fraction of precipitation reported here is considerably greater than the Benton et al. (1950) estimate that 10% or less of precipitation on the entire Mississippi watershed (a larger territory than this study's North American region) has its evaporative source within the watershed. Their estimate was based on distinguishing between continental and maritime air masses, with the assumption that evapotranspiration is greater into the former and precipitation greater from the latter. The methods in the present study lump all atmospheric motion during one month into the same averaging process, essentially ignoring any correlation between evapotranspiration or precipitation and air-mass source; section 6 discusses how this averaging process leads to an overestimation of the recycling ratio. Benton et al.'s (1950) set of assumptions is more physically realistic. Where adequate data are available, the Budyko model estimate would be improved by averaging over shorter time periods in order to capture more adequately the correlations between the physical processes.

When averaged over a period (annual) such that atmospheric and subsurface moisture storage change terms vanish, the convergence of water vapor into the atmosphere overlying a region must balance the surface and subsurface runoff loss (divergence) from the region (Lufkin 1959). The atmospheric water vapor divergence estimates for the North American region in Table 2 indicate that there is a net annual divergence over the region. This implies that the region as delineated (Fig. 2) has a net inflow of river and subsurface water. The rivers and subsurface flow fields across this region experience a net loss due to the dominance of evaporation rates in excess of infiltration into the soil. The seasonal cycle in the atmospheric water vapor divergence (Table 2) is due to both this net surface–subsurface water convergence and the seasonal regime in soil moisture storage change.

c. South American region

The results for the South American region are presented in Fig. 5. The estimates are based on the values listed in Table 3. Referring to Fig. 5, $P_m/P$ reaches a maximum of 0.32 in December and a minimum of 0.14 in June. Divergence calculations in Table 3 show the region to be a sink of atmospheric moisture throughout the year but particularly in the first four months of the year, corresponding to the southernmost excursion of the intertropical convergence zone (ITCZ) during the Southern Hemisphere (SH) summer. The

![Fig. 4. North American region: the ratio of locally evaporated to total precipitation.](image-url)
assimilated data from the European Center for Medium-Range Weather Forecasts and a spatially distributed recycling model. Their estimate that 25% of the basin’s annual rainfall is recycled is in good agreement with this study’s results.

d. African region

The results for the African region are presented in Fig. 6. The estimates are based on the values listed in Table 4. Referring to Fig. 6, two peaks appear in the annual march of $P_m/P$: 0.41 in February–March and 0.48 in August. The February–March peak corresponds to fairly high $E$ and low $P$ in those months, while the July–August peak corresponds to a season of high $E$ and high $P$ (Table 4).

The estimated monthly rates of moisture convergence (Table 4) indicate that, except during May and June, the territory defined as the African region is a net source of moisture to the atmosphere. The defined region includes the expansive lower Niger marsh lands where a large volume of the water drained from the Guinea highlands and the tropical upper Niger evaporates into the semi-arid surroundings. There is considerable excess of evaporation over precipitation in this region during the months when the highland river discharge spreads over the expansive marsh land and evaporates into the atmosphere. During May and June when the river experiences its lowest monthly flow rates (and even dries completely in some years), the marsh lands disappear and the regional evaporation is accordingly reduced. During this season, the precipitation water is mostly derived from advected moisture, as evident in Fig. 6. Furthermore, there will be a net regional convergence of moisture ($\nabla \cdot \bar{Q} > 0$) during these months when the sources of water vapor lie outside the defined territory.

The significant contribution to precipitation of locally evaporated moisture in July and August can readily be inferred from Table 4. Precipitation is at its maximum during these months, yet the influx of atmospheric water vapor is much less than in June and

<table>
<thead>
<tr>
<th>Month</th>
<th>$-\text{Div} Q$ (mm/mo)</th>
<th>$P$ (mm/mo)</th>
<th>$E$ (mm/mo)</th>
<th>$F_m/\bar{Q}$ (mm/mo)</th>
<th>$P_m/P$</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>104</td>
<td>256</td>
<td>152</td>
<td>205</td>
<td>0.27</td>
</tr>
<tr>
<td>February</td>
<td>117</td>
<td>249</td>
<td>132</td>
<td>223</td>
<td>0.23</td>
</tr>
<tr>
<td>March</td>
<td>113</td>
<td>288</td>
<td>175</td>
<td>212</td>
<td>0.29</td>
</tr>
<tr>
<td>April</td>
<td>110</td>
<td>250</td>
<td>140</td>
<td>197</td>
<td>0.26</td>
</tr>
<tr>
<td>May</td>
<td>59</td>
<td>183</td>
<td>124</td>
<td>204</td>
<td>0.24</td>
</tr>
<tr>
<td>June</td>
<td>51</td>
<td>120</td>
<td>69</td>
<td>219</td>
<td>0.14</td>
</tr>
<tr>
<td>July</td>
<td>18</td>
<td>95</td>
<td>77</td>
<td>177</td>
<td>0.18</td>
</tr>
<tr>
<td>August</td>
<td>17</td>
<td>73</td>
<td>56</td>
<td>155</td>
<td>0.15</td>
</tr>
<tr>
<td>September</td>
<td>33</td>
<td>100</td>
<td>67</td>
<td>174</td>
<td>0.16</td>
</tr>
<tr>
<td>October</td>
<td>30</td>
<td>145</td>
<td>115</td>
<td>144</td>
<td>0.29</td>
</tr>
<tr>
<td>November</td>
<td>56</td>
<td>187</td>
<td>131</td>
<td>147</td>
<td>0.51</td>
</tr>
<tr>
<td>December</td>
<td>54</td>
<td>213</td>
<td>159</td>
<td>169</td>
<td>0.32</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Month</th>
<th>$-\text{Div}Q$ (mm/mo)</th>
<th>$P$ (mm/mo)</th>
<th>$E$ (mm/mo)</th>
<th>$F^+ A$ (mm/mo)</th>
<th>$P_m/P$</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>-36</td>
<td>2</td>
<td>38</td>
<td>116</td>
<td>0.14</td>
</tr>
<tr>
<td>February</td>
<td>-94</td>
<td>6</td>
<td>100</td>
<td>73</td>
<td>0.41</td>
</tr>
<tr>
<td>March</td>
<td>-77</td>
<td>23</td>
<td>100</td>
<td>72</td>
<td>0.41</td>
</tr>
<tr>
<td>April</td>
<td>-24</td>
<td>51</td>
<td>75</td>
<td>100</td>
<td>0.27</td>
</tr>
<tr>
<td>May</td>
<td>14</td>
<td>88</td>
<td>74</td>
<td>146</td>
<td>0.20</td>
</tr>
<tr>
<td>June</td>
<td>74</td>
<td>130</td>
<td>56</td>
<td>265</td>
<td>0.10</td>
</tr>
<tr>
<td>July</td>
<td>-27</td>
<td>181</td>
<td>208</td>
<td>116</td>
<td>0.47</td>
</tr>
<tr>
<td>August</td>
<td>1</td>
<td>234</td>
<td>232</td>
<td>124</td>
<td>0.48</td>
</tr>
<tr>
<td>September</td>
<td>-71</td>
<td>173</td>
<td>244</td>
<td>188</td>
<td>0.39</td>
</tr>
<tr>
<td>October</td>
<td>-71</td>
<td>67</td>
<td>137</td>
<td>202</td>
<td>0.25</td>
</tr>
<tr>
<td>November</td>
<td>-114</td>
<td>9</td>
<td>123</td>
<td>120</td>
<td>0.34</td>
</tr>
<tr>
<td>December</td>
<td>-72</td>
<td>5</td>
<td>77</td>
<td>105</td>
<td>0.27</td>
</tr>
</tbody>
</table>

September; much of the precipitation must be supplied by local evapotranspiration.

There may be considerable errors in the mass-flux and specific humidity profiles over regions such as West Africa where the analyzed fields are based on sparse soundings. The convergence and vapor flux climatology based on these fields are generally consistent (at least in sign) with the hydrology of the region as discussed above. Nevertheless, significant error may exist in the magnitude of the estimates. (The potential sources of errors in the analyses are discussed in section 6).

6. Discussion

The precipitation recycling estimates presented in section 5 are based on an extension to two dimensions of Budyko’s linear model of the water vapor content of the air moving over a land region. The relaxation of the linear model results in the more detailed model outlined in section 3.

To compare the measure defined by (13) with the considerably simpler (16), the percentage difference between the two is taken and the under- or overestimation is plotted across a feasible parameter space. The fraction of the total rainfall derived from local recycling ($P_m/P$) is estimated with both (13) and (16). Both measures of the recycling rate are completely defined by the two dimensionless climatic numbers $\Lambda^*$ and $\Sigma$. The percentage differences between the former and the latter estimates of $P_m/P$ are plotted as contours over the ($\Lambda^*$, $\Sigma$) space in Fig. 7; also included are monthly ($\Lambda^*$, $\Sigma$) estimates for all four study regions. It is apparent that the estimation of precipitation recycling according to the linear Budyko model in (16) underestimates the rate when compared to the estimation by (13). In Fig. 7, it is seen that the underestimation is rather small, generally by a factor less than 10% unless both $\Lambda^*$ and $\Sigma$ are rather small. The underestimation may rise by up to 30% when the region considered is large (low $\Lambda^*$ value) and the linear averaging inherent in the Budyko model becomes restrictive.

The critical assumption of both models is that locally evaporated and advected water vapor molecules condense and fall as precipitation proportionally to their respective fraction of the total number of vapor molecules contained in the entire atmospheric column. Stidd (1967) argued against the fully mixed assumption: “since precipitation is normally associated with rising currents of air, it is logical to suppose that most of the excess moisture associated with a storm has come recently from a layer of air close to the ground.” The large fraction of locally evaporated moisture in the lower layers then would imply a large fraction of local moisture in convective precipitation. On the other hand, if nonprecipitating convection mixes the troposphere vertically (Paluch 1979; Taylor and Baker 1991), the assumption of a well-mixed atmospheric moisture reservoir is strengthened for regions and seasons in which convection is active.

Another assumption in the recycling estimation methods is the use of areally averaged precipitation over arbitrarily defined regions. Precipitation is a highly variable process in space and time. Although time-averaging smooths the series at a point in space, substantial spatial variation may exist, depending upon the topography (and other surface properties) and the atmospheric-circulation characteristics of a region. As an extreme example, the model is invalid if all of a region’s precipitation falls near the upwind boundary before any locally evaporated moisture has been added to the air. In such a situation, $P_m/P$ would be overes-

![Fig. 7. Percentage difference in the estimation of $P_m/P$ made by (13) and (16) for all four study regions. Contour lines represent the percentage factor by which (16) underestimates the value when compared to (13). The monthly estimates for the study regions are symbolized by squares for the North American, diamonds for the South American, circles for the African, and triangles for the Eurasian regions.](image-url)
timated because the model assumes that the precipitation rate is spatially homogeneous over the region and therefore that the local evaporation is the source of some of the precipitating water.

Similarly, the use in this study of time-averaged data introduces errors that will result in the overestimation of the recycling ratio. The calculations are based on monthly average values of water vapor flux, as if a steady circulation existed throughout each month. Correlations among the time-varying direction and water content of incoming air masses and precipitation and evaporation rates are neglected. To illustrate the problem, consider the following extreme situation: during the first half of the month, moist air flows across the territory releasing a steady uniform precipitation (without evaporation). During the last half of the month, the direction of flow is reversed and evaporation occurs at the same steady uniform rate as precipitation during the first half of the month. In this case, $F^+ / \Delta$ would be zero, implying total recycling, whereas in fact there is no recycling. (It should be noted that although spatial averaging is intrinsic to the model, time averaging is not. The method could be applied on shorter time scales.)

The quantities reported here are subject to uncertainties in the analyzed observations. Errors in the gridded data result from both sampling and analysis techniques. First, the wind and humidity measurements are taken at most twice daily, which is probably insufficient to capture the diurnal cycle in water vapor flux (Rasmussen 1967). Second, routine radiosonde measurements are not well resolved in the planetary boundary layer (surface to 850 mb), where a significant part of vapor transport occurs. Standard reporting levels covering this layer are 1000 and 850 mb. (The GFDL analysis includes values of the variables at 950 and 900 mb, but these are based on fewer observations than the analyses at the standard levels.) Third, the use of different sensors and instrument packages introduces time- and space-varying biases in the humidity measurements (Elliott and Gaften 1991). Finally, separate analysis of the velocity components, with heavy smoothing in some locations, has the result that certain dynamical constraints are not satisfied. For example, integration of total airmass flux across the boundaries of the North American study region (where radiosonde data available for analysis are rather densely distributed, compared to the rest of the globe) indicated a net airmass divergence over the year, in violation of the principle of mass conservation. Given the existence of such errors over the most data-rich region of the study, greater errors may be expected for the other study regions, particularly West Africa. Various correction techniques have been applied to analyzed datasets (Aalestalo 1983; Savijärvi 1988); however, for this study, no corrections were applied based on the judgment that all techniques would be equally arbitrary given the other uncertainties in the data.

The local precipitation-recycling ratio, as estimated herein, is a diagnostic measure of the predominant climatic regime. The values are only general indicators of the importance of land–atmosphere interaction to regional climate. As such, the estimates are not predictive; for example, it does not follow that reducing evapotranspiration to zero in the Eurasian region during June (even if that were possible) would reduce June precipitation by 35%. Many complex, interrelated factors, both internal and external to the region, control the region's precipitation.

It is clear that a great deal of variety exists in the importance of the precipitation recycling process to different regions and at different seasons. Therefore, Budyko's (1974) conclusion that the relative role of local recycling is limited, and Lettau et al.'s (1979) conclusion that it is quite important are not contradictory. The reason is that the process depends not only on the region's length scale (or tortuosity of the airmass streamtube over the territory) and evapotranspiration rate, but also on the presence of the physical mechanisms that cause precipitation to occur at all.

If a sufficiently high orographic barrier is present at the downwind boundary of a region, most of the moisture present in the air, regardless of its evaporative source, precipitates on the windward side. By the definition of recycling used in this study, the moisture content due to evapotranspiration within the region—which lies simultaneously upwind and downwind—falls back into the region as recycled precipitation. This mechanism illustrates the sensitivity of the recycling notion to the choice of control volume and sheds some light on why increased irrigation could increase rainfall in the Columbia River basin (Stidd 1967), whereas building a large lake on the southern border of Arizona would not increase rainfall in that state (McDonald 1962). Adding moisture to the air at low levels tends to reduce the stability of the atmosphere by building up a reservoir of latent heat in the lower layers (Stidd 1967). By this contribution, increased evapotranspiration could enhance convective precipitation. If so, advective precipitation ($P_a$) would likely be increased as well as $P_m$, whereas with the orographic mechanism, increased evapotranspiration would only increase $P_m$.

In this study, the partition of precipitation between local and advective sources has been emphasized rather than the oceanic–terrestrial distinction. For studies of the interactions between the land surface and regional climate, the internal versus external supply of water vapor is of interest, whether that external supply comes from an ocean or a land surface. As noted before, the definitions merge if the control volume is taken as an entire continent. Rosen and Omolayo (1981) present such an analysis wherein they compute the boundary flux of atmospheric water vapor across Northern Hemisphere coastlines. When the ocean–land distinction is of the primary interest, analysis methods based on stable and radioactive isotopes have been demon-
strated to be useful techniques in distinguishing the evaporation sources of precipitating moisture (Baudet and Lesne 1975; Baudet and Abi 1979).

7. Summary

Analyzed fields of observed humidity and wind profiles provide quantitative estimates of water vapor transport in the atmospheric branch of the hydrologic cycle. Surface precipitation is water derived from the water vapor reservoir in the overlying atmospheric air column in the same proportion as the abundance of vapor of different origins. The fraction of the total precipitation derived from within a region, as opposed to precipitation of water due to water vapor advected into a region, is considered recycled water.

For four large innercontinental regions, covering parts of Eurasia, North America, South America, and Africa, the fraction of precipitation derived from evapotranspiration from within the region is determined for each month of the year based on ten years of observed data. There is generally a strong annual cycle in the estimated ratio; the mean (range) of the monthly values for the Eurasian region is about 0.11 (0.31), 0.24 (0.19) for the North American region, 0.24 (0.18) for the South American region, and 0.31 (0.38) for the African region. These climatic diagnostics are approximate measures of the importance of land-surface hydrology in maintaining the climate of the regions. The supply and partial control of regional precipitation by local hydrologic and soil-moisture processes are potentially significant mechanisms of land-surface–atmosphere interaction, which may contribute to some characteristics of climatic variability.

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