Investigation of the Large-Scale Atmospheric Moisture Field over the Midwestern United States in Relation to Summer Precipitation. Part II: Recycling of Local Evapotranspiration and Association with Soil Moisture and Crop Yields

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(Manuscript received 17 April 2003, in final form 16 January 2004)

ABSTRACT

The relative contributions of locally evapotranspired (i.e., recycled) moisture versus externally advected water vapor for the growing-season precipitation of the U.S. Corn Belt and surrounding areas ($1.23 \times 10^6$ km$^2$) are estimated in this paper. Four May–August seasons with highly contrasting precipitation and crop yields (1975, 1976, 1979, and 1988) are investigated. A simple recycling equation—developed from the traditional atmospheric moisture budget and involving regional evapotranspiration and atmospheric water vapor inflow—is applied on daily, monthly, and seasonal time scales. Several atmospheric moisture budget components {moisture flux divergence [MFD], storage change [or change in precipitable water (dPW)], and inflow [IF]} are evaluated for 24-h periods using standard finite difference and line integral methods applied to objectively analyzed U.S. and Canadian rawinsonde data (50-hPa vertical resolution, surface to 300 hPa) for 0000 and 1200 UTC. Daily area-averaged precipitation ($P$) totals are derived from approximately 600 evenly distributed (but ungridded) recording rain gauges. Evapotranspiration ($E$) is estimated as the residual of the moisture budget equation for 24-h periods; values compare favorably with the few existing observations.

Traditional budget results show the following: $E$ is weakly related to $P$ on monthly and seasonal time scales; there is surprising interannual constancy of seasonal $E$ cycles and averages given the large variation in resulting crop yields; and monthly and seasonal variability of the export of the $E - P$ surplus is determined largely by the horizontal velocity divergence component of MFD. New recycling analyses suggest that the contribution of local $E$ to $P$ (i.e., $P_e / P$) is relatively small and remarkably consistent (largely 0.19–0.24) for monthly and seasonal periods, despite large $P$ and crop yield variations. However, the monthly/seasonal averaging process is found to completely mask a striking decrease of daily $P_e / P$ (from approximately 0.30 to 0.15) with increasing $P$ from 0 to 8 mm day$^{-1}$. Unique and detailed analyses of $P$-stratified daily moisture budget results provide key insights into apparent contradictions between daily and monthly/seasonal recycling and related results and concomitant interannual variability, especially for the very dry 1988 season. Interpretation is facilitated by the use of modeled daily global radiation values, measured (instantaneous) and modeled (monthly) soil moisture, United States Department of Agriculture (USDA) crop yield estimates, and satellite normalized difference vegetation index (NDVI) imagery. This paper shows that land–atmosphere interactions are intimately involved in pronounced seasonal climate anomalies for the world’s richest agricultural region, but apparently with considerable complexity that includes plant behavior, solar radiation forcing, and challenging time-scale interrelations.

1. Introduction

Land–atmosphere interactions are central to the natural environment, involving and affecting individual weather systems, regional climate, the hydrological cycle, soil and vegetation status, and agricultural production (e.g., Shukla and Mintz 1982; Rind et al. 1992; Chahine 1992a; Beljaars et al. 1996; Hahmann and Dickinson 1997, 2001; Zeng and Neelin 2000). This importance is manifest in several long-term, international biological and geophysical programs, including the Global Energy and Water Cycle Experiment (GEWEX) of the World Climate Research Programme (WCRP 1990; Chahine 1992b). The first major GEWEX activity, the GEWEX Continental International Project (GCIP), focused on the land–atmosphere interactions of the greater Mississippi River basin during the second half of the 1990s (WCRP 1990, 1992; Chahine 1992b; In-
ternational GEWEX Project Office 1993, 1994a,b). Consistent with broad GEWEX/GCIP goals, Zangvil et al. (2001, hereafter Part I) investigated the time-scale relationships among the large-scale atmospheric moisture budget components over the midwestern United States (35% of the GCIP domain) in relation to summer precipitation. In this companion study, we more specifically focus on the role of land–atmosphere interactions in the same large-scale atmospheric moisture budget context.

A land–atmosphere interaction problem of utmost environmental importance involves the classical issue of the relative contributions to regional precipitation of locally evaporated and transpired (i.e., recycled) moisture versus externally advected atmospheric water vapor (e.g., Benton et al. 1950; Budyko 1974, 239–243; Bru-baker et al. 1993; Burde et al. 1996; Bosilovich and Schubert 2001; Burde and Zangvil 2001a,b). The locally supplied moisture or upward flux of water vapor can be from evaporation of in situ open water or soil moisture, or from plant transpiration. Henceforth, we employ the collective term “evapotranspiration” for these processes. Using a simple formulation, involving an extension of the conventional moisture budget treatment in Part I, we present the first comprehensive estimation of the intraseasonal and interannual variability of locally recycled versus externally advected moisture sources for the growing season precipitation of the world’s most productive, largely unirrigated, agricultural region: the Corn Belt and surrounding areas of the midwestern United States (Fig. 1; cf. Fig. 6a) which, as noted, occupy 35% of the GCIP domain. Consistent with its GCIP collocation, this region is considered representative of the midlatitude, midcontinent land–atmosphere interactions that are vital for global water resources and food production. Here, we extend our investigation of the time-scale relations among the moisture budget terms initiated in Part I to include recycling ratios, the precipitation amounts supplied by the local moisture sources, and variables having a direct connection to that local moisture—solar radiation forcing and soil moisture support of evapotranspiration, a satellite-based vegetation index, and crop yields.

Since Part I contains a comprehensive review of previous global and regional atmospheric water vapor budget studies, along with a full treatment of the theory and computational procedures involved, most of that material is not repeated here. However, some repetition is needed to provide the context of the present inquiry. The current section 2 extends and tailors moisture budget theory and methodology to our recycling focus and also explains the development of three additional sets of environmental data (solar radiation, soil moisture, and crop yields) used to interpret the moisture budget results. Section 3 employs a new treatment of the traditional moisture budget analysis as a lead-in to a detailed examination of the relations between moisture budget terms and moisture recycling. In section 4, we attempt to interpret the complex land–atmosphere interactions involved in terms of plant behavior, solar radiation forcing, and challenging time-scale interrelations.

2. Data and methods

a. Moisture budget equations

As noted in Part I, the size of the present study region (1.23 × 10⁶ km²) approximates or exceeds the typical scales of most individual weather systems that affect the region during May–August and the resulting mid-latitude, midcontinent growing week-to-month-to-season precipitation anomalies (Richman and Lamb 1985, 1987; Kunkel and Angel 1989; Lamb and Richman 1990; Bell and Janowiak 1995). Also, this size exceeds the 0.6–1.0 × 10⁶ km² minima recommended/used for atmospheric moisture budget studies with monthly and seasonal time scales (Rasmussen 1968, 1971; Yanai et al. 1973; WCRP 1992). Following Rasmussen (1968, 1971), Yanai et al. (1973), and Part I, the traditional atmospheric moisture budget equation for such a region for time period Δt can be expanded to take the following form:

\[
\frac{1}{g} \frac{\partial}{\partial t} \int_s \sigma \, dp + \frac{1}{g} \int_s \nabla \cdot q \sigma \, dp + \frac{1}{g} \int_s q \nabla \cdot V \, dp = E - P,
\]

\[
dPW \quad HA \quad HD
\]

where \( g \) is the acceleration due to gravity, \( q \) is specific humidity, \( p \) is atmospheric pressure (\( S \) and \( T \) indicate the earth’s surface and an appropriate upper integration limit, respectively), \( V \) is the horizontal wind vector, and \( E \) and \( P \) are the surface evapotranspiration and precipitation rates, respectively. All terms are time and space averaged (see below) and expressed as water depth equivalents (mm day⁻¹) across the study region. Physically, in Eq. (1), \( dPW \) is the time change of atmospheric water vapor [precipitable water (PW)], HA gives the horizontal water vapor advection, HD includes the horizontal velocity divergence and is equal to the vertical water vapor advection (VA), and HA + HD is the moisture flux divergence (MFD).

Atmospheric moisture budget studies have traditionally employed Eq. (1) directly in an attempt to quantify the moisture sources for \( P \) (e.g., Benton et al. 1950; Thompson et al. 1979; Bosart and Sanders 1981; Kunkel 1989). This approach involved neglecting the possible contribution from atmospheric storage depletion (−\( dPW \), justifiable; see below) and assuming that \( E \) and negative MFD (i.e., water vapor convergence) proportionately contribute to \( P \). However, this equating of the external moisture source for \( P \) with negative MFD ignores the fact that MFD is the difference between atmospheric inflow (source) and outflow (sink) for a region. As documented in section 3b, this traditional ap-
proach underestimates the contribution of external moisture to \( P \), which causes overestimation of the role of \( E \).

As just noted, standard evaluations of the MFD terms (individually or collectively) in Eq. (1) cannot distinguish the relative contributions of \( E \) and imported water vapor to \( P \). Instead, they only can yield information on the relationship between the difference \((E - P)\) and the total MFD plus the atmospheric storage change (dPW). The first step in surmounting this problem is to express the MFD terms in Eq. (1) as

\[
\text{HA + HD} = \frac{1}{g} \int_{s} \nabla \cdot qV \, dp = \frac{1}{Ag} \int_{s} \bigotimes qV_n \, dl \, dp
\]

\[
= \frac{\text{OF}}{A} - \frac{\text{IF}}{A} + \text{dPW},
\]

where \( A \) is the area of the region, \( V_n \) is the component of the wind normal to the region's boundary, \( dl \) is a unit length of that boundary, and \( \text{OF} \) and \( \text{IF} \) are the total water vapor mass outflow from and inflow to the region, respectively. Substitution of Eq. (2) into Eq. (1) yields

\[
E - P = \frac{\text{OF}}{A} - \frac{\text{IF}}{A} + \text{dPW}.
\]

The advantage of Eq. (3) over Eq. (1) is that it identifies \( \text{IF}/A \) as the source of externally advected water vapor that supplies the part of the precipitation \( (P_{\lambda}) \) not derived from local evapotranspiration \( (P_{E}) \), that is, \( P = P_{\lambda} + P_{E} \).

However, like Eq. (1), Eq. (3) cannot be used to quantify the relative contributions of \( E \) and \( \text{IF}/A \) to \( P \). The next step in addressing this problem is to depart from Eq. (3) and separately develop a conceptual model to treat moisture recycling. The model's fundamental assumption is that water vapor of advected (subscripted \( A \)) and evapotranspired (subscripted \( E \)) origins is fully mixed in an atmospheric “tank” over the study domain (within which all moisture fields slowly vary) and thus makes proportionate contributions to \( P \). Using the above notation and a subscripted “0” to denote the start of time period \( \Delta t \), this model requires that

\[
\frac{P}{P_{E}} = \frac{PW_{E0} + PW_{A0} + E(\Delta t) + (\text{IF}/A)\Delta t}{PW_{E0} + E(\Delta t)}
\]

\[
= 1 + \frac{(\text{IF}/A)\Delta t + PW_{A0}}{E(\Delta t) + PW_{E0}}
\]

and

\[
\frac{PW_{A0}}{PW_{E0}} = \frac{(\text{IF}/A)\Delta t}{E(\Delta t)}.
\]

Equations (4) and (5) can be combined and rearranged to yield the following relation \(^1\) for the precipitation fraction supplied by evapotranspired moisture:

\[
\frac{P_E}{P} = \frac{E}{E + \frac{\text{IF}}{A}}.
\]

Note that Eq. (6) was not derived formally from Eq. (3) but involves processes defined and treated in the more traditional equations (1)–(3).

Because of its above assumptions, this tank formulation is applicable not only on monthly and seasonal time scales, but also to sets of (not necessarily contiguous) days with similar large-scale moisture characteristics (atmospheric inflow, evapotranspiration, precipitation, etc.) Burde et al. (1996) and Burde and Zangvil (2001b) showed that, under highly perturbed conditions (nonparallel flow fields), Eq. (6) gives somewhat similar values as a two-dimensional generalization of Budyko's (1974, 240–241) classical, more restrictive, one-dimensional streamline approach. Otherwise, Eq. (6) inherently yields higher \( P_{E}/P \) ratios than the original Budyko-type treatments (Budyko 1974, 241–243; Brubaker et al. 1993) by a factor ranging from 1 to 2 depending on the relative values of \( E, \text{IF}, \) and \( A \). We have estimated this factor to be 1.5–1.7 for the following results—those we present below; Budyko's (1974, 241–243) estimates for the European part of the former Union of Soviet Socialist Republics (USSR); and Brubaker et al.'s (1993) linear generalization of Budyko's original one-dimensional formulation to two dimensions, which is different from the aforementioned generalization discussed in Burde et al. (1996) and Burde and Zangvil (2001b). The Budyko (1974) and Brubaker et al. (1993) equations are given in the appendix.

Equation (6) is similar to other recycling formulas developed more recently for different regional applications (Brubaker et al. 1993; Eltahir and Bras 1994; Schär et al. 1999). See the appendix for formulas used in other studies for the central United States. However, our assumptions, derivation, and application are more straightforward and versatile. Specifically, our approach has no parallel or uniform flow requirements (cf. Brubaker et al. 1993), involves no averaging of moisture inflow and outflow (cf. Brubaker et al. 1993), requires no assumptions about neglecting atmospheric moisture storage or the source partitioning of atmospheric moisture outflow (cf. Eltahir and Bras 1994; Schär et al. 1999; see also Trenberth 1999), involves the application of a “bulk” formula to an appropriately sized region rather than to small grid cells within such a region (cf. Eltahir and Bras 1994; see also Bosilovich and Schubert 2001), does not include moisture inflow/outflow corrections for eddy contributions in the model output used (cf. Schär et al. 1999), and is employed to treat the daily time scale as well as traditional monthly and seasonal periods (cf. Brubaker et al. 1993; Eltahir and Bras 1994; Schär et al. 1999; see also Bosilovich and Schubert 2001). Note in particular that our unique treatment of the daily time scale, involving sets of days with similar

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\(^1\) As noted in Zangvil et al. (1992) and Burde and Zangvil (2001a), we first presented the above derivation of Eq. (6) in 1992 at the Seventeenth Annual Climate Diagnostics Workshop (University of Oklahoma, Norman, Oklahoma) and the Yale Mintz Memorial Symposium on Climate and Climate Change (Jerusalem, Israel).
large-scale moisture characteristics, is permitted by the key assumption embodied in Eq. (5) above.

b. Moisture budget computations

The budget terms and associated parameters in Eqs. (1)–(6) were evaluated for the region delineated in Fig. 1 for all 24-h (1200–1200 UTC) periods during the highly contrasting (Table 1) May–August growing seasons of 1975, 1976, 1979, and 1988. Area averages were computed from gridpoint or point values within the region, and 24-h time averages were obtained by applying a 0.25–0.50–0.25 filter to sequential 1200–0000–1200 UTC values. These procedures were applied to estimates of (i) HA and HD (surface to 300 hPa) for contiguous 1200–0000–1200 UTC sequences given by finite difference evaluations of rawinsonde data (50-hPa resolution) from all stations in Fig. 1 that were interpolated to a 190-km grid via a Barnes-type objective analysis scheme (Achtemeier 1986), and adjusted using the O’Brien (1970) method; (ii) dPW (surface to 300 hPa) from successive 1200 UTC values of similarly gridded rawinsonde data (50 hPa); (iii) \( P \) from 24-h totals for approximately 600 relatively evenly distributed (but un-gridded) hourly recording rain gauges; and (iv) IF/A (surface to 300 hPa) for contiguous 1200–0000–1200 UTC sequences given by line integration using the gridded rawinsonde data. In addition, 24-h area-averaged estimates of \( E \) were obtained as residuals of Eq. (1); negative \( E \) values (5.3%) were small and set to 0, with their associated budgets being rebalanced by equally compensating for the water added to \( E \) among the other
terms. Finally, 24-h values of \( P_E/P \) and \( P_E \) were calculated using Eq. (6) and the above \( E \), IF/A, and \( P \) estimates.

All above 24-h budget term/associated parameter values were then averaged over individual calendar months, May–August periods, and sets of days for five \( P \) categories (\( P \leq 0.6 \), \( 0.6 < P \leq 2 \), \( 2 < P \leq 4 \), \( 4 < P \leq 8 \), \( P > 8 \) mm day\(^{-1} \)). Results appear in Tables 2 and 3 (and Figs. 2, 4, and 5 later). In addition to the \( P_E/P \) averages obtained in this way, Table 3 gives \( P_E/P \) ratios in parentheses that were calculated from their individual \( P_E \) and \( P \) totals accumulated from 24-h values. The generally close agreement between these two sets of \( P_E/P \) results strongly supports the assumptions used to derive Eq. (6) and its application to sets of (not necessarily contiguous) days with similar large-scale moisture characteristics.

See Part I for details not given above, including data sources, uncertainties, quality control procedures, and additional climate characteristics of the study seasons. As in Part I, our atmospheric moisture budget results are considered not to be significantly affected by recently emphasized rawinsonde humidity uncertainties (stemming from varying measurement, conversion, reporting, automation, and archiving practices) that are more likely to bias longer-term climate studies (e.g., Schwartz and Doswell 1991; Elliott and Gaffen 1991, 1993). The atmospheric residual estimates of \( E \) obtained (Table 3: Figs. 2 and 5 later) are in good general agreement with (Crop Estimation through Resource and Environmental Synthesis) CERES-Maize model-based estimates for July for a similar region (Kunkel 1992). Our nonuse of recently available reanalysis data in the above computations stems from the prior initiation of the project and relevant reanalysis fields being relatively unreliable (Kalnay et al. 1996; Mo and Higgins 1996; Trenberth and Guillemot 1998) and follows Part I.

c. Related environmental data

Interpretation of the above moisture budget estimates required the development of four additional sets of related environmental data for the study region and seasons: daily area averages of modeled global solar radiation (SR; Tables 2 and 3; Figs. 3 and 5 later), instantaneous point measurements and monthly area-averaged model estimates of soil moisture (Figs. 6 and 7 later), seasonal area-averaged yields for the region’s principal row crops (corn, soybeans) in the context of yield and seasonal precipitation variability for 1972–90 (Table 1; Figs. 8c,d,f later), and seasonal area-averaged precipitation anomalies relative to 1972–90 variability (Table 1).

Daily SR area averages for all study seasons were

<table>
<thead>
<tr>
<th>( P )</th>
<th>( E )</th>
<th>( E-P )</th>
<th>HA</th>
<th>HD</th>
<th>MFD</th>
<th>dPW</th>
<th>IF/A</th>
<th>( P_E/P )</th>
<th>( P_E )</th>
<th>SR</th>
</tr>
</thead>
<tbody>
<tr>
<td>( P )</td>
<td>+1.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>( E-P )</td>
<td>-0.72</td>
<td>+0.32</td>
<td>+1.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>HA</td>
<td>-0.40</td>
<td>-0.14</td>
<td>+0.31</td>
<td>+1.00</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>HD</td>
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<td>+0.92</td>
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<tr>
<td>MFD</td>
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<td>+0.95</td>
<td>+0.42</td>
<td>+0.93</td>
<td>+1.00</td>
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<td></td>
</tr>
<tr>
<td>dPW</td>
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<td>-0.30</td>
<td>-0.19</td>
<td>-0.45</td>
<td>-0.37</td>
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</tr>
<tr>
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<td>-0.39</td>
<td>-0.09</td>
<td>-0.22</td>
<td>-0.23</td>
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</tr>
<tr>
<td>( P_E/P )</td>
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<td>+0.54</td>
<td>+0.49</td>
<td>+0.01</td>
<td>+0.46</td>
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<tr>
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<td>-0.56</td>
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<td>-0.37</td>
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<td>+0.53</td>
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<td>-0.30</td>
</tr>
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</table>

Table 2. Atmospheric moisture budget relationships for study seasons. Values are linear correlation coefficients between individual monthly mean values of atmospheric moisture budget components and related parameters in Fig. 2 for all four years combined (\( N = 16 \)). Moisture budget component abbreviations and sign conventions are as defined in section 2a. Consistent with Eqs. (3) and (6), IF/A is considered positive. See section 2c for an explanation of SR. According to the \( t \) test described in section 2c, the 0.1%, 1%, and 5% significance levels for the correlation magnitudes are 0.74 (italic) 0.62 (bold), and 0.50 (italic–bold), respectively.
computed from objective analysis (same 190-km grid and method as for rawinsonde data) of SR estimates given by a semophysical model for 53 locations in and adjacent to the present study region (Petersen et al. 1995). See Petersen et al. (1995) for details of model structure and required meteorological input data. Use of this model was necessitated by the lack of SR observations.

Two sources of soil moisture information were used. Neutron-probe measurements made during the very dry year of 1988 at three representative Illinois Climate Network stations (Hollinger and Isard 1994) were obtained from the Global Soil Moisture Data Bank (Robock et al. 2000) Web site (http://climate.envsci.rutgers.edu/soil.moisture/). These measurements were made at intervals of 1 or 2 weeks and yielded the total soil moisture in the top 2 m. Such measurements were not available for the other study seasons for Illinois or for the rest of the domain for all study seasons. This situation necessitated the use of modeled soil moisture estimates (Huang et al. 1996) for the region as a whole for 1988 and the much wetter 1979, which were also obtained from the above Web site. These estimates were monthly averages for the National Oceanic and Atmospheric Administration (NOAA) climate divisions (75 in the study region) given by a one-layer hydrological model that treats soil moisture through the depth that “it participates in land-surface processes, that is usually in the upper 1–2 m of soil” (Huang et al. 1996, p. 1352). See Hollinger and Isard (1994) and Huang et al. (1996) for further details concerning the basis of these soil moisture data.

County-scale crop production and hectare estimates for 1972–90 were obtained from data bases maintained by the U.S. National Agriculture Statistics Service and are available on their Web site (http://www.nass.usda.gov). Normalized corn and soybean yield anomalies (ν) were computed for our study seasons from detrended
Fig. 2. Mean values of atmospheric moisture budget components and related parameters for 24-h periods during individual months (thin bars, letter is first letter of month) and entire seasons (thick bars, number is last number of year) for May–Aug 1975, 1976, 1979, and 1988 (mm day$^{-1}$, except for dimensionless $P/E$). Budget component/parameter abbreviations and sign conventions are as defined in section 2a and Table 2.

(via linear regression) time series for 1972–90 ($N = 19$). Anomalies were obtained separately for the entire study region (Table 1) and its 81 United States Department of Agriculture (USDA) Crop Reporting Districts (Figs. 8c,d,f later). This detrending effected a conventional removal of technology-induced (through pesticide and plant hybrid development) yield increases. The corn (soybean) regression line for the entire study region had a slope of $-0.09$ Mg ha$^{-1}$ yr$^{-1}$ ($-0.02$ Mg ha$^{-1}$ yr$^{-1}$) with an explained variance fraction of 22.5% (23.3%) and a significance level of 4.0% (3.6%) according to an $f$ test (equivalent of a two-tailed $t$ test) with $N - 2$ degrees of freedom. Normalized “combined crop” yield anomalies ($\sigma$) were computed from hectare-weighted averages of normalized corn and soybean yield anomalies. The above crop analyses did not include data for the extreme southern tip of Ontario, which occupies 1.76% of the study region (Fig. 1).

Normalized seasonal (May–August) precipitation anomalies ($\sigma$) for the entire region were used to rank the 19 yr during 1972–90 and indicate the relative moistness of our study seasons (Table 1). These anomalies were computed from actual precipitation totals (not detrended) for 145 evenly spaced stations in the Richman–Lamb daily precipitation dataset (e.g., Richman and Lamb 1985, 1987; Lamb and Richman 1990). The absence of a trend in this precipitation anomaly time series is evidenced by its linear regression fit having a slope of only $-0.0087$ mm yr$^{-1}$ with an explained variance fraction of only 0.0005% and a significance level of 92.9%, according to the above $f$ test.

3. Results and discussion

a. New treatment of the traditional moisture budget analysis

Results from the traditional moisture budget analysis [Eq. (1)] for the study months and seasons are included in Table 2 and Fig. 2 (and Fig. 4). This section focuses on various interrelations between $P$, $E$, and the bulk $E - P$, the total MFD and its HA and HD components, and $\delta P$—but not the moisture sources for $P$. Quantification of the moisture sources for $P$ appears in the
next section, where the treatment expands to include the new parameters $\frac{\text{IF}}{\text{A}}, P_e/P$, and $P_e$. The present section begins with a conventional presentation and brief discussion of traditional results and includes repetition of a small amount of material in Part I. This provides a necessary lead-in to a new treatment of traditional moisture budget results with which the section concludes. See Part I for a more complete presentation of our traditional budget results.

Figure 2 shows that $P$ varies substantially on both monthly (intraseasonal and interannual) and seasonal time scales but is weakly related to $E$. Their monthly correlation is small (+0.43) and not statistically significant (Table 2). This correlation is little changed when the seasonal cycles are removed from each parameter (+0.46; see Table 7 in Part I). Except in May 1979, $E$ is larger and more consistent than $P$ on monthly and seasonal time scales; the interannual constancy of the seasonal $E$ cycles and averages is striking (Fig. 2), particularly given the large variation in the resulting crop yields (Table 1). As a result, $E - P$ is positive for all cases except May 1979 and varies principally (and inversely) with $P$ rather than $E$. The monthly correlation between $E - P$ and $P$ is $-0.72$ (significant at a 1% level), whereas it is only $+0.32$ and statistically insignificant for $E - P$ versus $E$ (Table 2). This correlation difference becomes even larger when the seasonal cycle is removed from each parameter ($-0.90$ versus $-0.03$; Table 7 in Part I).

According to Fig. 2, most of this $E - P$ atmospheric moisture surplus is exported from the region (positive MFD), with little being stored locally (positive dPW). In 1--2 months of every season, the export even slightly depletes PW. Although this export consistently involves at least moderate (dry) horizontal water vapor advection (positive HA; Fig. 2), its substantial monthly and seasonal fluctuations are strongly determined by the horizontal velocity divergence (HD; Fig. 2). This situation is documented statistically by the large difference between Table 2’s monthly correlations of MFD with HD (+0.93, 0.1%) and HA (+0.42, not significant), which is reduced only slightly by seasonal cycle removal (+0.92 versus +0.51; Table 7 in Part I). Figure 2 contains revealing examples of this atmospheric behavior—the export of the large $E - P$ moisture surpluses associated with the low $P$ of July 1975, August 1976, and June 1988 involved substantial HD supplements to strong HA, whereas the reasonably similar HA contributions to the MFD for the much wetter May--July 1979 were opposed by the effect of convergent horizontal velocity fields (negative HD). In the extreme case of May 1979, this rendered the total MFD convergent, which, in turn, supported the largest monthly PW increase and the only monthly $E - P$ deficit of the study periods.

The above traditional budget results confirm and extend, on an individual monthly/seasonal basis, the key findings of earlier multiyear average investigations (e.g., Benton et al. 1950; Rasmusson 1968, 1971; Brubaker et al. 1993). Our results are consistent with the overwhelmingly convective origin of the precipitation, the development of which is favored by large-scale horizontal velocity convergence and ascent (hence above HD importance) and produces relatively small/short-lived cloud elements that only weakly modulate SR (hence above $E$ constancy). Concerning the latter, the interannual variability of seasonal mean $SR$ ($21.00$--$23.88$ MJ m$^{-2}$ day$^{-1}$) and $E$ ($3.99$--$4.50$ mm day$^{-1}$) both span only 13% and surprisingly are negatively related (Figs. 2 and 3).

Table 2 shows that this negative correlation between SR and $E$ also prevails on the monthly time scale ($-0.55, 5\%$), which is surprising given the relatively strong positive correlation between monthly SR and dPW ($+0.68, 1\%$). However, these SR correlations with $E$ and dPW principally reflect seasonal cycle behavior, as they decline to $+0.16$ and $-0.03$, respectively, when that influence is removed. Some understanding of these weak seasonal and monthly relations among SR (and hence cloudiness), $E$, and dPW is offered in subsequent sections.

Greater insight is obtained into some of the mechanisms of the atmospheric moisture budget—but still not the moisture sources for $P$—when results from the traditional approach [Eq. (1)] are stratified by the daily $P$ amount in an analysis not performed before this work. Full documentation of the analysis is presented in Table 3 (except for SR and the last three columns on the right-hand side), and selected results appear in Fig. 4. Note that while Part I (Table 5) included all-year composite results drawn from this work, it did not investigate the underlying interannual variability, which we undertake here. While a distinctive feature of Table 3 and Fig. 4 is the generally strong within-$P$-category (i.e., interannual) uniformity and between-$P$-category separation of the $E - P$, HA, HD, and MFD results, several interannual differences that involve particularly the two overall driest study seasons (1988 and 1976) are equally distinctive.

In Table 3 and Fig. 4, days with $P \leq 0.6$ mm are of
course characterized by large $E - P$ atmospheric moisture surpluses, most of which are exported (primarily through HD) from the region with the remainder being stored locally. When daily $P$ increases to $0.6 < P \leq 2$ mm, $E - P$ is still positive but substantially reduced, the atmosphere still exports most of this (diminished) surplus (but now through HA as well as HD), and local storage is reduced, even to the point of becoming slightly negative in the overall driest season (1988; Table 1; Fig. 2). These changes become more pronounced as daily $P$ increases further to $2 < P \leq 4$ mm (Table 3; Fig. 4). For this $P$ category, the $E - P$ atmospheric moisture surpluses are small (except for the driest year, 1988, when there was a small deficit) and, along with generally modest local storage depletions, support weakly divergent MFDs that result from divergent HA slightly exceeding small moist HD. These changes are accentuated considerably when daily $P$ increases further to $4 < P \leq 8$ mm (Table 3; Fig. 4). Here, moderate ($E - P$) deficits are largely balanced by local storage depletions and, to a lesser extent, by weakly convergent MFDs that result from more strongly convergent HD slightly exceeding strong dry HA. These results are especially characteristic of the two overall driest seasons (1988 and 1976; Fig. 2), when both $E$ (Table 3) and crop yields (Table 1) were very low.

Some of these results for $4 < P \leq 8$ mm day$^{-1}$ are further accentuated when the daily $P$ exceeds 8 mm (Table 3; Fig. 4), but interesting differences also emerge for the highest $P$ category that is represented by a much smaller and presumably less reliable sample. Here, the $E - P$ deficits are larger despite $E$ being higher (including the two overall driest seasons and presumably due to surface moisture abundance) than when $4 < P \leq 8$ mm day$^{-1}$ (Table 3), but they are now primarily balanced by strongly convergent MFDs that result from
strenthened HD convergence substantially exceeding weakened dry HA. Some of this convergent MFD is also stored locally in most seasons.

b. Relationships between moisture budget terms, moisture recycling, soil moisture, and crop yields

Important new insight into midlatitude, midcontinent, growing season land–atmosphere interaction is added to the above results of traditional analyses through stratification of the new atmospheric moisture budget recycling parameters [IF/A, \( P_e/P, P_e \); Eqs. (4)–(6)] by the daily P amount (Table 3; Fig. 5). In particular, the recycling ratio (\( P_e/P \)) quantifies the contribution of local \( E \) (versus imported atmospheric moisture) to \( P \). As for the parameters treated in the preceding section, Table 3 and Fig. 5 contain both generally strong within-P-category uniformity and between-P-category separation of the IF/A, \( P_e/P \), and \( P \) results, and distinctive interannual differences for the driest seasons (1988 and 1976). To facilitate interpretation of these recycling results, Fig. 5 also includes the related parameters \( E \) and SR.

Here, \( P_e/P \) shows a striking decrease (by an average of 50%) as \( P \) increases from 0–8 mm day\(^{-1} \) (Table 3, categories A–D; Fig. 5), and then levels off or possibly slightly increases as \( P \) exceeds that threshold (Table 3, category E; Fig. 5). Specifically, the mean \( P_e/P \) progressively decreases from \( P \leq 0.6 \) (0.27–0.34 range for individual seasons and 0.30 for all seasons combined), to 0.6 < \( P \) ≤ 2 (0.18–0.22 and 0.21), to 2 < \( P \) ≤ 4 (0.13–0.22 and 0.19), and to 4 < \( P \) ≤ 8 mm day\(^{-1} \) (0.11–0.17 and 0.15). Once the daily \( P \) exceeds 8 mm, this \( P_e/P \) decrease is suggested to level off or even reverse slightly (0.16–0.19 and 0.17). However, the number of days when \( P > 8 \) mm (26) was much smaller, and hence presumably less reliable, than for the four lower \( P \) categories (99–127). Table 3 and Fig. 5 also show that this \( P_e/P \) decrease as \( P \) increases from 0–8 mm day\(^{-1} \) is associated with a pronounced elevation in the magnitude of the region’s water vapor inflow (IF/A), which is further enhanced as \( P \) exceeds 8 mm day\(^{-1} \) (when \( E \) is also higher; see below). These results strongly imply that IF/A thresholds of 14, 16, 20, and 25 mm day\(^{-1} \) must be surpassed for \( P \) to exceed 0.6, 2, 4, and 8 mm day\(^{-1} \), respectively.

Note that while \( P_e/P \) decreases by an average of 50% as \( P \) increases from 0–8 mm day\(^{-1} \), the \( P_e \) magnitude inflates approximately tenfold over the same range (Ta-
ble 3; Fig. 5). Thus, when $P_{sv}/P$ is largest, only miniscule amounts of water vapor are recycled (0.06–0.08 mm day$^{-1}$). In contrast, this absolute recycling averages almost 1 mm day$^{-1}$ when the relative recycling is lowest (for $4 < P \leq 8$ mm day$^{-1}$) and apparently increases further to around 2 mm day$^{-1}$ when daily $P$ exceeds 8 mm. This predominately negative daily relation between $P_{sv}/P$ and $P$ is not captured on the monthly time scale, for which the correlation coefficient is $+0.18$ (Table 2; $+0.15$ after seasonal cycle removal).

The inverse relationship between daily $P_{sv}/P$ and $P$ strongly characterizes individual seasons, despite the wide variations in their monthly and seasonal $P$ totals (Fig. 2; Table 1) and crop yields (Table 1). However, several distinctive interannual variations are apparent within this overall relationship (Table 3; Fig. 5). Most prominent are the especially low $P_{sv}/P$ values for the days when $0.6 < P \leq 8$ mm during the very dry and unproductive 1988 growing season. For these days, $E$ was strikingly low and offset by compensating and consistently high IF/A, especially when $2 < P \leq 4$ mm day$^{-1}$. This suppressed $E$ was not due to anomalously low SR (Table 3; Fig. 5): instead, it likely reflects a poor crop status and precursor soil moisture consistent with the overall extremely low $P$ and crop yields in that year (Table 1). Low atmospheric humidity levels can inhibit plant transpiration (Cowan 1984) and, in this case, would also have stemmed from low soil moisture limiting the evaporranspirative water vapor source from both the plants and the soil. Most of the above 1988 results are also characteristic of days during the other overall very dry year (1976) when $4 < P \leq 8$ mm (Table 3; Fig. 5). Presumably, the delay of the major rainfall deficit in 1976 until July–August (versus May–June in 1988; Fig. 2) led to a better July–August soil moisture and crop status in 1976, which may have supported the higher $E$ and $P_{sv}/P$ for 1976 days when $0.6 < P \leq 4$ mm (Table 3) and prevented less severe 1976 crop yield reductions (Table 1).

Figures 6 and 7 confirm the above suggested role of deficient precursor soil moisture for the strikingly low $P_{sv}/P$ and $E$ values obtained for the 1988 days when $0.6 < P \leq 8$ mm. The occurrence of low $P_{sv}/P$ and $E$ during most of the 1988 events in this $P$ category is clearly apparent in Figs. 6a,b. This association weakens only toward the end of the post–15 July multiday periods when $P > 0.6$ mm day$^{-1}$, apparently reflecting some short-term soil moisture replenishment. Comparison of Figs. 6d and 7b suggest that the three Illinois soil moisture time series in Fig. 6d are representative of the entire study region for 1988. These soil moisture time series reveal that the lack of precipitation through mid-July (Fig. 6a) was accompanied by a monotonic decrease in soil moisture, which, in turn, was associated with generally declining $E$ and $P_{sv}/P$ after early June (Fig. 6c). The overall increases of $E$ and $P_{sv}/P$ through early June (Fig. 6c) presumably reflect initial crop development supported by still relatively high soil moisture (Fig. 6d).

Figures 6a,d and 7b suggest that the increased precipitation after mid-July largely arrested the previous soil moisture decline, but soil moisture remained low through August 1988. The contrasting soil moisture abundance of the much wetter and highly productive 1979 growing season (Table 1) is clearly conveyed by Fig. 7a.

The very poor season-long crop status in 1988 is confirmed strongly by the spatial patterns of crop yield and satellite-derived normalized difference vegetation index (NDVI) anomalies in Fig. 8. Here, the prevalence of deep brown NDVI shading across the core of our study region in August (Fig. 8e) indicates the extremely poor final condition of the corn and soybean crops caused by the devastating spring–early summer 1988 drought. The NDVI spatial pattern across the study region in Fig. 8e closely matches the distribution of the resulting county-scale 1988 corn yields (Fig. 8b) and Crop Reporting District corn and soybean yield anomalies (Figs. 8c,d,f). The very low 1988 $E$ was confirmed by surface eddy correlation measurements for a location near the center of our study region (Kunkel 1989), and by CERES-Maize plant model simulations for a region similar to our entire region (Kunkel 1992) and for the western half of our region (Tsvetsinskaya et al. 2001a,b).

According to Riebsame et al. (1991, p. 56), the losses and costs caused by this 1988 drought included $15$ billion in reduced farm production, $4$ billion in extra agricultural services (including crop insurance and emergency feed assistance), and $10$ billion in food price increases. These losses/costs provide further evidence of the very poor 1988 crop status documented in Fig. 8.

The above strong daily $P_{sv}/P$ dependence on $P$ is completely masked by the averaging process on the standard monthly and seasonal time scales. For example, the seasonal mean $P_{sv}/P$ ratios (Fig. 2) are confined to a remarkably narrow intermediate range (0.20–0.23) and are poorly related to $P$. The same is largely true of the monthly mean $P_{sv}/P$ ratios in Fig. 2; although these have extremes of 0.14 and 0.26, 69% are between 0.19–0.24. The correlations between monthly moisture budget components and related parameters in Table 2 show $P_{sv}/P$ to be significantly related (at the 5% level) to only $E$ ($+0.54$) and IF/A ($-0.51$), which is consistent with Eq. (6). These correlation magnitudes increase (to $+0.62$ and $-0.70$, respectively) and become statistically more significant (1% and almost 0.1%, respectively) with seasonal cycle removal, further emphasizing the strength of the $P_{sv}/P$ versus IF/A association. The extremely weak monthly relation of $P_{sv}/P$ to $P$ ($-0.06$; Table 2; $-0.10$ with seasonal cycle removed) is illustrated further by the mean $P_{sv}/P$ ratios, 0.22–0.25, for the four driest months (July 1975, August 1976, and May–June 1988) approaching 0.26 for the wettest month (July 1979). The complex time-scale interactions that underlie these differences between the monthly/seasonal and daily results are discussed in section 4 below.
c. Comparison with previous recycling results for the central United States

To place these new $P_e/P$ results for the world’s most productive agricultural region (the Corn Belt) in perspective, we next compare them with previous estimates for various regions in the central United States. This comparison recognizes that $P_e/P$ generally increases with the horizontal dimension of a study region along its mean flow [Eqs. (2) and (6)]. Comprehensive comparative information is presented in the appendix. To summarize, the above $P_e/P$ ratios on all time scales are 1) greater than Benton et al.’s (1950) multiyear (1940s) annual average $P_e/P$ estimate for the (much larger) entire Mississippi watershed ($<0.10$), obtained using pan evaporation, precipitation gauge, and pibal/radiosonde data in conjunction with air mass discrimination; 2) generally less than Brubaker et al.’s (1993) multiyear (1963–73) growing season monthly mean $P_e/P$ estimates for a larger ($1.98 \times 10^6$ km$^2$) central United States region (0.22–0.34), made using objectively analyzed rawinsonde data, despite that study’s $P_e/P$
formulation inherently yielding smaller values (by 59%–66%; see section 2a and the appendix);

3) also mostly less than Bosilovich and Schubert’s (2001) 15-yr monthly mean $P_E/P$ estimates obtained for May–August for the above Brubaker et al. (1993) central United States region (0.28–0.39), and individual monthly estimates for the extremely dry 1988 (0.26–0.63) and much wetter 1993 (0.20–0.31) for the same area/season, calculated by applying a recycling equation (from Eltahir and Bras 1994) that is similar to our equation (see below and the appendix) to reanalysis data from the National Aeronautics and Space Administration (NASA) Goddard Earth Observing System (GEOS-1) Data Assimilation System;

4) substantially smaller than Kunkel’s (1989) midwestern $P_E/P$ inferences made from single-point surface $E$ measurements (July–August 1988) for the center of our domain (>0.50); and

5) also substantially smaller than Koster et al.’s (1986) $P_E/P$ inferences made from a single July GCM “tracer” simulation, for which the terrestrial $E$ contribution to $P$ for two midwestern GCM grid cells (spanning 8° latitude × 10° longitude, 0.79 × 10⁶ km² per cell) was from the (much larger) entire North American continent (0.62–0.80).

Particularly notable among the above comparisons are the large differences between our $P_E/P$ estimates for 1988 and the much higher values obtained by Bosilovich and Schubert 2001, hereafter BS) for that season. However, it is unlikely that these differences stem from our respective recycling equations. Their application of the Eltahir and Bras (1994) equation was on a grid cell basis, rather than in a bulk sense for the entire region (see the appendix). Eltahir and Bras’s (1994) comparison of these approaches for the Amazon basin produced a higher recycling value for the bulk application (0.29), which boils down to the use of our Eq. (6), than for the grid cell application (0.25). Instead, several other factors may have contributed to these recycling differences between BS and present studies. The BS region is 57% larger than our region and extends considerably farther west (to the Rocky Mountains) and south (almost to the Gulf Coast), and does not include the most northern and eastern parts of the Corn Belt. Using typical values of $E$ and IF, we estimate that a 57% domain size increase would inflate our $P_E/P$ estimates by 40%–45%. As noted by BS, the restriction of their analysis to the monthly time scale excludes diurnal- or synoptic-scale variability that likely is important for land–atmosphere interactions in the central United States (Schubert et al. 1998). Through our $P$-category stratification of daily $P_E/P$ estimates, we have provided insight on these shorter time scales. Perhaps the main agreement between BS’s results and our 1988 results is the relatively low recycling suggested for August.

4. Concluding remarks

The above results for the globally important midwestern precipitation and crop yields suggest that midlatitude, midcontinent land–atmosphere interactions are complex and reflect plant behavior and important timescale interactions. Most notably, the interannual near-constancy of the seasonal $E$ and $P_E/P$ averages surprisingly fails to mirror the large (and similar) interannual variability in seasonal $P$ and crop yields (Fig. 2; Table
Fig. 8. Patterns of crop yield, crop yield anomalies, and the NDVI (the “greenness” index) anomalies for the study region for 1988. Rectangular boxes enclose the study region. (a) The average corn yield (tonnes ha$^{-1}$) for the USDA north central region for 1972–91 (from Gage 2003). (b) The 1988 corn yield (tonnes ha$^{-1}$) for the USDA north central region (from Gage 2003). (c) The 1988 soybean yield anomaly ($\sigma$ from detrended 1972–90 mean; see section 2c). (d) The 1988 corn yield anomaly ($\sigma$ from detrended 1972–90 mean; see section 2c). (e) The NDVI anomaly for Aug 1988 (from NASA Goddard Space Flight Center Scientific Visualization Studio, available online at http://svs.gsfc.nasa.gov/stories/drought/na.html), where deeper brown (green) indicates lesser (greater) greenness. (f) The 1988 combined corn–soybean yield anomaly ($\sigma$ from detrended 1972–90 mean; see section 2c). Panels (a) and (b) have county-scale resolution; (c), (d), and (f) have USDA Crop Reporting District resolution (9 per state, except 6 for Tennessee and Kentucky); and (e) has a 1.1-km resolution [based on (Advanced Very High Resolution Radiometer) AVHRR/2 imagery].
Superficially, this suggests that vegetation and soil moisture status are less important for \( P \) than is widely assumed (Shukla and Mintz 1982; Kunkel 1989; Oglesby and Erickson 1989; Rind et al. 1992; Chahine 1992a; Brubaker et al. 1993; Beljaars et al. 1996; Hahmann and Dickinson 1997, 2001; Zeng and Neelin 2000). However, these seasonal results mask several crucial subtleties evident in the daily data.

One subtlety explains the surprisingly high seasonal \( E \) for seasons of low \( P \). This counterintuitive association is a major contributor to the interannual near-constancy of seasonal \( E \). The situation arises from the relatively high daily \( E \) for the large number of days when \( P \leq 0.6 \) mm during the dry 1976 and (especially) extremely dry 1988 seasons. Table 3 (category A) and Figs. 4 and 5 show that these days were also characterized by relatively low \( P \) within the \( P \leq 0.6 \) mm day\(^{-1}\) category, associated high SR forcing of \( E \), and pronounced PW decreases indicating significant local storage of \( E \). Clearly, abundant SR sustained high \( E \) during the driest days of the driest growing season during 1972–90 (1988).

Another subtlety in the daily data for individual seasons resolves the apparent paradox involving (i) the close similarity of the seasonal mean \( P_e/P \) ratios for the extremely dry 1988 and very wet 1979 seasons (and also the dry 1976 and overall near-average 1975 seasons; Fig. 2; Table 1) despite (ii) the striking general decrease of daily \( P_e/P \) with increasing \( P \) (Table 3; Fig. 5). Central to this apparent paradox are the surprisingly low seasonal \( P_e/P \) ratios for the very dry 1988 and dry 1976 seasons, relative to their wetter 1979 and 1975 counterparts, respectively. These low 1988 and 1976 seasonal \( P_e/P \) ratios principally stem from low \( E \) on days when agriculturally beneficial \( P \) (4 < \( P \leq 8 \) mm) occurred and significantly contributed to the seasonal \( P \) totals (Table 3, category D; Fig. 5). As already noted, this low \( E \) was likely due to poor crop and precursor soil moisture status (Figs. 6–8) and not low SR (Table 3D; Fig. 5). The low \( E \) was offset by strongly convergent HD and pronounced PW depletion (Table 3, category D; Fig. 4) or, in other words, by the atmosphere compensating for the terrestrial moisture deficiency. These results were also largely true of the 1988 days when 2 < \( P \leq 4 \) mm (Table 3, categories B and C; Figs. 4 and 5). The opposite findings characterized the wetter and higher-yielding growing seasons of 1975 (2 < \( P \leq 8 \) mm day\(^{-1}\)) and especially 1979 (4 < \( P \leq 8 \) mm day\(^{-1}\)), when \( E \) was a more important contributor to \( P \) (Tables 3C,D; Figs. 4 and 5). Note that SR was especially low when 4 < \( P \leq 8 \) mm day\(^{-1}\) during the record/near-record 1979 production of corn/soybeans (Tables 1 and 3 (category D); Fig. 5). Finally, the importance of \( E \) for drought season \( P \) appears to increase and augment HD for days when \( P > 8 \) mm, probably due to widespread abundant surface moisture as suggested above.

Thus, our nontraditional \( P \)-stratified analysis of daily atmospheric moisture budget estimates confirms the widely held view that land–atmosphere interactions are intimately involved in pronounced seasonal regional climate anomalies. This confirmation did not emerge from a more conventional analysis that was limited to seasonal and monthly time scales. The underlying complex time-scale interactions of atmospheric moisture budget components, which involve plant behavior and solar radiation forcing, pose a challenge for the simulation and prediction of the natural environment. The needed modeling capability (e.g., Hong and Kalnay 2000) must be able to treat and interrelate the daily-to-interannual time scales.

Acknowledgments. During the early stages of this research, we benefited from a number of helpful discussions with the late Tzvi Gal-Chen and are pleased to dedicate this paper to his memory. This research was supported by the Climate Dynamics Program of the National Science Foundation and by the National Oceanic and Atmospheric Administration. We thank M. Neil Ward and Eugenia Kalnay for insightful discussions and comments on earlier versions of the manuscript and Tonya Rollins, Sue Weygandt, and Michele Coroiu for important technical assistance. The constructive comments and suggestions of two anonymous reviewers led to several key improvements in the presentation of our methodology and results.
Comparison of the present study with previous investigations of moisture recycling in the central United States. Symbols are as defined previously, except where indicated. Where possible, notation in other recycling equations has been converted to be consistent with our section 2a and recycling equation (below).

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<th>Study</th>
<th>Domain</th>
<th>Time parameters</th>
<th>Basic data sources</th>
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<td>Benton et al. (1950)</td>
<td>Entire Mississippi watershed (3.24 × 10^6 km²)</td>
<td>Data years: 1942–48 (varied</td>
<td>Individual station rawinsonde and pilot balloon wind and humidity measurements;</td>
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<td>Koster et al. (1986)</td>
<td>Two 8° lat × 10^6 Ion GCM grid cells in central United States between 34.5°–42.5°N and 105°–85°W (0.79 × 10^6 km² per cell)</td>
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<td>Monthly mean grid cell values of water vapor traced from 10 source regions that contributed to P for that cell; total P for a cell was equal to the sum of the P contributions traced to the 10 source regions.</td>
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<td>months, sets of days in same P</td>
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</tr>
<tr>
<td></td>
<td></td>
<td>category (within same season and</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>across all seasons)</td>
<td></td>
</tr>
<tr>
<td></td>
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</tr>
</tbody>
</table>
APPENDIX
(Extended)

<table>
<thead>
<tr>
<th>Recycling treatment</th>
<th>Recycling equation</th>
<th>Recycling values</th>
</tr>
</thead>
<tbody>
<tr>
<td>Used moisture budget equations for continental air masses, maritime air masses,</td>
<td>None</td>
<td>“Probably not more than 10 per cent”</td>
</tr>
<tr>
<td>continent, and ocean to deduce percentage of $P$ over the Mississippi Watershed</td>
<td></td>
<td></td>
</tr>
<tr>
<td>that used water vapor from continental and maritime sources; assumed ratio of</td>
<td></td>
<td></td>
</tr>
<tr>
<td>pan $E$ into those two air masses was the same as the ratio of their natural $E$.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Estimated percentage of $P$ from $E$ from the entire North American continent versus</td>
<td>None</td>
<td>Jul = 0.62 (western grid cell)</td>
</tr>
<tr>
<td>surrounding oceanic moisture sources.</td>
<td></td>
<td>Jul = 0.80 (eastern grid cell)</td>
</tr>
</tbody>
</table>

Inference (not a direct calculation) based on estimation of surface energy budget and components of atmospheric water vapor budget (see section 2a).

- Linear generalization of Budyko’s (1974) 1D formulation to 2D, which assumed the flow is parallel and uniform within the region. In Budyko’s equation to the right, $L$ is the length scale of the airstream tube, and $u$ is the velocity normal to the airstream tube.

  $\frac{P_L}{P} = \frac{E}{E + \frac{2IF}{A}}$

  derived from Budyko’s

  $\frac{P_L}{P} = \frac{E}{E + \frac{2PW}{L}}$

- Used bulk diagnostic recycling model after Eltahir and Bras (1994) to calculate recycling ratio for each grid cell in region; recycling equation is similar to our Eq. (6) (section 2a); method yielded a spatially variable (but scale independent?) recycling ratio, values of which were applied across the region to yield regional recycling value; in equation to right, subscripted $L$ denotes local moisture inflow into the grid cells in the region, subscripted $O$ denotes inflow into the grid cell that emanates entirely from outside the region; in the extreme case of the entire region being treated as one grid cell (not done), equation to right reduces to our Eq. (6).

  $\frac{P_L}{P} = \frac{E}{E + \frac{IF_L}{A} + \frac{IF_O}{A} + E}$

  Used following Eltahir and Bras equation for a grid cell

  We extracted manually from their Figs. 2 and 3 the following values for the 15-yr mean, 1988, 1989, 1993, and 1995, respectively:

  - May = 0.23
  - Jun = 0.22
  - Jul = 0.34
  - Aug = 0.33

  May = 0.28, 0.40, 0.25, 0.30, 0.28
  Jun = 0.35, 0.63, 0.28, 0.26, 0.38
  Jul = 0.39, 0.35, 0.58, 0.26, 0.42
  Aug = 0.38, 0.26, 0.29, 0.31, 0.17

  See Table 3 and Figs. 2 and 5.
REFERENCES


International GEWEX Project Office, 1993: Data collection and operational model upgrade. Vol. 1, Implementation plan for the GEWEX Continental-Scale International Project (GCIP), IGPO Documentation Series, No. 6, GEWEX, 83 pp. + appendices.

[Available from International GEWEX Project Office, 1010 Wayne Ave., Suite 450, Silver Spring, MD 20910.]


Trenberth, K. E., 1999: Atmospheric moisture recycling: Role of advection and local evaporation. J. Climate, 12, 1368–1381.


