

ATMOSPHERIC MOISTURE RESIDENCE TIMES AND CYCLING: IMPLICATIONS FOR RAINFALL RATES AND CLIMATE CHANGE

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Abstract. New estimates of the moistening of the atmosphere through evaporation at the surface and of the drying through precipitation are computed. Overall, the e -folding residence time of atmospheric moisture is just over 8 days. New estimates are also made of how much moisture that precipitates out comes from horizontal transport versus local evaporation, referred to as 'recycling'. The results depend greatly on the scale of the domain under consideration and global maps of the recycling for annual means are produced for 500 km scales for which global recycling is 9.6%, consisting of 8.9% over land and 9.9% over the oceans. Even for 1000 km scales, less than 20% of the annual precipitation typically comes from evaporation within the domain. While average overall atmospheric moisture depletion and restoration must balance, precipitation falls only a small fraction of the time. Thus precipitation rates are also examined. Over the United States, one hour intervals with 0.1 mm or more are used to show that the frequency of precipitation ranges from over 30% in the Northwest, to about 20% in the Southeast and less than 4% just east of the continental divide in winter, and from less than 2% in California to over 20% in the Southeast in summer. In midlatitudes precipitation typically falls about 10% of the time, and so rainfall rates, conditional on when rain is falling, are much larger than evaporation rates. The mismatches in the rates of rainfall versus evaporation imply that precipitating systems of all kinds feed mostly on the moisture already in the atmosphere. Over North America, much of the precipitation originates from moisture advected from the Gulf of Mexico and subtropical Atlantic or Pacific a day or so earlier.

Increases in greenhouse gases in the atmosphere produce global warming through an increase in downwelling infrared radiation, and thus not only increase surface temperatures but also enhance the hydrological cycle, as much of the heating at the surface goes into evaporating surface moisture. Global temperature increases signify that the water-holding capacity of the atmosphere increases and, together with enhanced evaporation, this means that the actual atmospheric moisture should increase. It follows that naturally-occurring droughts are likely to be exacerbated by enhanced potential evapotranspiration. Further, globally there must be an increase in precipitation to balance the enhanced evaporation but the processes by which precipitation is altered locally are not well understood. Observations confirm that atmospheric moisture is increasing in many places, for example at a rate of about 5% per decade over the United States. Based on the above results, we argue that increased moisture content of the atmosphere therefore favors stronger rainfall or snowfall events, thus increasing risk of flooding, which is a pattern observed to be happening in many parts of the world. Moreover, because there is a disparity between the rates of increase of atmospheric moisture and precipitation, there are implied changes in the frequency of precipitation and/or efficiency of precipitation (related to how much moisture is left behind in a storm). However, an analysis of linear trends in the frequency of precipitation events for the United States corresponding to thresholds of 0.1 and 1 mm/h shows that the most notable statistically significant trends are for increases in the southern United States in winter and decreases in the Pacific Northwest from November through January, which may be related to changes in atmospheric circulation and storm tracks associated with El Niño–Southern Oscillation trends. It is suggested that as the physical constraints on precipitation



apply only globally, more attention should be paid to rates in both observations and models as well as the frequency of occurrence.

1. Introduction

Characterizing all aspects of the hydrological cycle accurately from observations and analyses is a difficult task, so that there remain substantial uncertainties in precipitation, evaporation, moisture transport in the atmosphere and surface runoff (e.g., Trenberth and Guillemot, 1996b, 1998). These uncertainties become magnified in attempts to project what changes may occur in any of these quantities as the climate changes. Although comprehensive climate system models can be used for these tasks, the complexity of the changes that occur, for example in precipitation, with an increase in carbon dioxide concentrations in the atmosphere, is considerable for a single model. And the complexity becomes even greater when results from different models are compared (e.g., Mitchell et al., 1987; Boer, 1993).

There are reasons for this complexity, of course. Precipitation processes often occur on scales not adequately resolved by climate models. Moreover, observed precipitation fields have tremendous structure on very small scales and there is a large essentially random component to them, so that the details are not reproducible in models run under the same climate forcings, and thus the details are not predictable. But some changes with larger-scale structures can be anticipated; examples include changes in the overall hydrological cycle and changes in monsoons and storm tracks.

How should rainfall, or precipitation, change as climate changes? Why are the patterns predicted from different models under increased greenhouse gas scenarios so different? What is the relationship among changes in evaporation, changes in moisture content of the atmosphere, and changes in precipitation? What are the factors that should be taken into account to explain the changes? The IPCC 1995 report (IPCC, 1996) in dealing with future climate prospects with increased greenhouse gases in the atmosphere states that 'Warmer temperatures will lead to a more vigorous hydrological cycle;² this translates into prospects for more severe droughts and/or floods in some places. . . ' 'Several models indicate an increase in precipitation intensity, suggesting a possibility for more extreme rainfall events'. This paper attempts to explain the processes involved that influence these results and addresses some of the questions by noting the importance of rainfall rates (or intensity) and rainfall frequency, not just accumulated amounts. We further examine the relative roles of moisture stored in the atmosphere, its advection, resupply, and how these may change.

One problem with using models for climate change experiments is that the models contain biases in their control climates, and so the observed rainfall patterns may not be well reproduced. Some models include a 'flux adjustment' to

help compensate for those biases and the modeling strategy is to take the difference between the experiment and the control climate to provide an estimate of the change in climate. The advantage is that this differencing technique removes effects of flux adjustment as well as systematic errors that are common to both runs. However, a problem that can not be alleviated by this approach occurs, for instance, if the control climate produces no rainfall in an area where it should occur (the rains may be in the wrong location). Then it is impossible for the rainfall to be reduced in a climate experiment and the only possible outcome is no change or an increase in rainfall. This example highlights the fact that the precipitation simulated in the model climate is every bit as important as the change that occurs, and so the change from one model to the next should not be expected to be the same even for the same change in climate forcing, unless they both accurately simulate the current climate. Mitchell et al. (1987) show that the regional response of climate models to a perturbation is highly dependent on the control simulation, especially for hydrological variables. This point has often not been adequately appreciated in interpreting results. Meehl and Washington (1988) show how the soil moisture sensitivity in two climate models depends on their differing control climate simulations.

Another important aspect (not the subject of this paper but briefly discussed for completeness) is the role of atmospheric circulation changes. Changes in natural modes of the atmospheric circulation have been documented and may be linked to anthropogenic climate change. In particular, the North Atlantic Oscillation (NAO), the Pacific-North American (PNA) teleconnection pattern, and El Niño-Southern Oscillation (ENSO) combine to influence the planetary wave structure over the Northern Hemisphere such that most temperature changes in recent years have been of warming over North America and Eurasia but cooling over the northern oceans (Hurrell, 1996). The NAO has been at exceptionally high levels for most of the past 20 years (Hurrell, 1995) (but not in the 1995–1996 winter), while ENSO has shown a statistically significant preference for the El Niño phase in the same period (Trenberth and Hoar, 1996). One way in which rainfall patterns can change in midlatitudes is through a shift in storm tracks associated with teleconnections. A dipole pattern of change is found over Europe, with lower rainfalls over southern Europe and wetter conditions in Scandinavia (Hurrell, 1995), as the NAO has been in a more positive phase. In the Pacific, pronounced changes occur in storm tracks over the North Pacific in association with ENSO and the PNA (Trenberth and Hurrell, 1994) leading to a dipole pattern of precipitation anomalies that extends to California at times and which has a component over the southeastern United States, with enhanced activity and rainfall to the south and diminished rainfalls to the north. Moreover, floods and droughts in different locations are associated with ENSO around the globe through teleconnections. Trenberth and Guillemot (1996a) show how storm tracks changed across North America to help bring about the spring-summer 1988 drought and 1993 floods. Therefore, the presence of natural

variability combined with possible changes from anthropogenic climate change should make the patterns of rainfall changes rather complex.

Increases in greenhouse gases increase the radiative forcing and thus not only increase surface temperatures but also enhance the hydrological cycle as much of the heating at the surface goes into evaporating surface moisture. Global temperature increases signify that the water holding capacity of the atmosphere increases and, together with enhanced evaporation, the actual atmospheric moisture increases, as is observed to be happening in many places (Hense et al., 1988, Gaffen et al., 1991, 1992; Gutzler, 1992; Ross and Elliot, 1996). Of course, enhanced evaporation depends upon the availability of sufficient surface moisture and, over land, this depends on the existing climate. This factor is one further complication in terms of the expected response and is likely to be model dependent. However, it follows that naturally-occurring droughts are likely to be exacerbated by enhanced potential evapotranspiration, both in nature and in models (regardless of whether they are very realistic), e.g., Gregory et al. (1997). But what are the processes by which rainfall is changed? Globally there must be a balance between evaporation and rainfall, but what is the relationship locally?

It is argued that increased moisture content of the atmosphere favors stronger rainfall events and thus increased risk of flooding, and this is also observed to be happening in many parts of the world (Karl et al., 1996). It is further argued that one reason why we should expect spotty increases in rainfall is because of mismatches in the rates of rainfall versus evaporation. It is shown that precipitating systems of all kinds feed mostly on the moisture already in the atmosphere, and so the moisture content directly affects rainfall or snowfall rates, but not so clearly total precipitation, at least locally. These aspects are explored by examining the cycling times of moisture into and out of the atmosphere, which we refer to as 'moistening' or 'restoration' and 'drying' or 'depletion' rates, and how much moisture that precipitates comes from horizontal transport versus local evaporation. The latter is often referred to as 'recycling' (see Eltahir and Bras, 1996, for a review). New estimates are provided of the residence times of moisture in the atmosphere and of recycling using new global precipitation datasets and analyses (the reanalyses) from the National Centers for Environmental Prediction (NCEP) using data from NCAR (Kalnay et al., 1996). The seasonal variations in recycling and a more complete discussion of the results is given in Trenberth (1998). The trends in frequency of precipitation are also examined, as it is the combination of frequency and intensity that makes up the total accumulated amounts.

A discussion of the datasets used is given in Section 2. Section 3 provides the main new results relative to atmospheric moisture cycling. It includes estimates of moistening rates through evaporation and drying rates through precipitation of the atmosphere, precipitation rates, and recycling estimates. These provide a basis for the overall arguments on how climate change influences the hydrological cycle and precipitation which is discussed in Section 4 along with the analysis of trends

in frequency of precipitation. The conclusions are given in Section 5 along with a recommendation for new dataset development.

2. Datasets

The sources and sinks of atmospheric moisture are evaporation and precipitation. Climatological values are quite uncertain for both fields, and time series of monthly means globally are only now just becoming available but require evaluation to ascertain their accuracy and usefulness. One source of precipitation data newly available is from the Global Precipitation Climatology Project (GPCP) (Huffman et al., 1997), for the period after 1987 and these are extended to cover the period after 1979 by Arkin and Xie (1994) and Xie and Arkin (1996, 1997), and called the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP). Over land these fields are mainly based on information from rain-gauge observations, while over the ocean they primarily use satellite estimates made with several different algorithms based on outgoing longwave radiation (OLR), and scattering and emission of microwave radiation.

Another source is from the NCEP/NCAR reanalyses produced on model (σ) surfaces with T62 spectral resolution and 28 levels in the vertical using four-dimensional-data assimilation, and these also provide estimates of the evaporation and moisture transport. The vertically integrated moisture transports are computed using four-times daily data in model coordinates at full resolution. An evaluation has been carried out of the NCEP moisture fields, the precipitation P and evaporation E , and the moisture transport and divergence in the atmosphere by Trenberth and Guillemot (1996b, 1998). The precipitable water w was compared with analyzed fields from NVAP (Randel et al., 1996) based primarily on Special Sensor Microwave Imager (SSM/I) over the oceans and rawinsonde measurements over land, plus TIROS Operational Vertical Sounder (TOVS). The moisture budgets were evaluated through computation of the vertical freshwater flux at the surface $E - P$ using residual techniques from the divergence of the total moisture transport, and this was compared with the reanalysis $E - P$ that is based upon a 6-hour integration of the assimilating model and thus depends on the model parameterizations. The P field was evaluated using Xie-Arkin CMAP estimates and, although it contains considerable uncertainties (Xie and Arkin, 1997), the patterns are believed to be reliable and good enough to show that there are substantial biases in the NCEP P .

The NCEP moisture fields contain significant biases in the tropics, the tropical structures are less well defined, and there is an underestimate of the variability from year to year. The NCEP model P generally reveals a pronounced double intertropical convergence zone in the central Pacific and the location of the South Pacific Convergence Zone is not well captured. Rainfall amounts are lower than observed in the oceanic tropical convergence zones. The variability in the central tropical Pa-

cific of P associated with El Niño-Southern Oscillation (ENSO) is underestimated in the NCEP reanalyses and, moreover, is not very well correlated with the CMAP product. A bias for too much rainfall in the model over the southeastern U.S. and southeast Asia is also present in northern summer. The comparison of $E - P$ from the moisture budget with the model result reveals some strong systematic differences. Biases in E are inferred in some places from the $E - P$ differences, and they probably arise from spurious land moisture sources in some cases.

Therefore we will make use of the monthly CMAP fields of precipitation from Xie and Arkin, precipitable water w from NVAP and, in spite of the deficiencies noted above, E and moisture transport values from the NCEP reanalyses. These are adequate for current purposes, although quantitative details should not be trusted.

We also make use of a new dataset for the United States of hourly precipitation rates (Higgins et al., 1996) to explore rainfall frequencies and how they have changed with time. The data are gridded onto a 2° latitude by 2.5° longitude grid, and results performed of the trend analysis were quite noisy, with bull's-eye grid-scale features emerging that were not robust with different thresholds of what counted as a rain event. Accordingly, a nine-point (3 by 3) smoother was used to bring out the larger spatial scales before the analysis was performed. This smoother removes 2 grid wavelength features entirely in both directions and damps 4 (8) grid wavelength features to 23% (71%) of their original amplitude. Accordingly, only scales larger than about 10° latitude are included.

3. Atmospheric Moisture Cycling

3.1. EVAPORATION AND PRECIPITATION

Climatological precipitation fields are much smoother than individual daily or even monthly mean fields. The annual mean P field for 1979–1995 is shown in Figure 1. This exhibits much sharper gradients and more prominent structures than the evaporation field (Figure 2) which has typical values of 3 to 6 mm/day, or less at high latitudes. The evaporation field does contain much stronger gradients in the northern winter, especially between the northern continents and oceans (e.g., see Trenberth, 1998). As noted by Trenberth and Guillemot (1995), the temporal variability of E is also much less than that of P and so the latter dominates both the spatial and temporal variability of the fresh water flux $E - P$. Average precipitation rates range from almost zero to over 10 mm/day (Figure 1), but as rain typically occurs only about 0 to 20% of the time, actual rates conditional on rain falling can average 10 mm/h or more.

This inferred imbalance in the drying versus moistening of the atmosphere implies that most of the moderate and heavy rain that falls comes from the precipitable water in the atmosphere, not directly from evaporation, and so the lifetime of moisture in the atmosphere and its availability to rain systems is a limiting factor

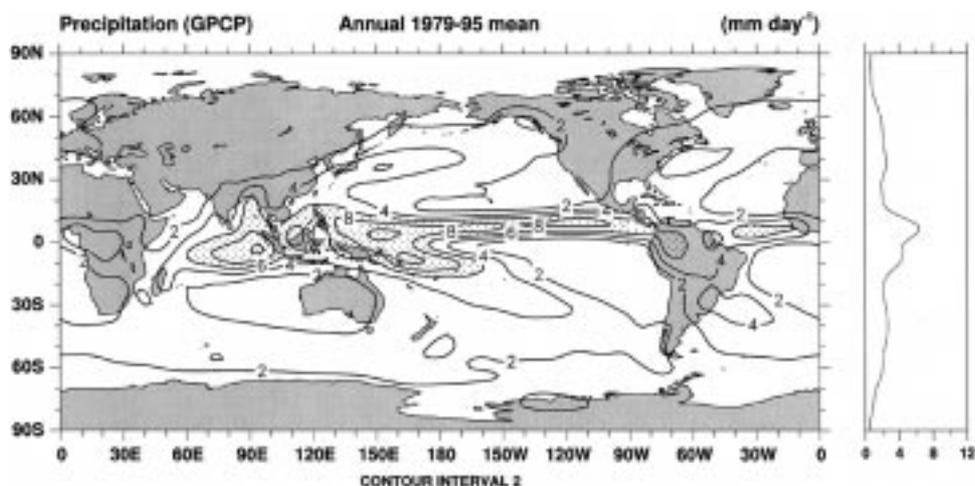


Figure 1. Annual mean precipitation from the Xie-Arkin product for 1979–1995 in mm/day. All quantities have been truncated to T31 for presentation in all figures, the contour interval is given below each plot and the units are at upper right. At right the zonal mean meridional profile is given. Values exceeding 6 mm/day are stippled.

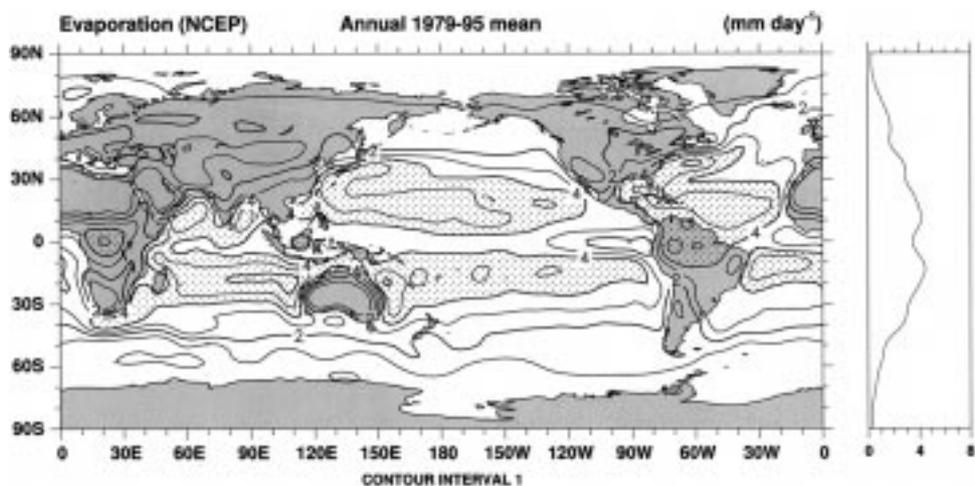


Figure 2. Annual mean evaporation from 1979 to 1995 from the NCEP reanalyses, based upon 6 hour model integrations with the assimilating model. Units mm/day. Values exceeding 4 mm/day are stippled.

on precipitation. The annual mean w is shown in Figure 3. This argument suggests that rainfall rates and the rate of restoration of moisture should be examined in more detail.

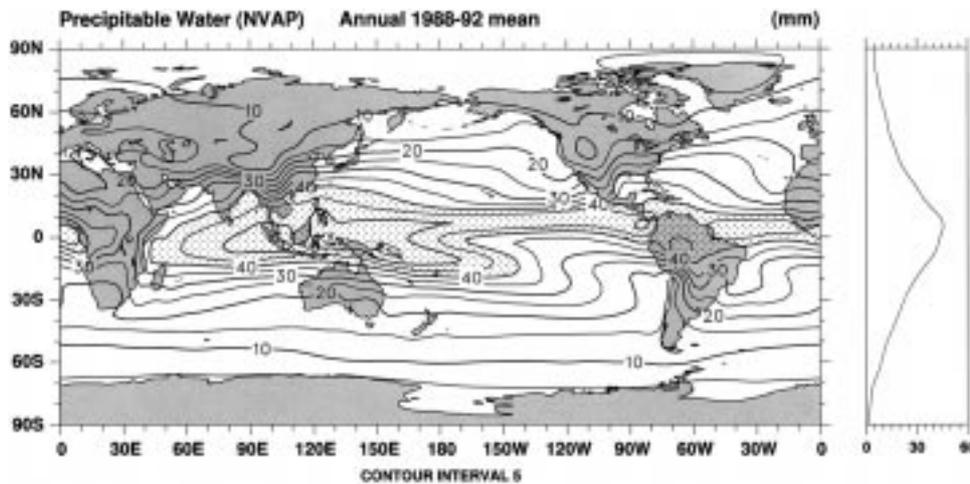


Figure 3. Annual mean precipitable water for the period 1988–1992 from NVAP in mm. Values exceeding 40 mm are stippled.

3.2. RESTORATION AND DEPLETION RATES OF ATMOSPHERIC MOISTURE

If it is assumed that the precipitable water change depends on the precipitable water w itself, which we have argued is the case for precipitation, then

$$\frac{\partial w}{\partial t} = -\lambda w \quad (1)$$

and λ is the e -folding inverse time constant. The assumption ignores the important role of moisture transport which is taken up in subsection 3.4. If the left hand side corresponds to P , then $\lambda = -P/w$ and λ^{-1} is the e -folding time constant for depletion of precipitable water by precipitation.

A similar expression to (1) can be considered to apply for E but λ is of opposite sign as the e -folding recharge rate. This may be less meaningful, as E is only partly dependent upon w , but the comparison is nevertheless of general interest. Because horizontal transport typically restores moisture amounts locally, it is desirable to smooth λ somewhat; here results are presented at T31 resolution.

The two time constants have been computed using the NVAP w , CMAP P and NCEP E and are shown in Figure 4. Where P or E are very small, the implied time constants are very long, and so the largest contour plotted is restricted to 30 days (locally values can exceed 100 days, although these are artificial as they result in part from using time averages of the fields, rather than time averages of the ‘time constants’). Zonal averages are found by averaging the λ values and taking the reciprocal to get the time constant, see Figure 5, and it shows time constants of around one to two weeks. For precipitation, local values of the depletion rate are about 5 days in the tropical convergence zones but they exceed a month in the dry zones in the subtropics and desert areas. The pattern has a strong resemblance to the P values themselves. The restoration rate shows largest values over

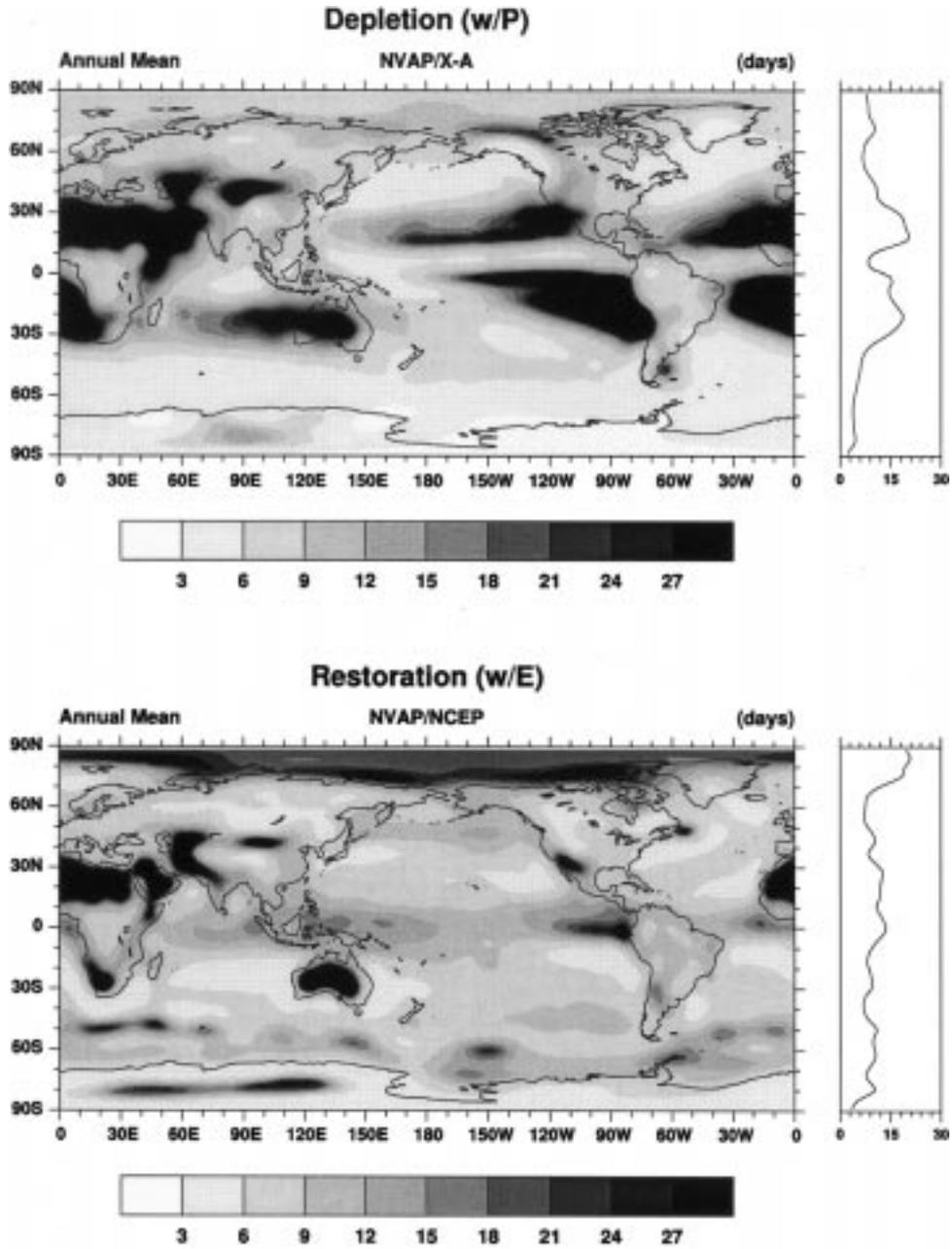


Figure 4. Average *e*-folding time constants for depletion of moisture by precipitation and for restoration of moisture by evaporation of precipitable water in the atmosphere, based upon NVAP estimates of precipitable water (Figure 3) and the fields in Figures 1 and 2.

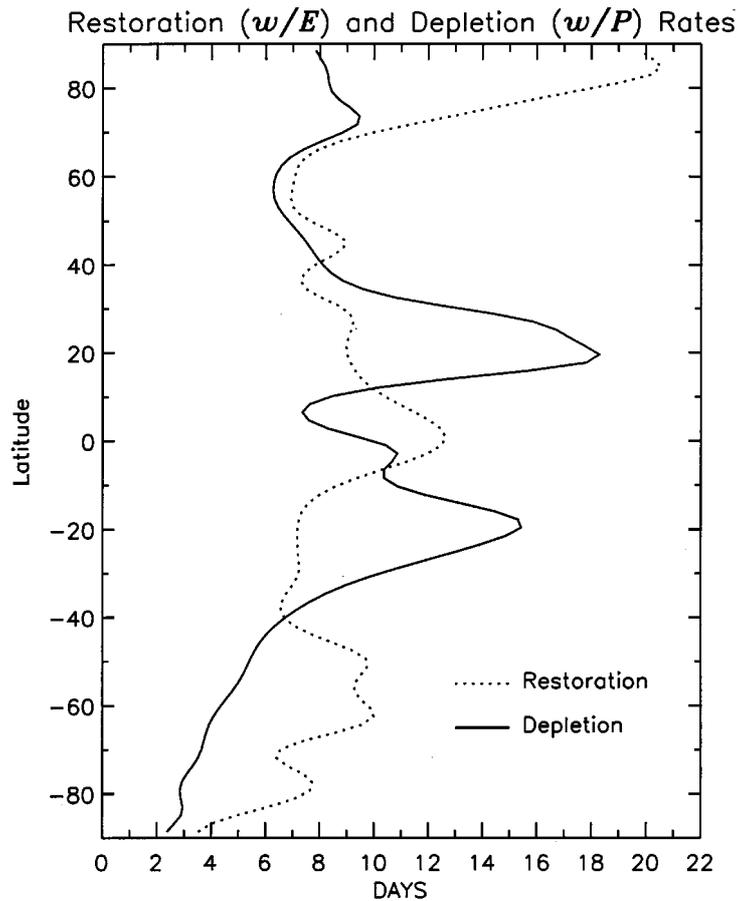


Figure 5. Zonal average time constants for precipitation (solid) and evaporation (dashed) for depletion and restoration of precipitable water in the atmosphere.

northern Africa, extending across to Saudi Arabia and Iran, and over Australia. Values average about 12 days in the tropical convergence zones and are lowest in the subtropical highs where evaporation is a maximum, yet moisture is trapped at low levels by subsidence and so precipitable water is limited. In many respects the restoration and depletion rates are mirror images of each other as the main evaporative sources occur in dry regions. Moisture transport accounts for the apparent inconsistencies. At high latitudes, values are unreliable as w , P and E are small and not well known.

Globally, these two λ values should be the same, although only if λ is computed from global means; there is no physical constraint on the global mean of λ . Using the global mean fields of w of 24.6 mm, E of 2.78 mm/day and P of 2.69 mm/day the time constants would be 9.1 days for precipitation (CMAP) and 8.9 days for evaporation (NCEP), showing that these fields are not fully compatible. As the P

TABLE I

Global mean values $\langle \rangle$ of moisture-related variables for the ocean, land and total areas. The time constants computed from both global means and as global means of the time constants are also given.

Variable	Units	Global	Ocean	Land
$\langle P(\text{CMAP}) \rangle$	mm/day	2.69	3.01	1.92
$\langle P(\text{NCEP}) \rangle$	mm/day	2.73	2.92	2.25
$\langle E(\text{NCEP}) \rangle$	mm/day	2.78	3.17	1.84
$\langle w(\text{NVAP}) \rangle$	mm	24.6	26.5	20.0
$\langle w \rangle / \langle P(\text{CMAP}) \rangle$	days	9.1	8.8	10.4
$\langle w \rangle / \langle E \rangle$	days	8.9	8.4	10.9
$\langle P(\text{CMAP}) / w \rangle^{-1}$	days	8.1	7.9	8.9
$\langle E / w \rangle^{-1}$	days	8.5	8.1	9.6

and E fields come from different sources, there is no guarantee of a global moisture balance. Nevertheless, the values are quite close. Table I provides the global means of the variables as well as their values for just the land and ocean areas. Global mean values of P from NCEP are also given as an indication of how different they are and thus to provide an idea of uncertainties. Differences are less than 2%.

However, the more meaningful physical quantity is the global mean of λ , for which the time constants of the globally-averaged fields are 8.1 days for precipitation and 8.5 days for evaporation (Table I). Note that the precipitation value drops by a day from that computed with global means owing to the heterogeneous P distribution and spatial correlation with w . To gain an idea of how sensitive these values are to different data sets, we use the NCEP P in place of that from CMAP which gives a time constant of 7.6 days. The method of computing these rates uses time-averaged quantities, rather than computing λ from the individual fields and taking the time average of λ . This would take into account correlations between fields, such as the influx of moisture ahead of cold fronts that actually feeds into precipitation. However, these refinements are not dealt with here as they would require an accurate and fully consistent dataset of w and P .

3.3. DURATION AND RATES OF PRECIPITATION

The time constants for depletion and restoration rates of atmospheric moisture are fairly similar overall but do not take account of the fact that rain falls only a small fraction of the time. In the tropics, for example, averaged over a network of stations within 120 km of Darwin, northern Australia, it rains only 1 to 2% of the time in

the pre-monsoon at a rate of 14 mm/h, while during the monsoon, it rains about 16% of the time at a rate of 8 mm/h (Short et al., 1993a,b). The latter includes more and longer periods of relatively light rain. However, as they show, the total accumulated rainfall comes mostly from high rainfall rates, so that 70% of the monsoon and 85% of the premonsoon rains occur at rates exceeding 10 mm/h. At this latter value, the depletion rate is about 3 to 4 hours in the deep tropics.

At individual stations 1 minute rain-rates can exceed 200 mm/h although not for very long, and light rain rates of less than 0.25 mm/h are usually not detectable from many recording gauges. A climatology of rain rates (Jones and Sims, 1978) indicates good reproducibility of rain rates from one year to another even in the presence of varying total rain accumulations. Frequency of rain varied from a low of 1.7% in midlatitude continental interior regions of the Northern Hemisphere to about 5% at many stations, using the 0.25 mm/h cutoff (i.e., light rain periods are not included). In Illinois, Jones and Sims show that 16 minute rain-rate maximum values are about half those of the 1 minute values (60 versus 120 mm/h) and 1 minute and 16 minutes cumulative frequencies are the same for rain rates of 2.4 mm/h. One minute or 'snap-shot' values are especially relevant for radar or satellite estimation techniques (e.g., Short et al., 1993a,b). While these latter applications are now widely used, information on rainfall rates as a function of space and time is not readily available. Clearly the rain rate depends upon the size of the averaging area and the period.

For the United States, we use the dataset of hourly precipitation rates (Higgins et al., 1996) and have chosen a threshold of 0.1 mm during one hour as a useful one for depicting the incidence of precipitation. Results are shown as a percentage for January and July (Figure 6). In January (winter) over 30% of the hours contain precipitation exceeding the threshold in the Northwest. Values are typically around 15 to 20% in the Southeast, and minimum values are found just east of the continental divide of less than 4%. In summer (July), minimum 'precipitation hours' are found in California of less than 2% and maximum values of over 20% are in the Southeast.

Therefore the depletion rate is about 3 to 4 hours in the deep tropics when rain is falling. In middle latitudes, typical unconditional rainfall rates are 3 mm/day, but with rain falling about 10% of the time and precipitable water amounts of 15 mm, the depletion rate of 5 days drops conditionally on rain falling to about 12 hours. This emphasizes that moderate or heavy precipitation feeds mostly on the moisture already in the atmosphere.

3.4. RECYCLING OF MOISTURE

Another aspect of the problem is revealed by examining how much evaporation in an area contributes to the precipitation in the same area, known as recycling of moisture. As the area is reduced to a point the evaporation contribution tends to zero and all the moisture precipitated is transported in. At the other extreme,

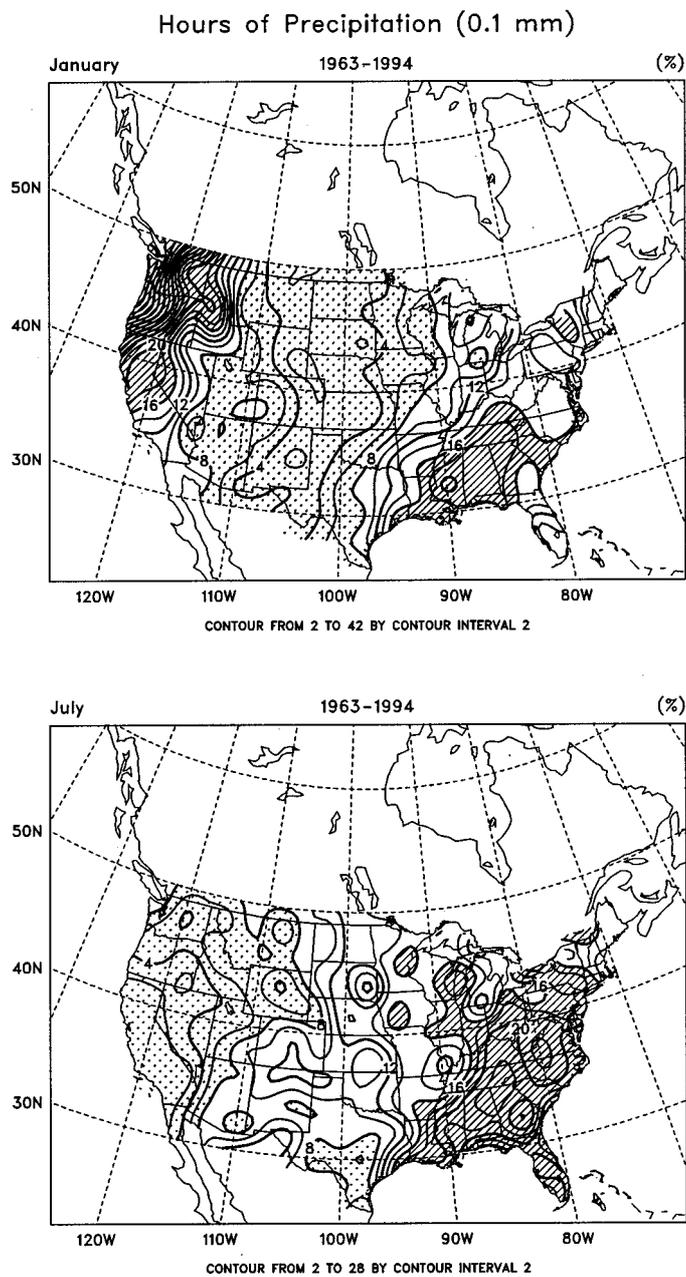


Figure 6. Percentage of hours with greater than 0.1 mm of precipitation for January and July. Values exceeding 16% are hatched and values less than 8% are stippled.

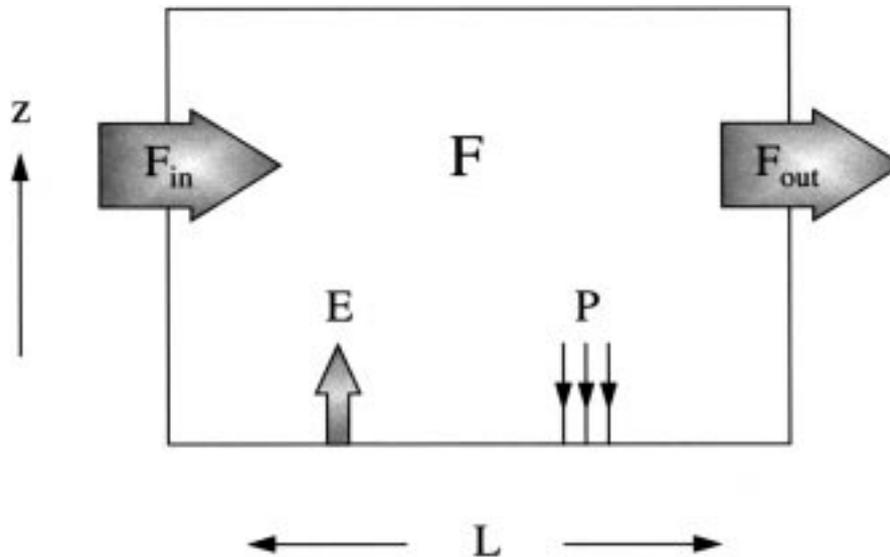


Figure 7. Schematic for the processes considered in recycling of moisture.

as the domain becomes global, the evaporation entirely replaces the precipitation. Thus the recycling ranges from 0 to 1 depending on the size of the area. Eltahir and Bras (1996) review estimates of precipitation recycling and Eltahir and Bras (1994) estimate that 25 to 35% of the rain that falls in the Amazon basin is contributed by evaporation within the basin (over scales of about 2500 km). In the Mississippi basin they find that recycling estimates range from 10% to 24% over about 1500 km scales.

Approximate values of recycling are computed following the approach of Brubaker et al. (1993) which makes assumptions that seem to be more justifiable than those in the formulae of Eltahir and Bras (1996) (see Trenberth, 1998). We assume equilibrium conditions so that there are no changes in atmospheric moisture content. Consider a domain of length L aligned along the trajectory of the air with a flux of moisture into the box of F_{in} and a flux out of F_{out} (Figure 7) and a total evaporation E and precipitation P in the box, then

$$F_{\text{out}} = F_{\text{in}} + (E - P)L \quad (2)$$

and the average horizontal moisture flux through the box is

$$F = 0.5(F_{\text{in}} + F_{\text{out}}) = F_{\text{in}} + 0.5(E - P)L. \quad (3)$$

If $P = P_a + P_m$, where P_a is the advective component and P_m is the component of precipitation arising from local evaporation, then the average horizontal flux of advected moisture over the region is $F_{\text{in}} - 0.5P_aL$ and the average horizontal flux of locally evaporated moisture is $0.5(E - P_m)L$.

An important assumption is that the atmosphere is well mixed so that the ratio of precipitation that falls arising from advection versus local evaporation is equal to the ratio of average advected to evaporated moisture in the air. Thus

$$\frac{P_a}{P_m} = \frac{F_{in} - 0.5P_aL}{0.5(E - P_m)L} \quad (4)$$

which is readily solved to give $\frac{P_a}{P_m} = \frac{2F_{in}}{EL}$ so that the recycling ratio ρ can be written

$$\rho = \frac{P_m}{P} = \frac{EL}{EL + 2F_{in}} \quad (5)$$

or, using (2)

$$\rho = \frac{EL}{PL + 2F} \quad (6)$$

This expression involves several other quantities that are of interest. The fraction of water vapor over a region that participates in the hydrological cycle $I = \frac{PL}{F}$ has been called the ‘intensity of the hydrological cycle’ by Drozdov and Grigor’eva (1965), although a more descriptive term might be the ‘precipitation efficiency’ as it gives the fraction of moisture flowing overhead that is precipitated out. However, this term is already used in thunderstorm literature with a slightly different meaning (see Section 4). Similarly, we can define a ‘moistening efficiency’ $M = \frac{EL}{F}$ as the fraction of moisture added. Then from (6), $\rho = \frac{M}{2+I}$.

This formulation has an advantage in that it can be evaluated locally, although formally the results apply to an area with a scale L . All these expressions for I , M , and ρ depend upon L . Note also from (6) that as $F \rightarrow 0$ then $\rho = E/P$. The scale dependence is explored in more detail in Trenberth (1998) which also examines the recycling as a function of season and documents the annual cycles in M and I as well.

Here we present annual mean results for L of 500 km. While interest has often been on estimates of ρ for large river drainage basins, the heterogeneity of the land surface is such that the recycling clearly varies substantially over the basins. All the computations were done locally in terms of I and M and have been smoothed to a T31 resolution. The above formulae are approximate, as they assume uniformity that does not exist in practice and Equations (2) and (3) do not hold exactly for the datasets that we have. Clearly, some of the assumptions made above are questionable. Note that the whole concept of recycling is most useful over land where moisture for evaporation is limited by the precipitation, whereas over the oceans the surface is wet regardless of whether it rains or not. In addition, as we assume that there is no change in moisture storage in the atmosphere, the results are most useful for a mean annual cycle. Thus the formulae are applied for annual mean values. Regions of mountains are screened out (where surface pressures are less than 800 mb) from the calculation as those are regions where the moisture flux is

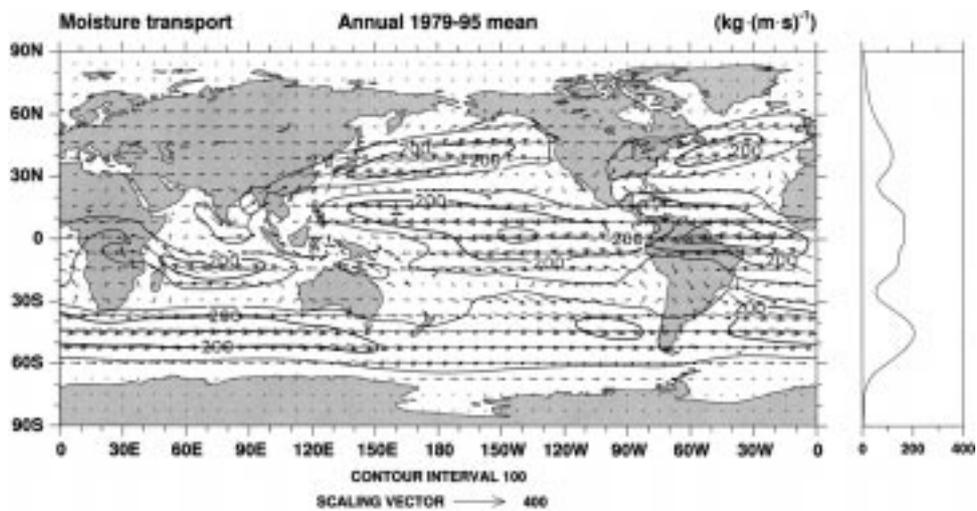


Figure 8. The annual mean vertically averaged moisture transport from the NCEP reanalyses for the period 1979–1995. The scale of the vectors is given below and contours indicate magnitudes in kg (m s)^{-1} .

small and there are huge variations over short distances owing to orographic effects on rainfall. Values for recycling vary substantially when the formulae are applied with seasonal mean data. This raises questions about how (6) should be applied and the effects of seasonal variations, but these are pursued in Trenberth (1998).

The recycling fraction depends greatly on the magnitude of the total moisture flux (Figure 8) which helps explain the patterns of recycling. The recycling results for annual means (Figure 9) reveal recycling percentages for 500 km scales of about 5 to 10% over land typically. The global mean is 9.6%, consisting of 8.9% over land and 9.9% over the oceans. Over the northern Amazon (north of 10°S), the average is about 6% and over the Mississippi basin about 7%. The Amazon results depend greatly upon whether or not the maximum over the southern part of the basin, where $\rho > 20\%$, is included. About half of it is included in the Brubaker et al. (1993) Amazonia region. Also, the latter chose a 'basin' of 25° longitude wide, so that L is over 4 times the value we used, and our results are compatible with the 25 to 35% recycling obtained by previous studies (see Trenberth (1998) for further discussion). In the Mississippi basin, Brubaker et al. used a region 20° longitude long (about 1800 km), for which we would obtain a recycling of about 21% compared with a monthly mean of 24% from Brubaker et al. (1993). It is worth pointing out that the larger values previously obtained for the Amazon versus the Mississippi are mostly a result of the scale of the domain.

Relatively high values ($> 20\%$) of recycling occur either in the subtropical highs, where E is high and the advective moisture flux is small, or in convergence zones where, again, the advective moisture flux is small (Figure 8). Low values

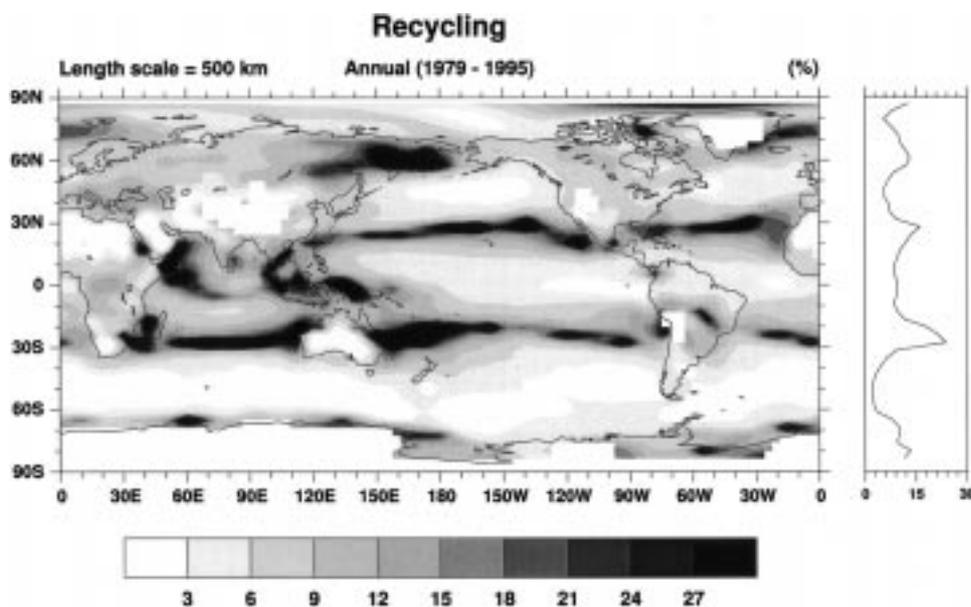


Figure 9. The recycling in percent, for annual mean conditions, computed from (6) for $L = 500$ km, and using E and F from the NCEP reanalyses (Figures 2 and 8) and P from CMAP (Figure 1).

occur over the southern oceans, the North Pacific, and the eastern equatorial Pacific, where the moisture flux is at a maximum (Figure 8).

All of these recycling values show that on average less than 20% of the precipitation that falls comes from evaporation within a distance of about 1000 km. Therefore the results reinforce the arguments given above concerning the importance of transport of moisture and local storage in feeding precipitation.

In the United States, much of the moisture for precipitation, especially in the winter half year, comes from moisture transported out of the subtropics often in a southwesterly flow ahead of cold fronts. For storms east of the Rockies, moisture flows northwards from the Gulf of Mexico or subtropical Atlantic. At advection rates of 12 m s^{-1} (which is the standard deviation of the northward velocity component at 850 mb just north of the Gulf of Mexico in January (Trenberth, 1992)), the moisture travels over 1,000 km in a day, and thus moisture from the Gulf can be readily precipitated out over the Great Plains or Ohio Valley just a day or so later. In major storms, transient northward advection rates often exceed 20 m s^{-1} at 850 mb. In the western United States, the moisture comes from the subtropical Pacific – several examples of this which resulted in flooding in the Pacific Northwest are given in Halpert and Bell (1997), see especially their Figures 37 to 43. Once again it takes only a day or so for the moisture to be transported from the subtropical Pacific to the Pacific Northwest and thus, based on our recycling results, greater than 75% of the moisture precipitated out comes from moisture already in the atmosphere.

4. Relevance to Climate Change

The above analyses show the mismatch between precipitation rates and evaporation so that moderate and heavy precipitation, which contributes most to the total accumulation, depends upon the moisture already in the atmosphere and the advection and resupply of moisture by the storm circulation. These points are pertinent to climate change experiments. However, most climate model studies have not analyzed the results in a way that throws light on these aspects. The surface heat budget is especially relevant because it limits the energy available for evaporation.

There are many feedback processes in nature that can either amplify or diminish the climate response to increases in greenhouse gases. The net forcing or 'warming' at the surface depends critically on these and the surface heat budget. The latter has been explored for three models in some depth by Gutowski et al. (1991) and for the Canadian model by Boer (1993). The most comprehensive analysis seems to be that by Mitchell et al. (1987). Roads et al. (1996) address the sensitivity of the hydrological cycle in the NCAR CCM2 model to differing levels of CO₂ in the atmosphere. For a doubling of carbon dioxide concentrations under equilibrium conditions, there is a surface warming of typically 1.5 to 4.5 °C (IPCC, 1996). In every case it seems that at the surface there is an increase in downwelling infrared radiation associated with both the greenhouse effect from carbon dioxide and changes in water vapor and clouds. In some models, changes in clouds also produce an offset by reducing shortwave radiation, but the net energy available from radiation at the surface is increased in spite of the greater surface emissions associated with the higher temperatures. Moreover, changes in the sensible heat flux also act to warm the surface because of stabilization of the lower atmosphere (Boer, 1993; Roads et al., 1996). This leaves only the latent heat flux through increased evaporation to compensate and balance the surface heat budget. The latent heat flux increase, which ranges from 3 to 10 W m⁻² for the four models considered by Boer (1993), determines the global enhancement of the hydrological cycle and average precipitation rate (of about 3 to 10%). However, the atmospheric moisture content increases by about 20% (Mitchell et al., 1987) or more (in the case of the CCM2, Roads et al., 1996) although with very little change in relative humidity. With other things kept constant, moisture convergence would be enhanced by the same amount and should lead to similarly enhanced precipitation. But a 20% increase in precipitation cannot occur because of the limitations associated with the surface energy budget. Nevertheless such mechanisms should take place for individual storms, whether thunderstorms, or extratropical cyclones, leading to increased rainfall rates. If this is the case, however, there are implications for either the frequency of storms or other factors that must come into play to restrict the total precipitation.

One major factor is that the moisture increases are not uniform. Bigger increases occur in lower latitudes in spite of larger increases in surface temperatures at high latitudes because of the non-linear nature of the Clausius-Clapeyron equation. Thus much of this moisture may not be within reach of many extratropical storms.

Another factor is the 'precipitation efficiency' defined as the ratio of the water mass precipitated to the mass of water vapor entering the storm through its base (e.g., Fankhauser, 1988) or the ratio of total rainfall to total condensation in modeling studies (e.g., Ferrier et al., 1996). These studies show that in typical storms the precipitation efficiency varies from about 20 to 50%, and 30% seems a typical value. The problem is that all the moisture available can not be utilized, thereby creating separable dry air. Dry downdrafts do form in thunderstorms but are often moistened by evaporating rain or cloud drops. Therefore, one interpretation is that the average air leaving the storm or remaining behind has a relative humidity of perhaps 70% versus 100% at cloud base (this ignores all the inhomogeneities from updrafts and downdrafts, and the fact that the air leaving is at a different temperature, etc.). In one cloud/thunderstorm model, greater moisture content in the atmosphere does produce more condensation and greater precipitation efficiency (Ferrier et al., 1996). However, warmer conditions could also imply that more moisture might remain if relative humidity is a key factor, as is likely. Therefore the rainfall may not increase in direct proportion to the moisture convergence because more moisture is left behind.

Another factor relates to extratropical storms and the overall baroclinicity, as argued by Held (1993). In most models, surface temperature increases with increased greenhouse gases are greatest in the Arctic, in part because of ice-albedo feedback, so that the surface temperature gradient and baroclinicity is reduced, although this may not be the case above the surface. Held notes that extratropical storms are greatly influenced by moisture in the atmosphere and that one effect of increased moisture content in the atmosphere is to enhance the latent heating in such storms and thereby increase their intensity. On the other hand, he also notes that more moist air would be transported polewards by transient eddies, reducing the required poleward energy transports normally accomplished by baroclinically unstable eddies and the associated poleward down-gradient heat transports. He therefore argues that this would contribute to 'smaller eddies' and suggests that this means a decrease in eddy amplitudes. While recognizing that both effects are important, Held suspects that the latter is dominant. There are other possibilities not considered by Held. In particular, individual storms could be more intense from the latent heat enhancement, but fewer and farther between. Changes in the vertical temperature structure (the lapse rate) will also play a role in such storms.

Therefore the other major factor worth considering here in more detail is the frequency of precipitation events. The above discussion would suggest that if rain rates increase faster than rain amounts, then the frequency of rain occurrence could decrease. We have examined changes in frequency of precipitation events for the United States using the same dataset as for Figure 6, and two thresholds to check for robustness. One was 0.1 mm per hour, as in Figure 6, the other was 1 mm per hour. The trends in hours of precipitation per decade have been computed over the period 1963 to 1994 for each month, season, and annual values. Results shown

Trend of Precipitation (0.1 mm threshold)

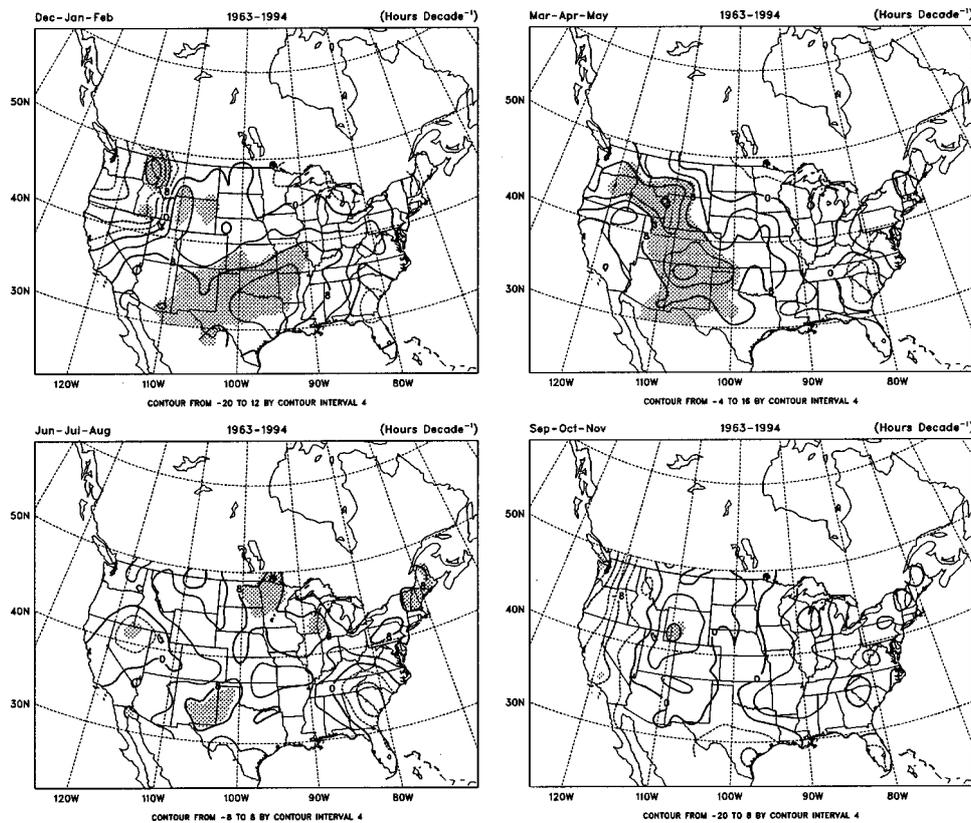


Figure 10. Linear trends in hours with 0.1 mm of precipitation per decade, for the period 1963–1994 for the four seasons. Stippling indicates statistical significance at the 5% level based on a two-tailed *t* test.

here are for smoothed values (see Section 2) that retain scales larger than about 10° latitude.

The results of the analysis of linear trends in the frequency of precipitation events for the United States corresponding to thresholds of 0.1 and 1 mm/h are given in Figures 10 and 11. For 0.1 mm/h there is a predominance of small but mostly positive trends in winter, spring and summer (Figure 10), and the average trend is 0.24 h/year of precipitation over the United States as a whole. But for 1 mm/h, the trends for the country as a whole are close to zero. The most notable statistically significant trends in precipitation frequency are for increases in the south central U.S. in winter and decreases in the Pacific Northwest from November through January, and increases in the South and Southwest in spring and summer. The storm track extending from Texas into the Ohio Valley has been enhanced at the expense of the storm track into the Pacific Northwest, especially in winter and, to some extent, in autumn. These changes appear to be related to changes

Trend of Precipitation (1.0 mm threshold)

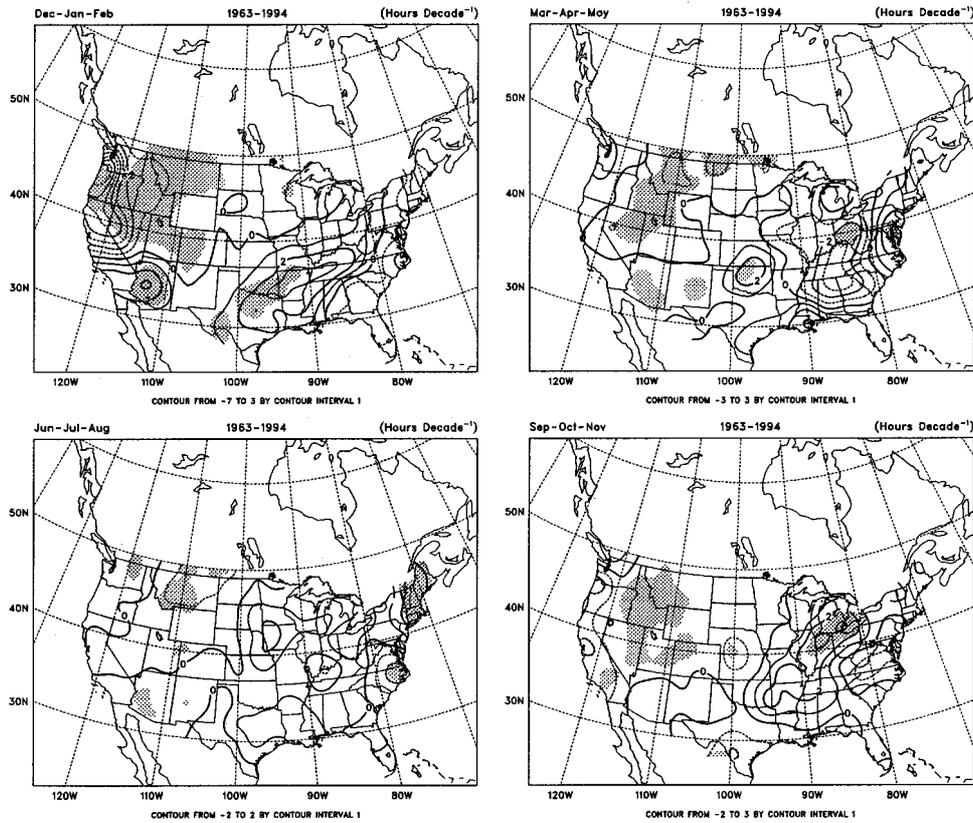


Figure 11. Linear trends in hours with 1 mm of precipitation per decade, for the period 1963–1994 for the four seasons. Stippling indicates statistical significance at the 5% level based on a two-tailed t test.

in storm tracks associated with ENSO trends (e.g., Kumar et al., 1994; Trenberth and Hurrell, 1994; Trenberth and Guillemot, 1996a). In summer, negative trends are significant in some regions in the Southeast. These changes say little with regard to the arguments given above and emphasize that regionally, the changes in atmospheric circulation patterns and especially those associated with ENSO, may produce dominant effects. How ENSO itself changes with global warming is an outstanding issue (Trenberth and Hoar, 1996).

All of these aspects of precipitation change have been explored only to a very limited extent in climate models. An extensive analysis has been done by Mearns et al. (1995) using a nested regional model with 60 km resolution for regions of the United States for current and doubled-carbon dioxide results. They explore the frequency and intensity of precipitation but only for daily values, not the true precipitation rates. Results reveal increased daily rainfall variability under doubled CO_2 . There are some areas where frequency of precipitation decreases but precipi-

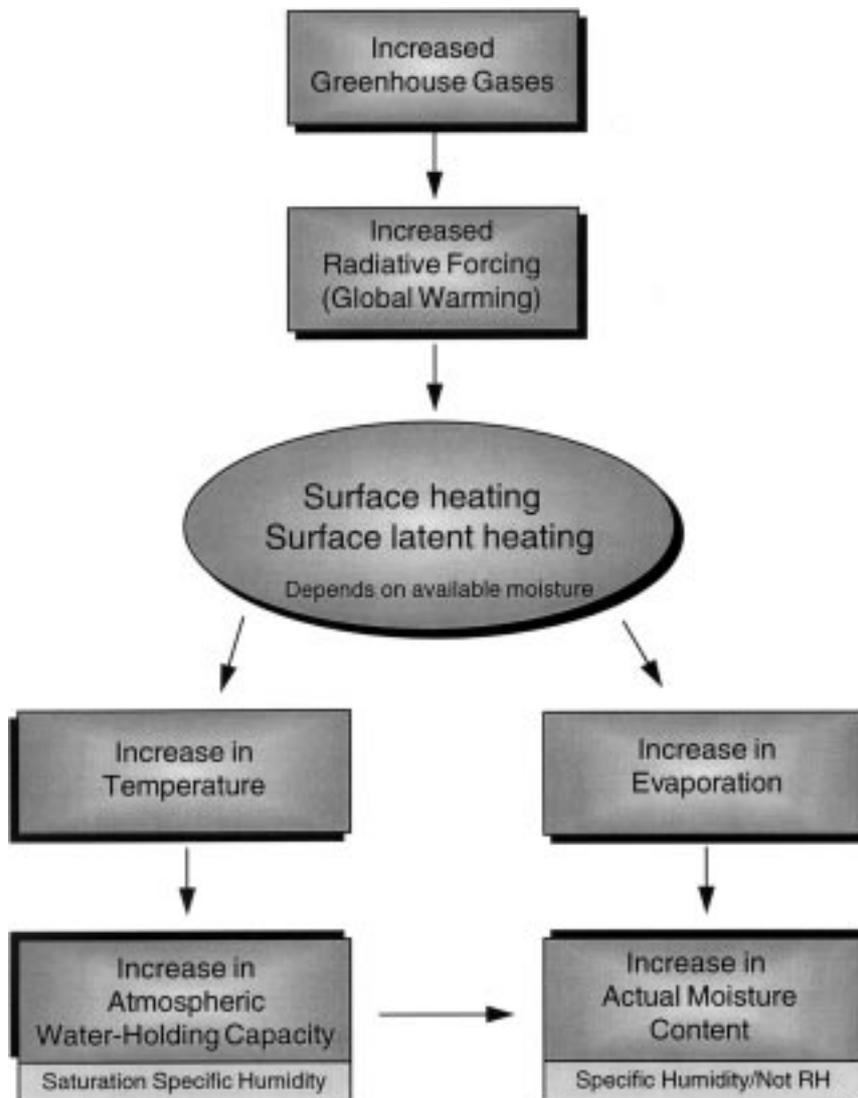


Figure 12a.

tation mean daily amounts increase. Overall, however, they find both increases and decreases of both precipitation frequency and intensity. Gregory et al. (1997) have explored various drought-related statistics including the distribution of daily rainfall amounts and frequency of long dry spells in a time dependent CO_2 experiment and find the possibility of increases in the severity of drought as CO_2 is increased.

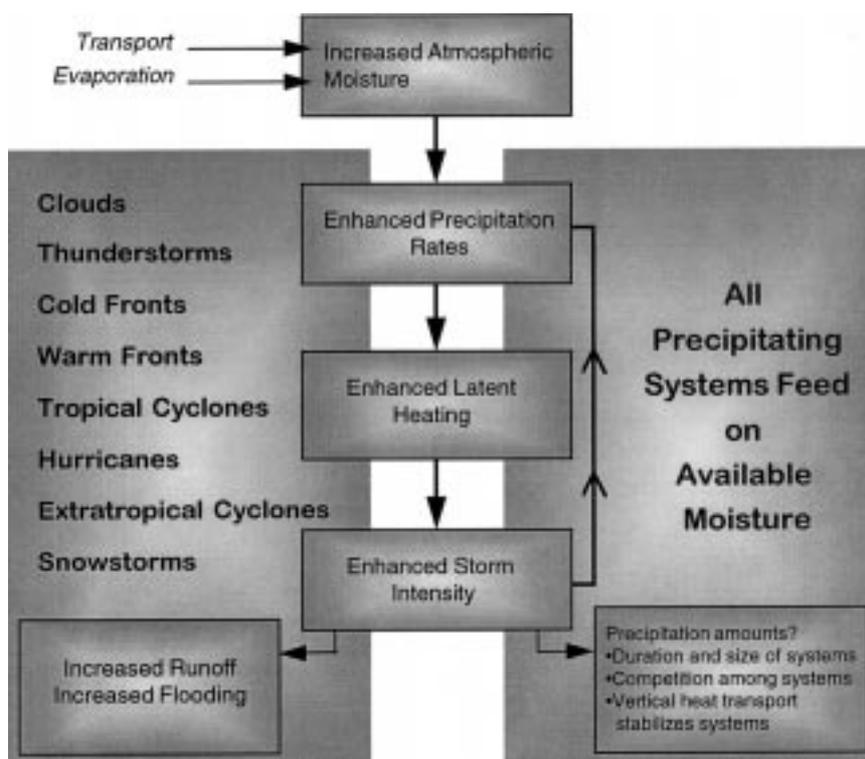


Figure 12b.

Figure 12. Schematic outline of the sequence of processes involved in climate change and how they alter moisture content of the atmosphere, evaporation, and precipitation rates. (a) Shows the factors involved in increasing moisture content of the atmosphere, while (b) shows how this alters rainfall rates as all precipitating systems feed upon the atmospheric moisture available within reach of the storm-scale circulation.

5. Conclusions and Recommendations

The arguments on how climate change can influence moisture content of the atmosphere, and its sources and sinks are assembled in the schematics in Figure 12. This provides the sequence described earlier. The sequence given is over-simplified as it does not indicate some of the feedbacks that can reinforce or interfere. For example, increased water vapor in the atmosphere provides a strong positive feedback as water vapor is a greenhouse gas. But in addition, an increase in atmospheric moisture may lead to increased relative humidity and increased clouds, which could cut down on solar radiation (enhance shortwave cloud forcing) and reduce the energy available at the surface for evaporation. Those feedbacks are included in the climate models and alter the magnitude of the surface heat available for evaporation in different models but not its sign. Figure 12 provides the rationale for why rainfall rates and frequencies as well as accumulations are important in understanding

what is going on in precipitation locally. All precipitating systems (e.g., as listed in Figure 12) feed upon the atmospheric moisture available within reach of the storm-scale circulation. The accumulations depend greatly on the frequency, size and duration of individual storms, as well as the rate (Byers, 1948) and these depend on static stability and other factors as well. In particular, the need to vertically transport heat absorbed at the surface is a factor in convection and baroclinic instability both of which act to stabilize the atmosphere. Increased greenhouse gases also stabilize the atmosphere. Those are additional considerations in interpreting model responses to increased greenhouse gas simulations.

There is firm evidence that moisture in the atmosphere is increasing in many places, as noted in the introduction. In particular, in the Western Hemisphere north of the equator, precipitable water amounts below 500 mb are increasing over the United States, Caribbean and Hawaii by about 5% per decade as a statistically significant trend from 1973 to 1993 (Ross and Elliott, 1996), and these correspond to significant increases in relative humidities of 2 to 3% per decade over the Southeast, Caribbean and subtropical Pacific. Precipitable water and relative humidities are not increasing significantly over much of Canada, however, and are decreasing slightly in some areas.

As noted earlier for the United States, much of the moisture for precipitation, especially in the winter half year, comes from moisture transported out of the subtropics in a southwesterly flow ahead of cold fronts. In both the West, where the moisture comes from the subtropical Pacific and the observed upward trends in moisture at Hawaii are pertinent, and east of the Rockies, where the moisture comes mainly from the Gulf of Mexico and subtropical Atlantic, the moisture amounts are strongly trending upwards. Because perhaps about 25% of the moisture comes locally from evaporation from the surface, the observed increased moisture amounts should be conducive to enhanced precipitation in individual storms and increase the risk of flooding.

Moreover, there are strong indications that rainfall rates have changed in the United States where the incidence of heavy rainfall events has steadily increased at the expense of moderate rainfall events throughout this century (Karl et al., 1996). In particular, this has been shown by an analysis of the percentage of the U.S. area with much above normal proportion of total annual precipitation from 1 day extreme events, where the latter are defined to be more than 2 inches (50.8 mm) amounts. The 'much above normal proportion' is defined to be the upper 10%. This quantity can be reliably calculated from 1910, and the percentage has increased steadily from less than 9 to over 11%, a 20% change. However, the analysis of linear trends in the frequency of precipitation events for the United States shows a predominance of small but mostly positive trends. Longer-term trends of U.S. precipitation amounts reveal statistically significant upward trends during the 20th Century of 8% (Karl et al., 1996; Karl and Knight, 1998). The latter study shows that heavy rainfall events are increasing at the expense of moderate events.

The above arguments suggest that there is not such a clear expectation on how total precipitation amounts should change, except as an overall average. With higher average temperatures in winter expected, more precipitation is likely to fall in the form of rain rather than snow, which will increase both soil moisture and run off, as noted by the IPCC (1996) and as is found in many models. In addition, earlier and/or faster snow melt in spring is likely to aggravate springtime flooding. In other places, as discussed in the introduction, dipole-like structures in changes in precipitation patterns should occur in places where storm tracks shift meridionally. Another illustration of possible changes in storm tracks maybe the changes seen in Figures 10 and 11 in winter. Beyond this, it is suggested that examining rainfall rates and how they change with climate change may be more important and fruitful in understanding what is happening in model predictions. Use of 'threshold' techniques and fitting of log-normal distributions (e.g., Short et al., 1993a,b; Shimizu et al., 1993) may be worthwhile ways of analyzing these data.

Precipitation rates are not well known. Of course overall average rates are known as well as the precipitation itself is known, but most of the time precipitation is not falling. Therefore, it is concluded that a great deal more attention should be paid to the rates (or intensity) of precipitation, both in observations and models, conditional on when it is falling, and the frequency of precipitation. It is further suggested that the focus should be on 1-hour average rates as a useful compromise that is reasonably compatible with the lifetime of the main precipitating systems in nature, but which goes beyond instantaneous values, and is feasible from models which typically have time steps of about half an hour. This would facilitate a detailed analysis of the diurnal cycle of precipitation both in models and in nature which should be very enlightening.

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Notes

¹ The National Center for Atmospheric Research is sponsored by the National Science Foundation.

² This statement is not scientifically correct, as shown later. It is the increased heating, not warmer temperatures, that lead to a more vigorous hydrological cycle.

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