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Preferred states in spatial soil moisture patterns: Local and nonlocal controls

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Abstract. In this paper we develop a conceptual and observational case in which soil water patterns in temperate regions of Australia switch between two preferred states. The wet state is dominated by lateral water movement through both surface and subsurface paths, with catchment terrain leading to organization of wet areas along drainage lines. We denote this as nonlocal control. The dry state is dominated by vertical fluxes, with soil properties and only local terrain (areas of high convergence) influencing spatial patterns. We denote this as local control. The switch is described in terms of the dominance of lateral over vertical water fluxes and vice versa. When evapotranspiration exceeds rainfall, the soil dries to the point where hydraulic conductivity is low and any rainfall that occurs essentially wets up the soil uniformly and is evapotranspired before any significant lateral redistribution takes place. As evapotranspiration decreases and/or rainfall increases, areas of high local convergence become wet, and runoff that is generated moves downslope, rapidly wetting up the drainage lines. In the wet to dry transitional period a rapid increase in potential evapotranspiration (and possibly a decrease in rainfall) causes drying of the soil and "shutting down" of lateral flow. Vertical fluxes dominate and the "dry" pattern is established. Three data sets from two catchments are presented to support the notion of preferred states in soil moisture, and the results of a modeling exercise on catchments from a range of climatic conditions illustrate that the conclusions from the field studies may apply to other areas. The implications for hydrological modeling are discussed in relation to methods for establishing antecedent moisture conditions for event models, for distribution models, and for spatially distributing bulk estimates of catchment soil moisture using indices.

1. Introduction

Near-surface soil moisture is a major control on hydrological processes at both the storm event scale and in the long term. It influences the partitioning of precipitation into infiltration and runoff and controls evapotranspiration by controlling water availability to plants and so also affects the partitioning of latent and sensible heat. In this way, soil moisture is a link between the surface energy and water balances. Improved understanding of the spatial distribution of soil moisture is useful for a range of applications in hydrology. For event-based hydrological modeling, correct definition of antecedent soil moisture conditions is critical to accurate simulation [*Stephenson and Freeze*, 1974], and the spatial distribution of long-term water balance requires knowledge about the parts of the landscape that are persistently wetter than average. Similarly, in solute transport and erosion modeling, spatial patterns of soil moisture are of key importance because of their impact on water flux patterns. At a much larger scale a better understanding of soil moisture and its spatial distribution may contribute to improved land components of climate models and global circulation models.

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Evidence that average soil water content changes seasonally comes from a number of sources. Everyone who waters a garden or grows a crop recognizes that in most climates, there are major differences in average soil water content throughout the year. Seasonal changes in precipitation and evapotranspiration tend to lead to periods where soils are persistently wetter or drier than average. Long-term field studies such as that undertaken at the 100 ha R5 catchment in Okalahoma [see *Loague*, 1992] illustrate the effect in more quantitative terms. Soil water content was measured at four sites and eight depths (down to 1.22 m) over a 4 year period and at 34 sites and eight depths over an additional 4 year period. Water contents were shown to vary seasonally from less than 10% to greater than 40% by volume over depths to 0.6 m, with the deeper sites rarely dropping below 20% [*Loague*, 1992, pp. 239–241, Figures 2–4].

Kalma et al. [1995] presented similar data over several years from neutron access tubes in the 26.1 km^2 catchment at Lockyersleigh in New South Wales, Australia. Their results closely resemble those of *Loague* [1992], showing clear seasonality to depths of over 1 m [*Kalma et al*., 1995, p. 213, Figure 6].

An important question for hydrology is "how does this temporal variation in soil moisture affect the spatial patterns of soil moisture?" Although there is a lack of field data on soil moisture patterns and their seasonal variation, there is a substantial body of literature (and field data) on source areas. The classic paper of *Dunne et al*. [1975], for example, presented patterns of the seasonal variation of saturation areas in the WC-2 and WC-4 Sleepers River watershed in Vermont, in the Randboro catchment in Quebec, and in a swale near Peterborough, Ontario. These results clearly demonstrated the seasonal shrinking of the saturation areas during spring when the catchments dried out. In late winter a significant part of each catchment was saturated, but only a narrow strip around the drainage lines was saturated by late spring. This observation has been repeated in a number of studies [e.g., *Anderson and Burt*, 1978a, b; *Anderson and Kneale*, 1980; *Petch*, 1988]. The focus of these studies on source areas is generally on runoff-producing (saturated) areas which occupy only a very small part of the landscape over most of the year. In this paper we take a broader look and examine how soil moisture patterns change throughout a catchment.

Despite the lack of field data, there are at least two basic approaches that have been used to represent spatial patterns in soil moisture. The first is to use distributed parameter hydrological models run in a continuous mode. These can, in principle, provide both spatial and temporal patterns of soil moisture throughout a landscape but require a great deal of information to apply accurately, are computationally intensive, and are subject to problems with the conceptualization of processes [*Beven*, 1989; *Grayson et al*., 1992; *Blöschl and Sivapalan*, 1995].

The second is to use terrain-based wetness index methods. These are based on simple assumptions about the dominant controls on spatial soil moisture and have gained widespread use, particularly since digital elevation models have become easily available. They were originally derived to estimate depth to a shallow water table and to predict saturated source areas, but they are also used to generally define spatial soil moisture conditions or saturation deficit. The best known are those of *Beven and Kirkby* [1979] and *O'Loughlin* [1981]. These wetness indices are also used in both distributed and distribution models which use their basic patterns to allocate saturation deficit across a landscape. In distribution models, areas of equivalent deficit are assumed to behave similarly and are modeled as such with the output from each similar class being summed to determine overall catchment response.

Although some of these models have separate components for the simulation of infiltration excess runoff, the use of a wetness index for distributing soil moisture imposes an inherent assumption about the controlling processes on soil moisture distribution, namely, that subsurface lateral flow dominates. In addition, there is a steady state assumption which implies that the saturation deficit (and hence soil moisture) at any point in a catchment is influenced by lateral flow from the entire upslope contributing area. This has been relaxed in the quasi-dynamic wetness index of *Barling et al*. [1994] in which the upslope area that influences soil moisture at a particular point is made a function of time of rainfall and soil properties. This enables a range of distributions to be computed for different periods of rainfall (i.e., catchment wetness). While the index of *Barling et al*. [1994] was better able to represent spatial soil moisture in some areas, it is still based on the assumption that only lateral subsurface flow controls the spatial patterns of soil moisture content. In some wet climates or during wet periods this assumption may be realistic, but at drier times, it is not.

There are a number of indices that do not use upslope contributing area. One example is surface curvature. The rationale of its use is that curvature is a measure of local convergence. Depressions (areas with high curvature) tend to be wetter than more planar areas where curvature is low [e.g., *Moore et al*., 1991]. This index is not based on the sort of theoretical development that underlies the more common wetness indices, and while it has performed reasonably well in some studies [e.g., *Zaslavsky and Sinai*, 1981], in others it was not well correlated with spatial soil moisture [e.g., *Moore et al*., 1988; *Anderson and Kneale*, 1982]. These problems have led us to reassess the observational data on soil water patterns and to consider, from a theoretical point of view, how we might expect soil moisture patterns to develop in temperate regions.

2. Hypothesis of Preferred States

We will argue that soil water patterns in temperate regions of Australia (and similar climates elsewhere) switch between two quite different preferred states, controlled by different processes. Furthermore, when switching occurs, it occurs over a relatively short period. In the following sections we first present the states in a descriptive form and propose conceptual mechanisms by which the switching may occur between states. These sections are based on simple physical reasoning without detailed empirical support. They are essentially the presentation of a hypothesis. We then present data and simple analyses to support the premise and, finally, discuss the implications for modeling and the applicability of the notion of preferred states to other geographical and climatic regions.

2.1. The States

In periods when precipitation continually exceeds evapotranspiration, spatial patterns of soil water are dominated by lateral water movement by both surface and subsurface pathways. Drainage lines and other areas of high topographic convergence are wetter than other parts of the catchment because of the concentration of shallow, lateral subsurface flow and the surface runoff that is generated from these wet areas. The topography upslope of a given point is the dominant control on the spatial patterns of soil moisture. We therefore denote this as nonlocal control.

In the periods when evapotranspiration continually exceeds precipitation the vertical fluxes of evapotranspiration and rainfall dominate, and moisture patterns reflect soil and vegetation differences, taking on a more random appearance. There is no significant lateral flow and therefore no "connection" between a point in the catchment and its upslope area. Only local terrain influences the dry pattern of soil moisture where areas of high local convergence may cause a temporary elevation of water content following rainfall. We denote this as local control.

2.2. Mechanisms for Switching Between States

The switching mechanisms are described in terms of the dominance of lateral over vertical water fluxes and vice versa, i.e., nonlocal over local controls. During the period when evapotranspiration generally exceeds rainfall the surface soil layers are relatively dry, and lateral hydraulic conductivity is low. Any rainfall that occurs during this period of high evaporative demand wets up the soil more or less uniformly and is evapotranspired before any significant lateral redistribution takes place. The lateral conductivity is low irrespective of whether matrix or macropore flow occurs. If there is matrix

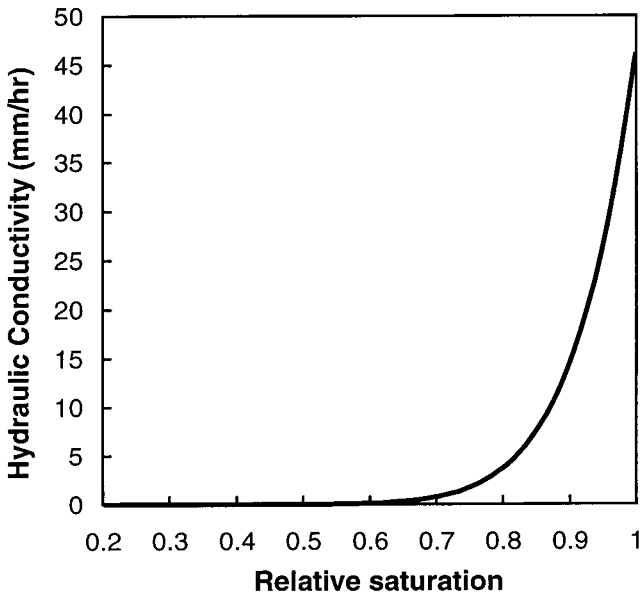

Figure 1. Hydraulic conductivity versus water content for a loam soil.

flow, the nonlinearity of hydraulic conductivity with water content ensures that a reduction in relative saturation, to say 70%, results in a decrease in conductivity of more than an order of magnitude. Figure 1 shows an example of a $K(\theta)$ relationship for a loam soil based on data used by *Grayson et al.* [1992]. Macropore flow can operate only when saturated conditions exist within walls of the macropore. Average soil water contents during dry periods are well below saturation, and the macropore flow mechanism cannot operate. This behavior of unsaturated soil means that even when rainfall occurs during the dry period, there will be negligible lateral flow unless the soil approaches saturation. Even if saturation is approached immediately after rain, high evapotranspiration demands rapidly dry the soil, and any lateral flow quickly ceases. This means that only local terrain can influence the soil water patterns during this dry period, and even then, the influence will be confined to periods immediately after rain. The major controls on soil moisture for these conditions are (local) soil properties, (local) vegetation, and possibly radiation if there are large aspect differences.

As evapotranspiration decreases and/or rainfall increases over the season, the areas of high local convergence are the first to become wet. These areas are generally in the upper parts of a catchment near the ends of depression lines. As these areas become wet, a progressively smaller amount of rain is needed to generate runoff. At some point, runoff is generated and moves down the depression, rapidly saturating the drainage lines from "above." The nonlinearity of hydraulic conductivity with soil water content and possibly the triggering of macropore flow provide a positive feedback to accelerate the dominance of lateral subsurface flow into these saturated areas, and the "wet" pattern is quickly established. This pattern is reinforced every time it rains. Soil water contents in the depression areas are at or near saturation, and elsewhere, they will also be high, so lateral flow will be significant, hence spatial patterns of wetness are influenced by more and more of the catchment terrain. The wet pattern is maintained throughout the period where precipitation exceeds evapotranspiration.

In the spring transitional period a rapid increase in evapotranspiration (and possibly a decline in rainfall) causes drying of the surface soil and the root zone. Macropore flow ceases as water contents drop below saturation, and the nonlinearity of hydraulic conductivity with soil water content "shuts down" lateral flow as the soil dries. Vertical fluxes dominate and the dry pattern is established.

There are two issues fundamental to the preceding argument. The first is that there are seasonal imbalances in climatic forcing that tend to create periods when average soil water contents are persistently low and when they are persistently high. The second is that the spatial distribution of soil moisture responds in a nonlinear way to these periods because of the relationship between soil water content and hydraulic conductivity and because there is an upper limit (saturation) and a lower limit (residual) to soil water content. Smooth cyclical changes in climatic forcing translate to rapid changes (i.e., switching) between the two preferred states and hence rapid changes in the control on the distribution of soil water from laterally dominated to vertically dominated.

There will be some environments that are totally dominated by one state or the other and do not switch between states. Desert and semiarid regions where potential evapotranspiration is always greater than precipitation will rarely (if ever) change from the "dry state." Similarly, very wet areas with low potential evapotranspiration will nearly always remain in the "wet state." The balance between precipitation and evapotranspiration will determine which of the preferred states is dominant, while this balance, and the capacity of the soil water store, will govern the rate of any switching behavior.

3. Evidence of Preferred States

In this section we describe data sets from two catchments to support the notion of preferred states. Brief descriptions of the catchments are followed by observational evidence of a switch between wet and dry states (local and nonlocal controls) of spatial soil moisture patterns following three lines of inquiry.

First, we present time series of soil moisture from a number of locations in the Wagga Wagga and Tarrawarra catchments. If soil moisture at all locations within a catchment acts in unison, we conclude that local controls dominate. If there is a delay in response between upslope and downslope sites, we infer nonlocal behavior. Second, we present measured spatial patterns of soil moisture from Tarrawarra. If high values are associated with drainage lines and depressions (i.e., zones with a large upslope contributing area), we infer nonlocal behavior. If values are random (or associated with local effects), we infer local controls. Third, we compute local water balances around neutron access tubes as well as examine piezometer data to determine the extent to which lateral flow (and hence nonlocal behavior) is occurring at different times of the year. Finally, to assess whether the wet/dry switching behavior observed in the three data sets is valid for other climatic regimes, we run a lumped conceptual catchment model on a number of sites around Australia.

3.1. Catchment Descriptions: Wagga Wagga and Tarrawarra Field Sites

The 7.5 ha Wagga Wagga catchment is located in southern New South Wales, Australia (Figures 2 and 3). It has an easterly aspect with an average slope of approximately 12% and a vertical relief of 45 m. Average annual rainfall is 675 mm

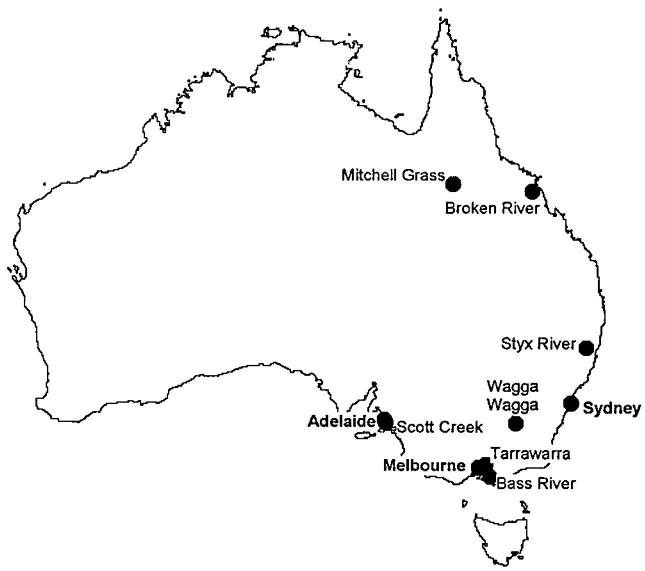

Figure 2. Location diagram for catchments referred to in the text.

ranging from 350 to 950 mm and is relatively evenly distributed through the year, although higher intensities tend to occur in the summer. Average annual class A pan evaporation is 1340 mm. The soils are duplex with a hard-setting sandy loam over clay B horizon. The depth of the A horizon varies from approximately 0.1 m in eroded areas to over 1 m in the lower depositional areas. Weekly soil samples have been collected from six sites over a 5 year period and used to measure gravimetric water content in the top 50 mm. The catchment and sample locations are shown in Figure 3.

The 10 ha Tarrawarra catchment is located in Victoria about 50 km east of Melbourne and 450 km south of the Wagga Wagga catchment (Figure 2). It has a range of aspect from east through south to north-northwest and typical slopes of 11–14%. The average annual rainfall is 1000 mm, and the class A pan evaporation is 1200 mm. The maximum mean monthly rainfall occurs in October (110 mm), and the minimum occurs in January (65 mm). The soil over most of the catchment is

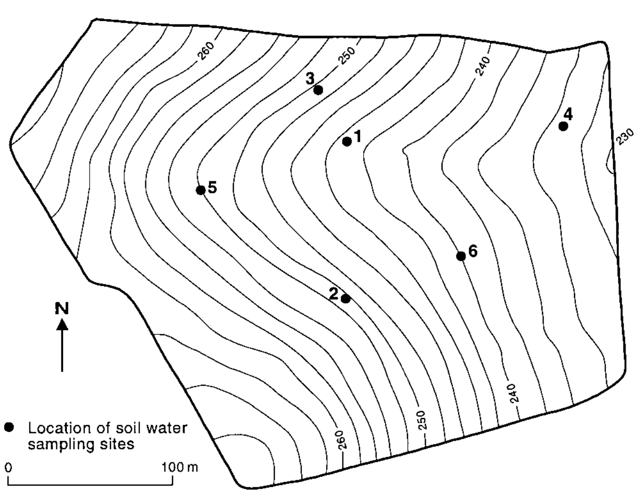

Figure 3. Wagga Wagga experimental catchment (area = 7 ha).

duplex with a loamy A horizon of 0.20–0.40 m overlying a clay B horizon. The depression areas are quite silty with the clay layer over 1 m below the surface.

A detailed investigation into the dynamics of soil moisture is being undertaken at the Tarrawarra catchment. The experiment is described by A. W. Western and R. B. Grayson (The Tarrawarra data set: Soil moisture patterns, soil characteristics, and hydrological flux measurements, submitted to *Water Resources Research*, 1997, hereinafter referred to as Western and Grayson, submitted manuscript, 1997) and involves the spatial measurement of soil moisture in the top 300 mm on a 10 by 20 m grid using time domain reflectrometry (520 sample points) as well as water content profiles using a Neutron Moisture Meter (20 sites) and piezometric levels (74 sites). Figure 4 is a map of the catchment showing the topography and the sites for fixed measurements.

3.2. Temporal Changes in Soil Water Content

Gravimetric soil water contents from the Wagga Wagga site are presented as a time series over 5 years in Figures 5a–5c. It is clear that the response over the summer and winter periods is quite different. In the summer all six sites wet up and dry down in unison. There is no delay in response between upslope and downslope sites and no sites which are consistently wetter or drier than others. The sites are affected by only local controls and are simply wetting up because of incident rainfall and drying down because of evaporation; that is, vertical fluxes are dominant. There is no evidence of lateral flow (or nonlocal controls).

Over the winter all the sites are persistently wetter than in the summer. There is evidence of general topographic influence (nonlocal control) with sites 1 and 5 in the convergent areas generally wetter than other sites on planar or divergent areas (e.g., site 2). This illustrates that convergence of subsurface and possibly surface flow is occurring, keeping those sites wetter. There is also some evidence of delays to drying during the winter period, particularly for site 1. Again, this indicates that lateral flow processes are occurring.

Given the long time series of these data, it is possible to gain some appreciation of the average time for transition from dry behavior, when all sites are changing in unison, to wet behavior. It appears to occur over periods of around a month, with the "wetting" being more rapid than the "drying." It is therefore apparent from Figure 5 that the transitional periods constitute less than 15% of the year.

There is insufficient information in these data to assess the mechanisms for switching between states; however, the work presented by *Barling et al.* [1994] from an adjacent catchment at the Wagga Wagga field station provides some insight into the dry-wet switch. The catchment referred to by *Barling et al.* [1994] is of similar size to that shown in Figure 3 and has similar topography. They showed that the areas to first become wet after the summer dry period were at the points of maximum local topographic convergence, at the heads of two depression lines [*Barling et al.*, 1994, p. 1035, Figure 6, p. 1042, Figure 14]. The depression lines themselves remained dry until, finally, runoff was produced from upper areas which flowed over the initially unsaturated depressions to the catchment outlet. Piezometric response down the depression line indicated that a shallow groundwater mound formed under the depressions to become, finally, saturated as runoff continued. This supports the switching mechanism proposed in section 2.2. Note that local terrain features determine the sites at which the soil first saturates by concentrating local subsurface

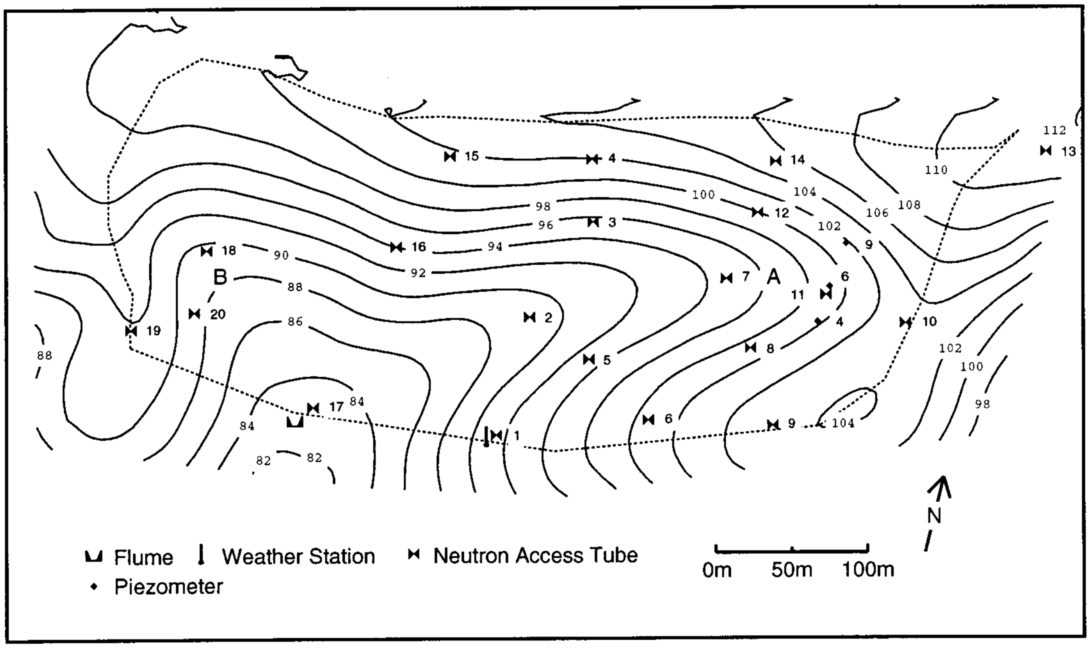

Figure 4. Tarrawarra experimental catchment (area = 10.5 ha).

flow. These are not the sites that would be first to saturate if steady state flow assumptions were applied. Once surface runoff occurs, there is rapid lateral redistribution, and nonlocal influences begin to dominate.

Figure 6 shows the scaled water content in the top 600 mm from a selection of the neutron access tubes on the Tarrawarra catchment over 1 year. The data have much better spatial coverage than those for the Wagga Wagga site, and they provide an estimate of water content in all of the hydrologically active zone rather than just at the surface. The data have been adjusted to give a scaled wetness W:

$$W = \frac{\theta - \theta_{\mathrm{driest}}}{\theta_{\mathrm{wettest}} - \theta_{\mathrm{driest}}}$$

where θ is the average water content in the top 600 mm, $\theta_{\mathrm{wettest}}$ is the wettest water content in the top 600 mm, and θ_{driest} is the driest water content in the top 600 mm.

The data show a clear distinction between the response during the summer (December–March), when all sites wet up and dry down in unison, and the response during the winter (May–September), when sites are either all wet or their relative wetness is heavily controlled by their topographic position. For example, sites 2, 7, and 18 in the gully areas are virtually always saturated, whereas in the spring of 1995, sites 8, 15, and 20 in the planar or divergent areas were drier. As with the Wagga Wagga data, there is a sharp transition from the dry to the wet behavior, i.e., from the time when vertical fluxes dominate over lateral fluxes or when local controls dominate over nonlocal controls. The actual range of water contents over which the switch from lateral to vertically dominated behavior occurs will be dependent on the soil characteristics and particularly the relationship between hydraulic conductivity and water content. Given the sort of behavior illustrated in Figure 1, switching would probably occur in the range of relative saturation between 0.6 and 0.8.

3.3. Spatial Soil Moisture Measurements

Patterns of spatial soil moisture in the top 300 mm have been measured at Tarrawarra throughout the year. They clearly show the spatial characteristics of the wet (Figure 7) and dry (Figure 8) states. The pattern in Figure 7 shows the dominance of nonlocal controls with the drainage lines being significantly wetter than the planar and divergent areas. This pattern is reinforced by both surface and subsurface flow. The pattern is relatively well described by the *Beven and Kirkby* [1979] wetness index ($r^2 \approx 0.5$), which assumes nonlocal control and steady state, meaning that the whole upslope area of a given point contributes to wetness at the point.

The dry pattern shown in Figure 8 is much more random in appearance. There is no evidence of redistribution caused by terrain; that is, there are no nonlocal controls evident. The range of moisture values can be explained by soil properties, vegetation, radiative exposure (aspect), and uncertainty in measurements.

In 1996 it was possible to track the switch from dry to wet states in some detail. Foci of high wetness began to form in the areas marked A and B on Figure 4 during the month of April (autumn). In the patterns measured on April 13, water contents in these areas were between 39% and 45%, whereas those in the main depressions were similar to the rest of the catchment (between 30% and 39%). Saturated water content is approximately 47%. On April 14, the first surface runoff since the spring of 1995 was produced as 18 mm of rain fell, saturating areas A and B. This runoff flowed down the depression lines but did not make it to the catchment outlet, infiltrating within the depressions. This surface flow brought the water content of the depressions to near saturation, and during a 2.6 mm event the next day, runoff was measured at the flume. Wetness measurements then followed the same basic pattern as Figure 7 until October when the soil began to dry.

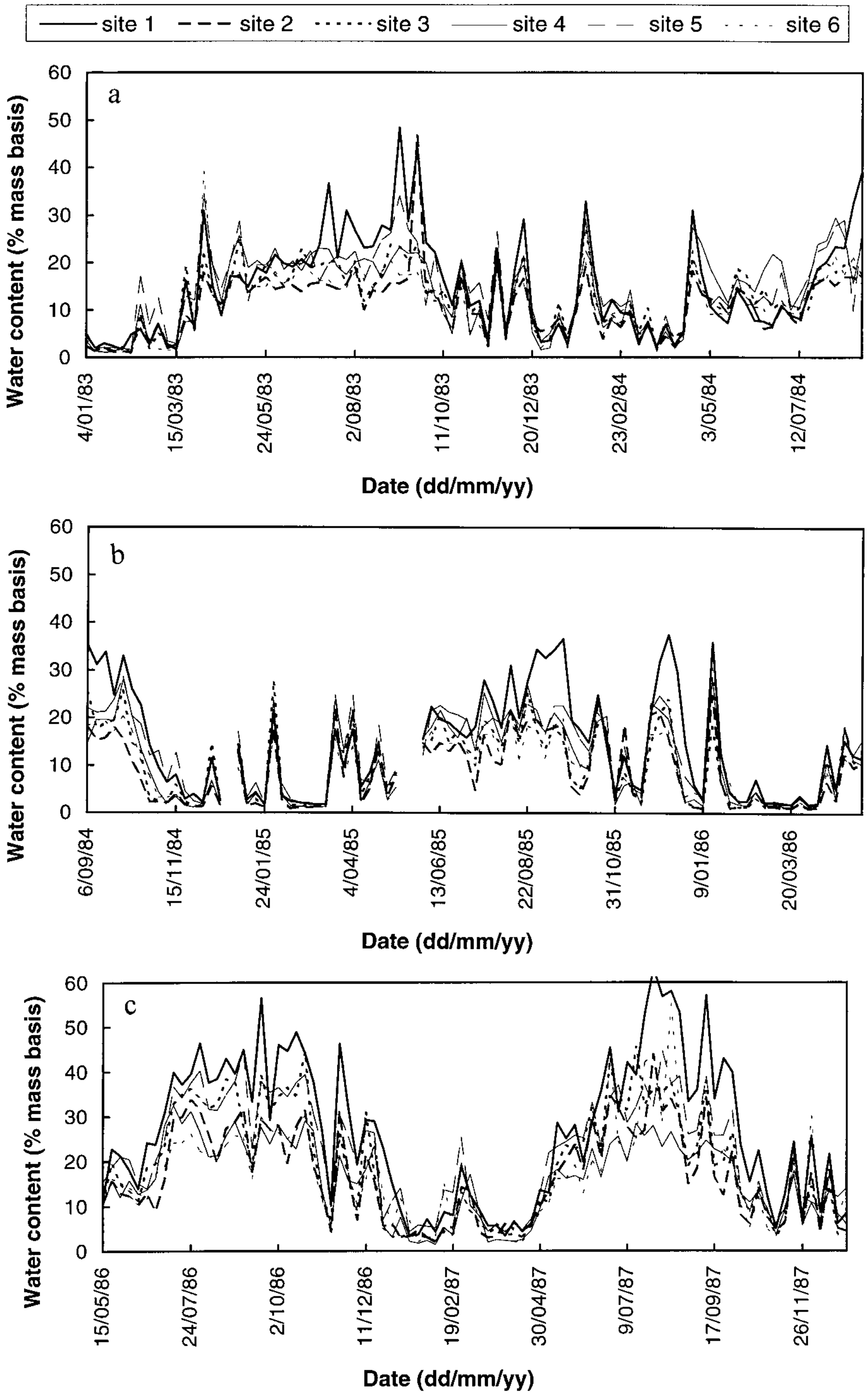

Figure 5. Time series of gravimetric soil water content in the top 5 cm from five sites on the Wagga Wagga catchment: (a) January 1983 to August 1984, (b) September 1984 to April 1986, and (c) May 1986 to December 1987.

3.4. Direct Evidence of Vertical and Lateral Fluxes

Water balance calculations for the top 600 mm around 19 neutron moisture meter access tubes from Tarrawarra were made for three events during the summer. This was used to determine whether the change in water content around an access tube could be explained by the vertical flux of rainfall (i.e., if change in water content could equal rainfall) or whether additional lateral flow was required to cause the measured water content change (i.e., if the change in water content is greater than rainfall). Evapotranspiration was ignored in the analysis because the water content measurements were generally made within 24 hours of the cessation of rainfall, so evapotranspiration would be less than a few millimeters. Rainfall for the three events was 33, 58, and 38 mm.

Table 1 shows the ratio of water content change to rainfall. Given the possible errors in the data, it may be considered that any deviation from 1.0 of greater than 20% is significant. For these summer events, changes in water content can be ex-

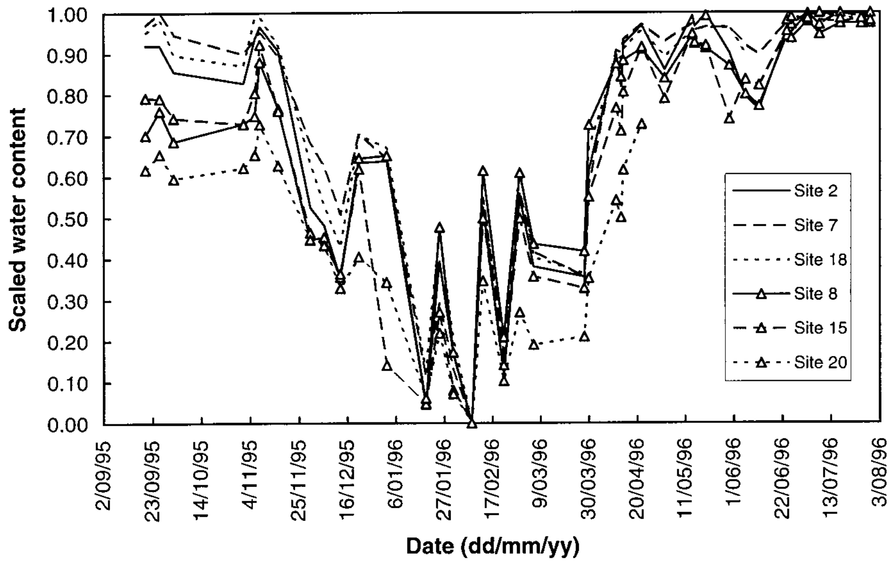

Figure 6. Scaled water content in the top 600 mm from six sites on the Tarrawarra catchment.

plained by infiltration of the rainfall for all but two sites (11 and 18). These sites became wetter than would have been expected if only vertical processes were occurring. It is also these sites where the switching process to wet conditions is initiated (A and B in Figure 4). Nevertheless, it is clear that lateral flow is occurring in very few places and is limited in areal extent. Note that the gully sites (2, 7, and 17) show no evidence of lateral redistribution.

This type of analysis is not possible for the wet periods because saturation and surface runoff occur. There are, however, 74 piezometers on the catchment, and these show rapid lateral redistribution of water following rainfall throughout the wet period. Figure 9 shows the levels in three tubes along one transect near point A in Figure 4. Tube 6 in the base of the depression is "fed" by lateral flow that keeps the water table almost at the surface, while tubes 4 and 9 drain more quickly. Other piezometers confirm this behavior. The other directly observable lateral redistribution mechanism is, of course, surface runoff, which reinforces the nonlocal pattern.

4. Other Climatic Regimes

To assess whether the switching behavior might occur in other parts of Australia, simulations were performed using a well-calibrated, conceptual hydrological model on catchments in different climatic regions. The sites are shown in Figure 2

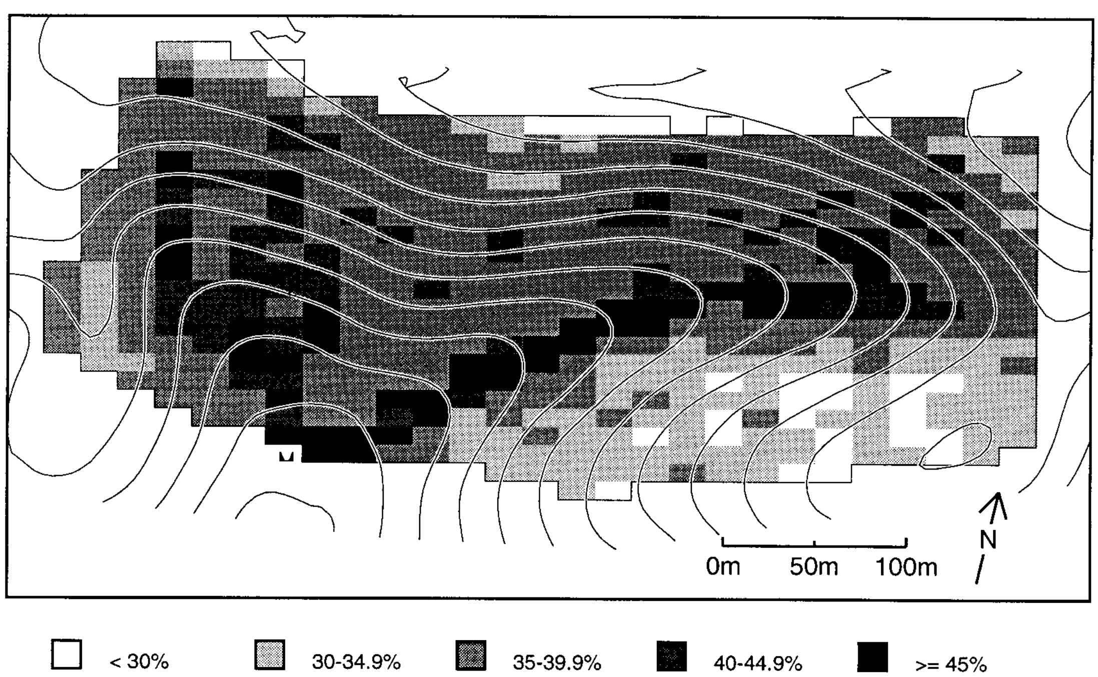

Figure 7. Example of the wet state soil water distribution measured at the Tarrawarra catchment (each cell represents one measurement). Soil water content in % vol./vol. Flume is marked at catchment outlet.

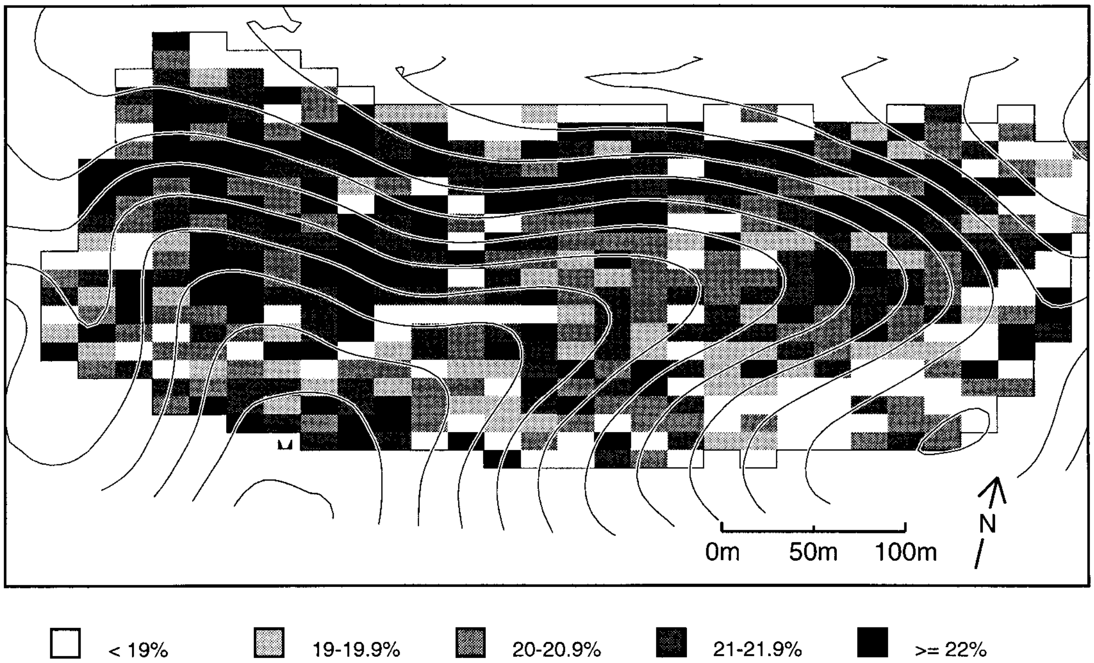

Figure 8. Example of the dry state soil water distribution measured at the Tarrawarra catchment (each cell represents one measurement). Soil water content in % vol./vol. Flume is marked at catchment outlet.

and cover both summer- and winter-dominated rainfall regimes. As is typical of Australian catchments, the "hydrologically active" soil depth is relatively shallow, and all sites have duplex soils where a heavy clay soil underlies a lighter textured topsoil.

The simulations are from the daily rainfall-runoff model MODHYDROLOG, which simulates runoff from rainfall and potential evapotranspiration data. The model attempts to simulate the hydrological processes using conceptual stores and empirical equations to mimic actual processes. Figure 10 is a schematic representation of the model structure. A more detailed description is given by *Chiew and McMahon* [1994].

The relative soil wetness is defined as the level of the soil moisture store divided by the soil moisture store capacity (SMSC). MODHYDROLOG is used in this application as a lumped model with a single set of parameter values. The optimized values of SMSC range between 120 and 180 mm. Because the model is run in a lumped configuration, the soil water store represents the entire catchment. It is therefore not possible to see any spatial behavior; however, it will be possible to determine if the temporal pattern of total catchment wetness shows persistent low and high periods. Low average catchment wetness would correspond to the dry state, where lateral redistribution is minimal and local controls dominate spatial patterns. High average catchment wetness would correspond to the wet state, where lateral redistribution is significant and nonlocal controls dominate spatial patterns.

The total 5 day rainfall and potential evapotranspiration

Table 1. Average Ratios of Water Content Change to Rainfall

Site	ΔWater Content/Rainfall
1	1.1
2	1.0
4	0.8
5	1.1
6	1.0
7	0.9
8	1.2
9	0.9
10	0.8
11	1.3
12	0.9
13	1.1
14	0.8
15	1.0
16	0.9
17	1.1
18	1.3
19	0.8
20	0.9

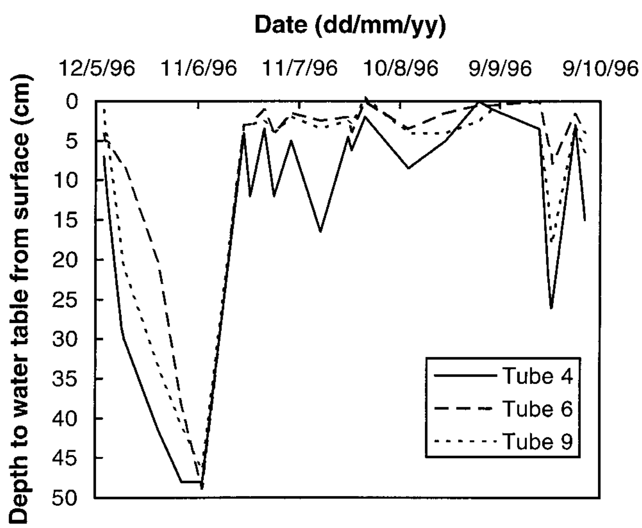

Figure 9. Piezometer response for sites 4, 6, and 9 at the Tarrawarra catchment (see Figure 4 for locations).

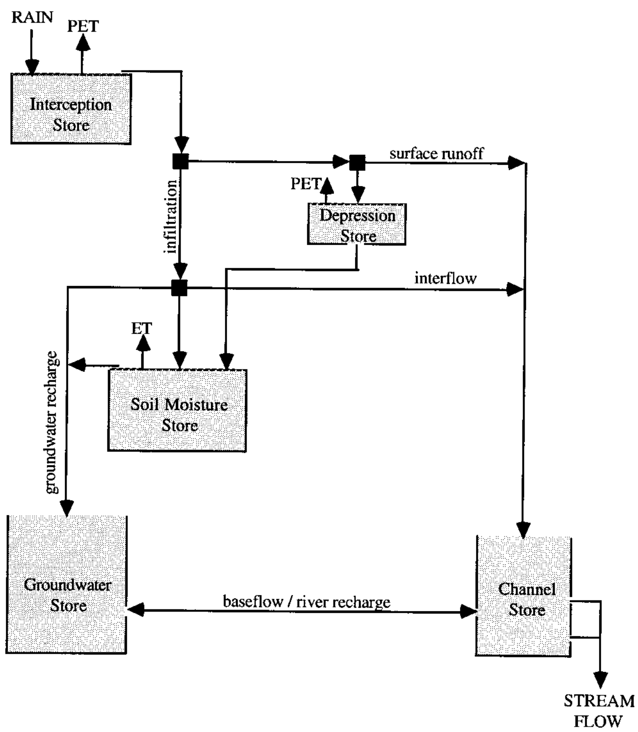

Figure 10. Conceptual structure of the MODHYDROLOG rainfall-runoff model.

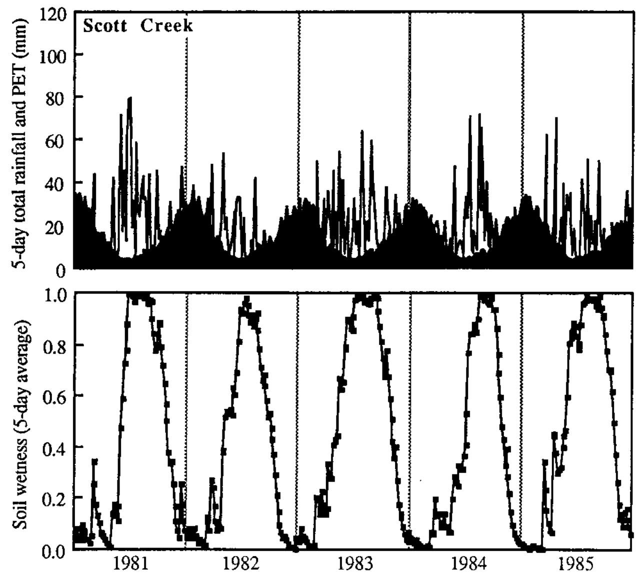

Figure 11. Rainfall, potential evapotranspiration (PET), and simulated soil wetness response of the Scott Creek catchment (27 km^2). See text for definition of soil wetness.

data and the average 5 day soil wetness estimated by MODHYDROLOG for five catchments (see Figure 2) are shown in Figures 11–15. Average soil wetness is defined as the level of the soil moisture store divided by the soil moisture store capacity (see Figure 10). The model is calibrated and validated against runoff data, and in the five catchments here the simulated and recorded runoff for independent validation periods is practically the same (correlation coefficients are greater than 90%; *Chiew and McMahon*, 1994). Eight to 14 years of data are used for the simulations, although Figures 11–15 present results from only the final five years.

It should be noted that MODHYDROLOG is calibrated against streamflow data only, and although it is conceptually based, the reliability of model simulations of soil wetness and other internal fluxes is unknown. The absolute values of these fluxes should therefore not be used directly. However, because the model mimics the actual processes and because runoff is fitted well, it is reasonable to study the relative changes in the soil moisture estimates. Further simulations with other conceptual models, as well as simulations using fewer parameters in MODHYDROLOG, show bulk soil wetness behaviors which are similar to those given in Figures 11–15 [see *Chiew and McMahon*, 1993; *Chiew et al.*, 1993].

Periods of persistent wet and dry soil moisture conditions are clearly evident in Figure 11, where the catchment has a winter-dominated rainfall (Scott Creek at Scotts Bottom, 27 km^2). Persistently wet soil conditions occur over winter (April–October) because there is more than enough rainfall to meet the evapotranspirative demands. In summer the soil water content is persistently dry, as rainfall is usually less or not much higher than potential evapotranspiration (PET). It usually takes less than 1 month to change from the dry to the wet conditions and from the wet to the dry conditions (see Figure 11).

Note that the actual switch from a wet state to a dry state will not require the catchment to be fully wet or fully dry because any lateral flow will reduce to negligible levels at water contents of say 70% of saturated values (e.g., Figure 1). We propose that the actual transition between lateral flow-dominated spatial patterns and vertical flow-dominated spatial patterns is likely to be in the range of relative wetness from 0.6 to 0.8.

The persistence of wet and dry conditions is also evident in the region with a more uniformly distributed rainfall (Bass River at Loch, 52 km^2; Figure 12). Like the winter rainfall region (Scott Creek), the change from persistently dry to per-

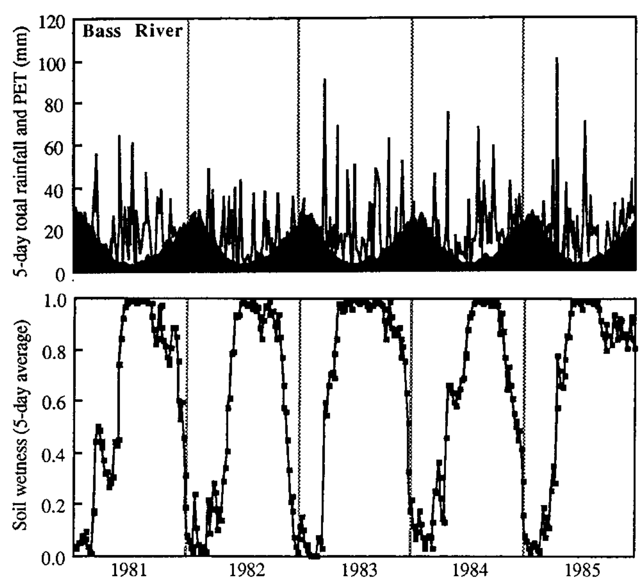

Figure 12. Rainfall, PET, and simulated soil wetness response of the Bass River catchment (52 km^2). See text for definition of soil wetness.

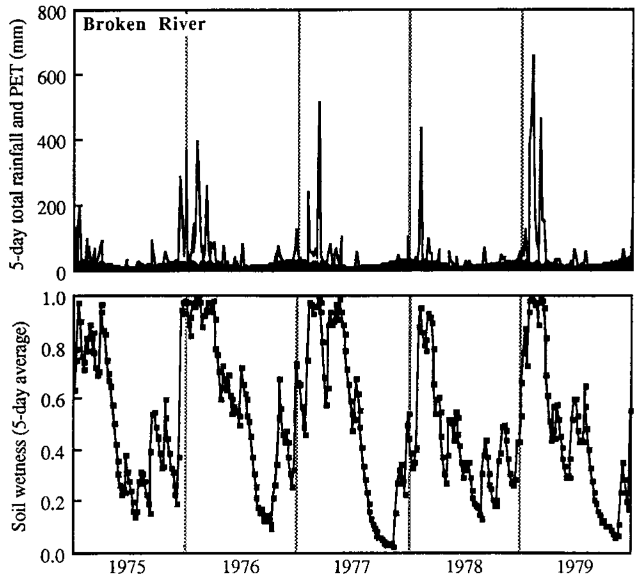

Figure 13. Rainfall, PET, and simulated soil wetness response of the Broken River catchment (41 km^2). See text for definition of soil wetness.

sistently wet usually occurs over several weeks, while the change from the wet to dry can sometimes take up to 2 months (Figure 12). However, as previously discussed, switching does not require full wet or dry conditions, so actual transition periods would be less than one month.

The simulations for a catchment with a summer-dominated rainfall (Broken River at Crediton, 41 km^2) demonstrates that persistent wet and dry conditions are also evident in this climate region (Figure 13). Practically all the rain here falls in the summer (75% of the annual rainfall occurs in the 4 months from December to March). The catchment is wet during this period because the rainfall is very much higher than PET. In the winter the catchment is dry because although PET is low, the winter rainfall is less or not much higher than PET. The change from persistently wet to persistently dry occurs over a longer period compared to the winter-dominated rainfall areas. This is because switching from a wet condition in the summer to a dry condition in the winter requires a long period of little rain because of the low PET.

The simulations for the 163 km^2 Styx River at Jeogla are presented in Figure 14 to show that although preferred states of average catchment soil moisture exist in a range of regions, the change between the two states is not necessarily well defined everywhere. This example shows that changing from persistently wet to persistently dry occurs more than once a year with the wet to dry change being somewhat slower than the dry to wet. The summer PET in this catchment is not much higher than the winter PET (compared to the clear summer-winter distinction in Scott Creek and Bass River), and except for 2 or 3 dry months in the winter, rainfall is fairly uniform through the year. This example demonstrates the difficulty in defining a consistently dry or wet condition in uniform rainfall areas without a clear summer-winter PET difference.

The simulations for the arid region (Mitchell Grass, 3 km^2) show an example of an area that is always in a dry state except for short periods every few years (Figure 15). High rainfall regions which are permanently in a wet state will also exist, but an example of such an area is not shown here.

Although these simulations do not directly verify the spatial behavior presented in section 3, the temporal behavior of average catchment wetness follows the same pattern as that proposed in the hypothesis and supported by the data presented in section 3. It is therefore reasonable to expect that the temporal behavior will be manifested in spatial behavior similar to that described in section 3.3. The persistently dry periods will have spatial patterns that are locally controlled, being dominated by vertical fluxes, and the persistently wet periods will have spatial patterns that are nonlocally controlled, being dominated by lateral fluxes of surface and subsurface flows.

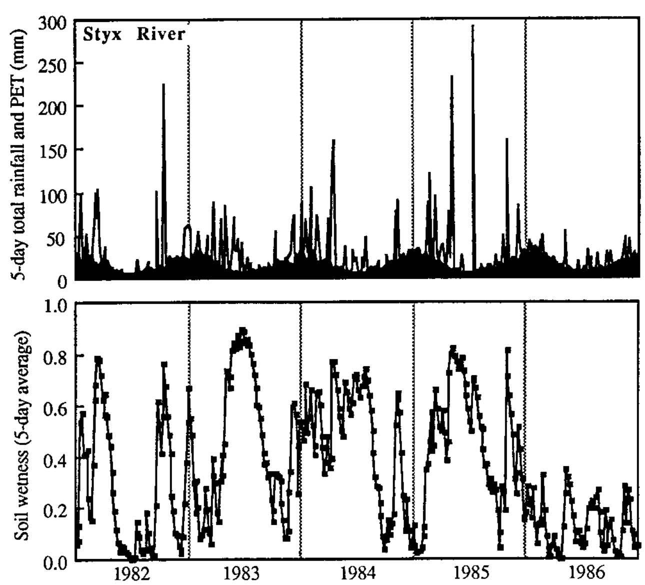

Figure 14. Rainfall, PET, and simulated soil wetness response of the Styx River catchment (163 km^2). See text for definition of soil wetness.

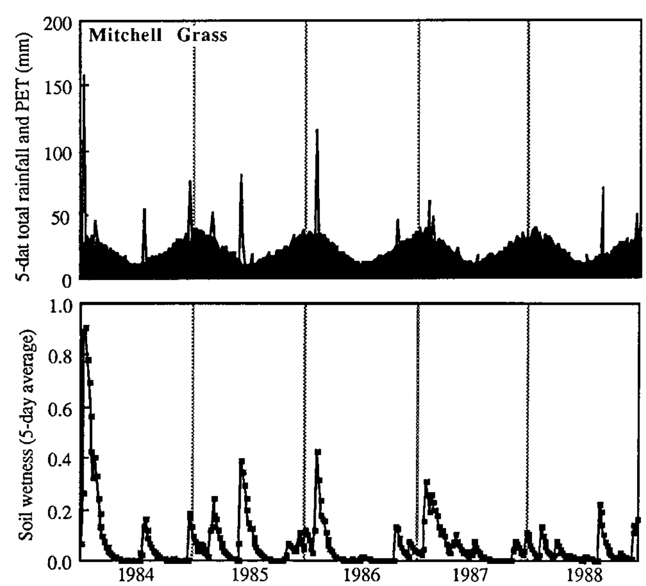

Figure 15. Rainfall, PET, and simulated soil wetness response of the Mitchell Grass catchment (3 km^2). See text for definition of soil wetness.

5. Implications and Applicability

The preceding data and analyses have illustrated that there is evidence for the presence of preferred states in soil moisture over a range of climatic regions. Temporal switching from persistently wet to persistently dry soil water conditions results in spatial patterns of soil moisture that are controlled by different processes.

The temporal switching is apparent in other published data [e.g., *Loague*, 1992; *Kalma et al*., 1995], but the influence on spatial patterns seem less widely recognized. Indeed much of the modeling of spatial soil moisture, such as that based on wetness indices, assumes that there are always nonlocal controls. Most of these approaches also assume that the nonlocal control is in the form of lateral subsurface flow from the entire upslope area of a particular point in the catchment. It is therefore clear that these approaches cannot be expected to represent spatial patterns accurately during the dry state when local controls and vertical fluxes dominate.

At first glance the ideas presented in this paper may suggest that the only way to represent soil moisture patterns throughout a year is to simulate them using a fully dynamic distributed model. This would imply that a great deal of catchment specific information is required and that the elegance of simple index-based distribution models must be sacrificed.

While the presence of preferred states means that a single wetness index cannot represent reality in these catchments, if the switching between states is rapid, it might be possible to use two separate indices, reflecting the appropriate dominant processes, along with a "switching criterion" to determine which state is appropriate. This approach would retain the basic elegance (and robustness) of the index-based modeling approaches for establishing antecedent conditions, for distributing spatially estimates of bulk catchment water content, and for use in distribution models (e.g., TOPMODEL).

In such an approach the wet index will be based on nonlocal controls such as surface flow and lateral subsurface flow (either matrix or through macropores). Existing steady state methods display some empirical correlation with observed patterns of soil moisture in wet conditions, although it should be noted that the mechanisms for formation of the wet areas are not always those underlying present indices (see description above of the dry to wet switch). Present indices may be improved by projecting areas of high local convergence downslope and/or by including aspect and other features (see A. W. Western et al., Observed spatial organization of soil moisture and relation to terrain, submitted to *Water Resources Research*, 1997).

The dry index will need to combine local terrain and soil characteristics to capture the key features of the soil water patterns during times when vertical fluxes dominate, i.e., when there is local control. This assumes that rainfall is uniformly distributed in space, which for small temperate catchments is often a reasonable assumption [e.g., *Adijizian and Ambroise*, 1996; Western and Grayson, submitted manuscript, 1997]. In areas where convective storm activity dominates the spatial distribution of rainfall may determine the spatial variability of short-term moisture content [*Goodrich et al*., 1994].

The switching criterion will be some measure of the imbalance between evapotranspiration on one hand and rainfall and soil water storage on the other. For areas with a strong seasonal rainfall pattern the simulations in section 4 suggest that the switch can be derived from a simple antecedent precipitation index. Where rainfall is more uniform, the criterion will need to be more sophisticated, incorporating perhaps potential evapotranspiration (or time of year as a surrogate) and rainfall, modulated by total soil water storage. More research is needed on this criterion.

On the basis of the limited set of catchments examined in this study it may be concluded that there are four basic modes of response in soil moisture: (1) always dry, (2) always wet, (3) both wet and dry with rapid transition, and (4) both wet and dry with slow transition. These modes depend on the climate and soil storage. From the point of view of modeling spatial soil moisture, only the last of these modes poses a problem for the use of index-based approaches. However, this work indicates that this mode does not occur very often. Therefore index-based approaches, either as a single or compound (two-state) index, may be useful for the majority of catchments provided a suitable switching criterion can be determined. This will be the subject of further study.

6. Conclusion

We have illustrated that in temperate regions of Australia where hydraulic conductivity decreases with depth (such as in duplex soils), spatial patterns of soil moisture display two preferred states and spend a relatively short period in transition. The wet state tends to occur in winter and is characterized by nonlocal moisture patterns. The dry state tends to occur in summer and is characterized by local soil moisture patterns. This behavior also seems to be likely in other climatic areas so long as there are extended periods where there is an imbalance between precipitation and evapotranspiration.

This behavior suggests that a realistic representation of spatial patterns of soil moisture in these regions cannot be achieved through the use of a single wetness index. If an index-based approach is to be maintained, it will be necessary to develop two indices based on different dominant processes to represent the wet and dry states as well as a switching criterion to determine the appropriate state. It is suggested that the presence of preferred states simplifies the accurate representation of spatial patterns of soil moisture using index methods because the transitional periods are short and may be ignored.

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