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A comparative modeling study of soil water dynamics in a desert ecosystem

Paul R. Kemp, James F. Reynolds, Yakov Pachepsky,[1] and Jia-Lin Chen
Department of Botany and Phytotron, Duke University, Durham, North Carolina

Abstract. We compared three different soil water models to evaluate the extent to which variation in plant growth form and cover and soil texture along a topographic gradient interact to affect relative rates of evaporation and transpiration under semiarid conditions. The models all incorporated one-dimensional distribution of water in the soil and had separate functions for loss of water through transpiration and soil evaporation but differed in the degree of mechanism and emphasis. PALS-SW (Patch Arid Lands Simulator-Soil Water) is a mechanistic model that includes soil water fluxes and emphasizes the physiological control of water loss by different plant life forms along the gradient. 2DSOIL is a mechanistic model that emphasizes the physical aspects of soil water fluxes. SWB (Soil Water Budget) is a simple water budget model that has no soil water redistribution and includes simplified schemes for soil evaporation and transpiration by different life forms. The model predictions were compared to observed soil water distributions at five positions along the gradient. All models predicted soil water distributions reasonably well and, for the most part, predicted similar trends along the transect in the fractions of water lost as soil evaporation versus transpiration. Transpiration was lowest (about 40% of total evapotranspiration (ET)) for the creosote bush community, which had the lowest plant cover (30% peak cover). The fraction of ET as transpiration increased with increasing plant cover, with 2DSOIL predicting the highest transpiration (60% of total ET) for the mixed vegetation community (60% peak cover) on relatively fine textured soil and PALS-SW predicting highest transpiration (69% of total ET) for the mixed vegetation community (70% peak cover) on relatively coarse textured soil. The community type had an effect on the amount of water lost as transpiration primarily via depth and distribution of roots. In this respect, PALS-SW predicted greatest differences among stations as related to differences in plant community types. However, since PALS-SW did not provide as good of fit with the soil moisture data as did 2DSOIL, the differences in the morphology and physiology of the life-forms may be secondary to the overall control of water loss by the primary factors accounted for in 2DSOIL: vertical distribution of soil moisture, degree of canopy cover, and evaporative energy budget of the canopy. Soil texture interacted with the amount and type of plant cover to affect evaporation and transpiration, but the effect was relatively minor.

Introduction

In arid ecosystems of the southwestern United States, average soil moisture is quite low and the patterns of seasonal precipitation and the quantity of soil water availability are highly variable [*MacMahon and Schimpf*, 1981]. Because of this, water is a major determinant of many ecosystem processes in desert ecosystems, including seed germination [e.g., *Kemp*, 1983], primary productivity [*Noy-Meir*, 1973], nutrient cycling [*Charley*, 1972], decomposition [e.g., *MacKay et al.*, 1987], and the distribution of plants and animals [e.g., *Cepeda and Whitford*, 1989; *Cornelius et al.*, 1991]. In turn, vegetation cover and biomass affect various hydrologic phenomena, including infiltration, runoff, interception, and erosion [*Spaeth et al.*, 1996]. Changes in distribution of soil moisture in arid regions could come about either directly through changes in the rainfall patterns (e.g., climatic fluctuations, El Niño patterns, drought cycles) or through numerous indirect changes (largely anthropogenic induced) [*Emanuel et al.*, 1985; *Balling*, 1991]. Some of these indirect changes include climate forcing via greenhouse gases, albedo changes via increased dust or vegetation loss, changes in vegetation composition, and changes in the soil surface that impact infiltration and runoff. If we are to accurately predict changing ecosystem structure and function in deserts in response to these external forces associated with global change as well as to local changes and feedbacks, we must accurately predict the distribution of soil water.

Prediction of soil water distribution in desert systems requires linking three fundamental units: soil, plant, and atmosphere. We suggest that an "ideal" model of soil water dynamics should respond to seasonal and yearly variation in weather, changes in kind and cover of vegetation, and perturbations that change soil texture or structure. While there have been numerous soil moisture models developed over the last three decades, few have dealt with arid lands having natural vegetation and extremely variable moisture inputs. Several investigators

[1]Also at Remote Sensing and Modeling Laboratory, Beltsville, Maryland.

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have modeled soil moisture in arid systems by adapting existing agricultural models [e.g., *Noble and Crisp*, 1979/1980; *Hanks*, 1981; *Lane et al.*, 1984; *Marion et al.*, 1985; *Moorhead et al.*, 1989]. Recently, *Paruelo and Sala* [1995] developed a model to examine the long-term dynamics of soil water in a Patagonian steppe, and *Walker and Langridge* [1996] proposed a simple model for semiarid savanna regions with limited data. These diverse models represent a variety of different assumptions and there has been limited evaluation of how atmospheric, plant, and soil factors interact to control water loss from the soil or how variation in vegetation and soil over a desert habitat may bring about changes in soil water distribution.

In this paper we evaluate three soil models (PALS-SW (Patch Arid Land Simulator-Soil Water), 2DSOIL, and SWB (Soil Water Budget)) with regard to their ability to predict the vertical distribution of soil water as a function of different soil properties, vegetative cover, precipitation, and microclimate. We compare the behavior of the models along a 3-km topographic gradient at the Jornada Long Term Ecological Research (LTER) site in southern New Mexico. These one-dimensional water balance models differ considerably in their treatment and assumptions regarding water transport within the soil and water extraction from the soil via evaporation and transpiration. PALS-SW emphasizes plant physiological controls of water loss, 2DSOIL is a mechanistic model that emphasizes the physical aspects of soil water dynamics, and SWB is a simple water budget model that includes all of the system components but treats them in an aggregated and simplified manner. To compare and contrast the behavior of these models in terms of their ability to integrate plant processes (e.g., rooting distribution and water uptake) and respond to different soil types, we make use of a data set collected along the transect. This semiarid site is characterized by a diversity of plant species and soil types. Each model was independently parameterized by the authors (PALS-SW by J.-L. C., 2DSOIL by Y. P., and SWB by P. R. K.) on the basis of a subset of these data, and its performance was evaluated using independent data (coordinated by J. F. R.). Our objective was to compare how the different models behaved over a gradient of soil types and vegetative cover as well as through the soil profile in different seasons in order to address a number of uncertainties about soil water dynamics in the Jornada Basin. What is the relative importance of vegetative cover and rooting depth as compared to changes in soil properties on the water budget? What is the relative contribution of transpiration and soil evaporation to evapotranspiration, and what are the factors that control these processes? What is the degree of complexity in model structure necessary to obtain reasonable predictions?

Site Description and Data Collection

Vegetation, weather, and soil water data were collected at the Jornada LTER site, 40 km NNE of Las Cruces, New Mexico, in south-central New Mexico (Doña Ana County) [*Wierenga et al.*, 1987]. The climate is semiarid with a mean annual precipitation of about 230 mm, two thirds of which occurs during summer and early autumn as rainfall from convective storms (based on long-term rainfall records from the nearby Jornada Experimental Range Headquarters, U.S. Department of Agriculture, and National Climatic Data Center, Asheville, North Carolina). This site is characteristic of the basin and range topography of the southwestern United States, with the numerous small mountain ranges and intervening broad valleys [*Brown*, 1982].

In 1982, a 2700-m transect was established on a gently sloping, northeast facing piedmont of Mt. Summerford, the northernmost peak of a very small mountain range, the Doña Ana. The transect extends in a SSW direction from an ephemeral lake (playa), located on the basin floor at 1310 m, up gentle alluvial fan slopes (bajada) associated with Mt. Summerford, to its base at 1410 m (Figure 1). Ninety sampling stations are located at 30-m intervals along the transect. A number of vegetation, microclimate, and soil data have been collected at various positions along the transect as part of the LTER study; the timing and specifics of data vary somewhat from year to year. We focused on 1986, which had the largest and most complete data set.

The geomorphology and soils of the study site have been described in detail by *Gile et al.* [1981] and *Lajtha and Schlesinger* [1988]. Soils on the lower half of the transect are oldest (late Pleistocene), with relatively distinct clay and calcium carbonate horizons. Soils are progressively younger toward the upper end of the transect (mid-Holocene) and have poorly developed horizons or none at all. Most soils are sandy loams or loamy sands (Figure 1), and the soil physical and chemical properties, e.g., sand, silt, clay, coarse fragments, $CaCO_3$, and organic carbon, at each station have been reported elsewhere [*Nash and Daugherty*, 1990]. Multiple calcic horizons can be found through the profile, with the shallowest usually between 30 and 50 cm deep. The water table is probably nearest to the surface in the vicinity of the playa but not detectable at 20 m depth, although there are "lenses" of moisture at various depths below the playa associated with its occasional flooding [*Jenkins et al.*, 1988].

Vegetation varies along the transect with generally high cover of grass (*Panicum obtusum*), annuals, and forbs in the playa (stations 1–7); honey mesquite (*Prosopis glandulosa*) and grass (*Muhlenbergia porteri* and *Hilaria mutica*) in the playa fringe (stations 8–10); a broad zone of relatively open mixed vegetation dominated by grasses (e.g., *Aristida longiseta* and *Erioneuron pulchellum*) and annuals and with subshrubs (*Xanthocephalum sarothrae*) and forbs (stations 11–57); a zone dominated by creosote bush (*Larrea tridentata*) in the middle bajada (stations 58–72); and a grassland (e.g., *Bouteloua eriopoda*, *Muhlenbergia porteri*, and *Erioneuron pulchellum*) on piedmont slopes at the base of Mt. Summerford (stations 73–90). See *Cornelius et al.* [1991] for further details.

We grouped the principal species that occur along the transect into five guilds that have similar rooting patterns, seasonal activity, and stomatal responses to soil water deficits (based on work by *Kemp* [1983] and *Cornelius et al.* [1991]): guild 1, annuals (winter or summer active species); guild 2, perennial forbs (species active from spring through autumn); guild 3, grasses (all are C_4, summer active species); guild 4, winter deciduous subshrubs (primarily *Xanthocephalum* and *Zinnia* spp.); and guild 5, evergreen shrub (*Larrea tridentata*). The vegetative cover of each species was measured along a 30-m line transect (perpendicular to the main transect) at each station during spring (mid-April) and autumn (mid-October) to provide estimates of peak above-ground standing crop for the winter/spring and summer periods, respectively. A more intensive estimate of cover and plant phenology that was obtained biweekly on 1-m^2 plots at each station during the 3 years prior (1982–1984) to the present study (1986) was used as an aid to estimate changes in cover during the growing periods

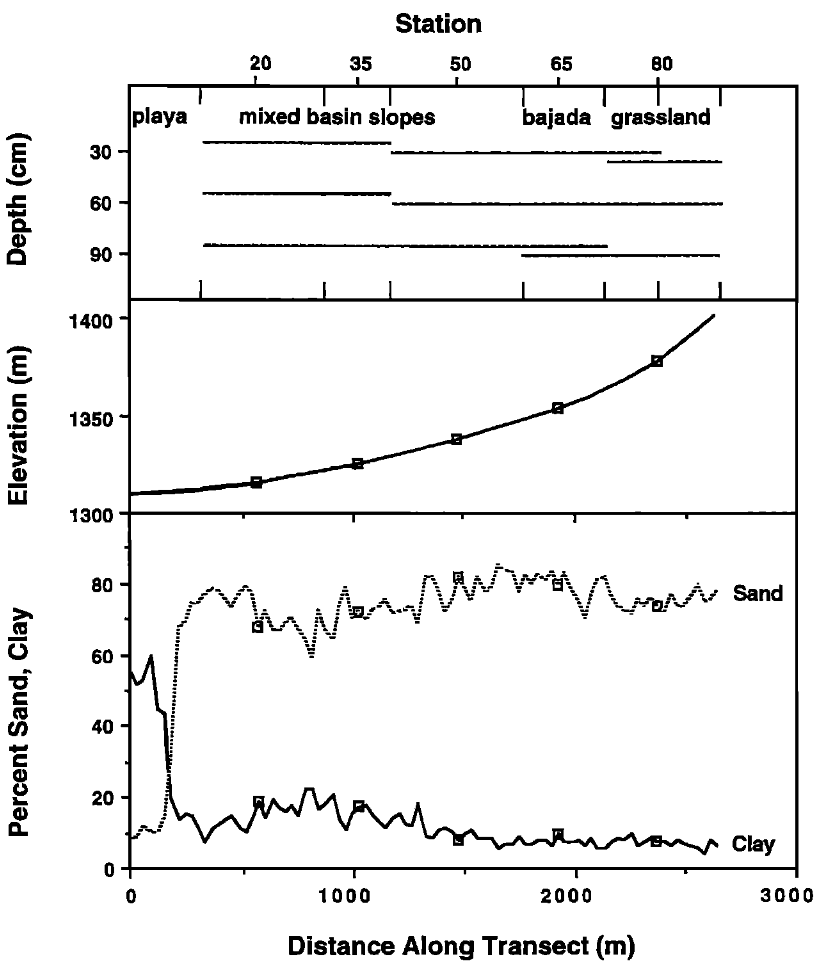

Figure 1. Elevation and distribution of sand and clay along transect at Jornada LTER site in New Mexico. Squares show locations of five stations used in this study. Results of repeated measures analysis of variance for soil water content observations at depths of 30, 60, and 90 cm are shown in upper panel. Significant differences between vegetation zones (see Figure 2) are indicated by nonoverlap in lines.

leading up to the development of maximum cover (Figure 2). We calculated leaf area from plant cover data on the basis of literature values for dominant species in each guild (Table 1).

Volumetric soil water content was measured biweekly at each station at depths of 30, 60, 90, 120, and 150 cm using a neutron-scattering probe (calibrated at the middle of the transect), and soil water potential was measured biweekly at every fifth station at depths of 5, 15, and 30 cm using thermocouple psychrometers. Each model incorporates a relationship between soil water content and moisture retention. Often, this relationship is determined in the laboratory using a pressure chamber. In this study we relied on the overlap in measurements at every fifth station of volumetric soil water content and water potential at a 30-cm depth. The paired measurements were taken from less than 1 m apart and were usually made within 1 or 2 days of each other with no rainfall between measurements. Some measurement pairs were eliminated from the set if they were made over a time interval greater than 4 days or if rainfall occurred between measuring dates. This yielded 17 sets of paired observations of water content and water potential (we omitted the two stations in the playa because they were occasionally flooded and the two stations at the uppermost end of the transect because their water contents occasionally fell below the calibration range of the neutron probe; locations are shown in Figure 1).

Precipitation was measured at each station using a small rain gauge and near the center of the transect using a recording, weighing rain gauge. Solar radiation, humidity, wind speed and direction, air temperature, and soil temperatures at depths of 1, 5, 10, 20, 50, 100, and 200 cm were measured continuously (with hourly averaging) near the center of the transect in an open area vegetated only with annual plants.

Overview of Models

A general comparison of the three models is presented in Table 2. PALS-SW is a mechanistic, one-dimensional model that explicitly accounts for redistribution of water within a soil profile and the extraction of water by evaporation and transpiration. Although soil water distribution is modeled in a manner similar to other models [*de Jong and Cameron*, 1979; *Federer*, 1979; *Hanks*, 1981], PALS-SW has been developed specifically to address questions of water uptake by diverse desert lifeforms undergoing water stress. The soil water model is designed to be used interactively with a plant growth model that

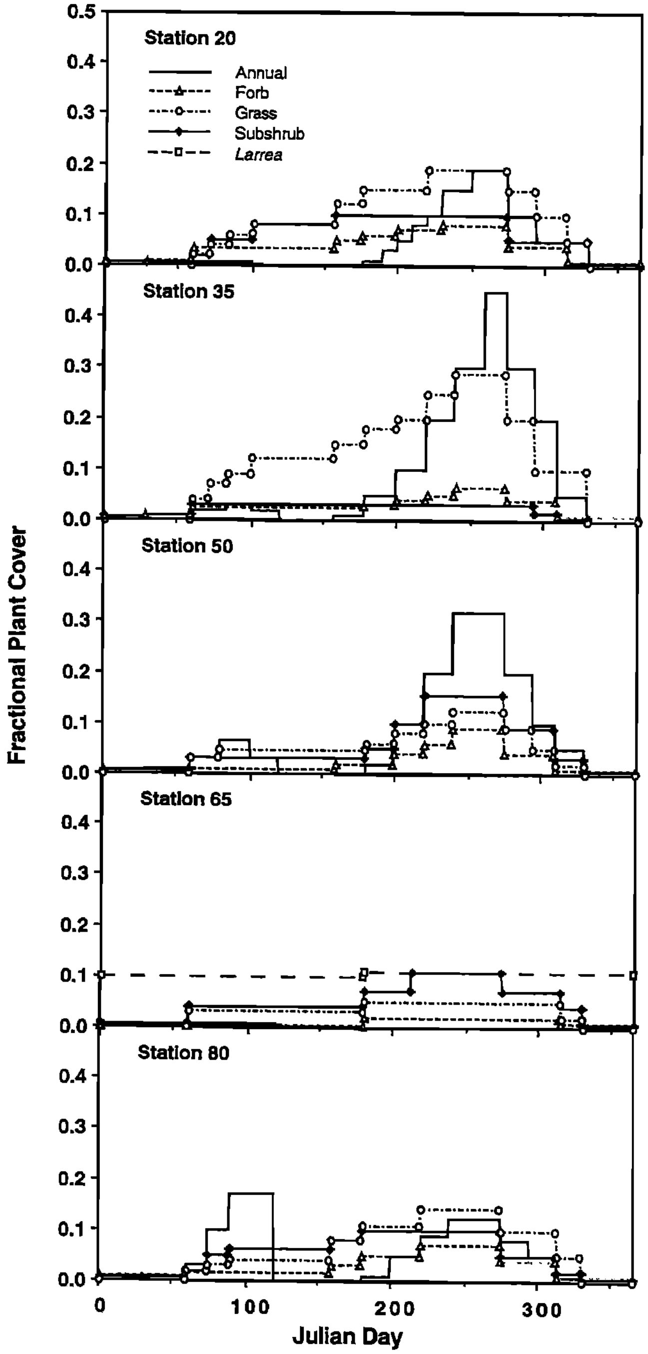

Figure 2. Fraction of the ground covered by each guild during 1986 at five locations along transect shown in Figure 1. These values are approximations based on actual measurements two times during the year (at approximately days 120 and 270) and on general patterns at this site recorded at biweekly intervals for 3 years previously.

predicts activity patterns of different life-forms in the desert [*Reynolds et al.*, 1996].

2DSOIL is a generic model developed to simulate the one or two-dimensional distribution of soil water and temperature in a variety of agricultural soils [*Timlin et al.*, 1996]. It includes detailed treatments of infiltration and redistribution of water and water extraction by plant roots. 2DSOIL has been successfully integrated into crop models to predict transpiration [*Pachepsky et al.*, 1993]. It was used here in a one-dimensional mode.

The SWB model is derived from the simple Versatile Budget model of *Baier and Robertson* [1966]. It was employed because water budget models have fewer parameters to estimate and are not strongly affected by boundary conditions, soil heterogeneity, hysteresis, and other factors that can introduce errors into mechanistic models [*Mustafa et al.*, 1983]. *Noble and Crisp* [1979/1980] successfully simulated soil water in arid regions of Australia using a modified Versatile Budget model. The original Versatile Budget model requires empirical determinations relating actual evapotranspiration to soil water content and daily evaporative demand as well as the change in root density in soil layers during the season. In the SWB model we substantially modified the way in which evaporation and transpiration were determined and made rooting density a function of above-ground cover.

Soil Water Distribution

Infiltration. PALS-SW and 2DSOIL use explicit infiltration routines employing soil hydraulic conductivity. In SWB, soil layers are recharged via rainfall in a cascade fashion, with each layer filled according to its water-holding capacity and with no redistribution among the layers. Each model considers a 1-m-deep soil profile, and while each has a scheme to account for water discharge or recharge at the bottom of the profile, there was no infiltration beyond 1 m, either simulated or in the data. We explored the incorporation of runoff and interception of rainfall into the models, but neither resulted in improvements in the models and therefore were not included. On these gentle slopes runoff is to some extent compensated for by run-on from above (see results and discussion sections).

Soil water flux. Both PALS-SW and 2DSOIL account for water flux between soil layers as defined by the one-dimensional form of the Darcy-Richards equation:

$$\frac{\partial \theta}{\partial t} = \frac{\partial Q}{\partial z} - U \tag{1}$$

where θ is volumetric soil water content (fractional), U is root water uptake rate (cm cm^{-1} d^{-1}), and Q is the downward flux (cm d^{-1}) with depth z (cm) expressed as

$$Q = K\left\{-\frac{\partial \Psi}{\partial z} + 1\right\} \tag{2}$$

where Ψ (in kilopascals; 1 kPa $\approx$ 10.2 cm) is soil matric potential and K (cm d^{-1}) is soil hydraulic conductivity. No horizontal water movement is taken into account. To solve (1), the following information is needed: the dependence of θ on Ψ

Table 1. LAI Calculated From Plant Cover Data

Plant Guild	LAI	Source
Annual	1.17 × cover	*Werk et al.* [1983]; *IBP* [1974]
Forb	2.37 × cover	*IBP* [1974]
Grass	3.60 × cover	*Williamson et al.* [1987]
Subshrub	5.70 × cover	*Ludwig et al.* [1975]; *Deputi and Caldwell* [1975]
Larrea tridentata	0.65 × cover	*Ludwig et al.* [1975]; *Barbour* [1977]

LAI was calculated from cover by first converting cover to leaf biomass (first reference listed) and then converting leaf biomass to leaf area (second reference listed).

Table 2. Comparison of Assumptions and Methods in the Three Models Used in This Study

Attributes	PALS-SW	2DSOIL	SWB
		Soil Layers	
Number	20	24	6
Thickness	5 × 1 cm, 5 × 2 cm, 3 × 5 cm, 7 × 10 cm	5 × 1 cm, 5 × 2 cm, 14 × 5 cm	2 × 10 cm, 4 × 20 cm
		Soil Water Flux	
	Darcy-Richard's equation, predictor-corrector	Darcy-Richard's equation, finite element	none
Moisture Retention	*Campbell et al.* [1993]	*van Genuchten* [1980]	*Campbell et al.* [1993] for water-holding capacity
Hydraulic Conductivity	*Gardner* [1958]	*Gardner* [1958]	none
		Root Distribution	
	Literature (idealized)	Uniform	optimized for SWB
		Evapotranspiration	
Transpiration	*f* (VPD, Stomatal conductance, leaf area, soil water potential)	*f* (canopy energy budget, leaf area, average soil water potential)	*f* (canopy energy budget, average soil water potential, VPD)
Soil Evaporation	*f* (Surface resistance, vapor gradient)	*f* (surface energy budget)	*f* (surface energy budget coupled with model of *Linacre* [1973])
Water Uptake	*f* (soil water potential in layer)	*f* (soil water potential in layer)	*f* (average soil water potential)

(moisture retention curve), the dependence of K on Ψ, root water uptake, and the time-dependent fluxes of water into and out of the soil profile through its boundaries, namely, through the soil surface and through a designated cross section at a significant depth. The Darcy-Richards equation deals with the movement of liquid along gradients of potential energy of water. In the very dry soils of desert regions it has been argued that substantial movement of water occurs in the form of vapor along water potential gradients or along thermal gradients [*Noy-Meir*, 1973]. In 2DSOIL, water movement as vapor along water potential gradients is accounted for by adjusting the relationship between hydraulic conductivity and soil water potential at low matric potentials. Neither PALS-SW nor 2DSOIL account for vapor movement due to thermal gradients; however, the amount of water that moves under these conditions is extremely small relative to the requirements of plants and to the amount lost through evapotranspiration [*Scanlon and Milly*, 1994]. Other studies have shown that incorporating thermal vapor movement into the model has not lead to improved predictions of soil water distribution in dry soils [*Hanks et al.*, 1967; *Jackson et al.*, 1974].

Rooting Distributions and Root Water Uptake

None of the models explicitly accounts for water uptake from soil via roots along a root-soil pathway. This would require detailed knowledge of rooting densities and size distributions, specific root activity and root resistances, and mechanisms that control these two factors. Our approach is to "subtract" water from the soil and distribute this loss through the soil profile in a way that is consistent with our understanding of the control of transpiration water loss in desert plants by stomatal conductance, root distributions, water potentials of the soil in proximity of the roots, and the vapor pressure. Although plant transpiration and root water uptake are calculated differently in each model, each utilizes a standard set of assumptions and information about plant cover and rooting patterns of the plants.

For each model a fixed root distribution is used for the entire year. This assumes that the primary roots of perennials (guilds 2–5) persist throughout the year in the soil profile in a fixed distribution, which represents the primary (nodal) root system from which the fine roots emanate. Primary structural roots in established plants tend to persist and are related to actual water uptake through rapid production and activity of feeder roots [*Nobel*, 1985]. We argue that even for annual plants (guild 1) the assumption of fixed root distribution is a reasonable approximation, since summer annual species grow roots very quickly to their maximum depth, and although winter annual species take considerably longer to reach maximum depth, this development occurs during cooler weather, when evapotranspiration is low [see *Mulroy and Rundel*, 1977]. Although the use of a fixed root distribution is appropriate for this warm desert ecosystem, there are some circumstances or ecosystems under which it would not be appropriate. For example, *Fernandez and Caldwell* [1975] suggested for cold deserts that due to winter soil water recharge, roots of perennials grew progressively deeper during the season to access stored soil moisture as the surface moisture was depleted. *Donovan and Ehleringer* [1994] found that some perennials could not utilize soil moisture in portions of their root zone (primarily shallow layers) during some periods of time because of lack of active roots. In warm deserts, moisture is readily used by the guilds of plants that are active above ground when the moisture comes [*Kemp*, 1983; *Kemp et al.*, 1992].

In PALS-SW the root water uptake from each soil layer (U_j, in centimeters per day) is the sum of the individual transpiration losses of each guild for that layer (T_{ij}), i.e.,

$$U_j = \sum_{i=1}^{5} T_{ij} \tag{3}$$

with T_{ij} given in (5). In 2DSOIL and SWB, total root water uptake from any soil layer (U_j) is the transpiration loss of the canopy (T_c) partitioned to that layer according to total root fraction in that layer ($\mathrm{Root}fr_{ij}$), i.e.,

$$U_j = T_c \sum_{i=1}^{5} \mathrm{Root}fr_{ij} \tag{4}$$

Table 3. Prescribed Root Distributions

Depth, cm	Guild				
	Annual	Forb	Grass	Subshrub	*Larrea*
		*Root Distributions**			
0–10	0.3	0.2	0.2	0.1	0.0
10–20	0.5	0.2	0.3	0.2	0.1
20–30	0.2	0.2	0.2	0.2	0.2
30–40	0.0	0.2	0.2	0.2	0.2
40–60	0.0	0.2	0.1	0.2	0.3
60–80	0.0	0.0	0.0	0.05	0.1
80–100	0.0	0.0	0.0	0.05	0.1
		Uniform†			
0–10	0.25	0.1	0.1	0.1	0.1
10–20	0.25	0.1	0.1	0.1	0.1
20–30	0.25	0.1	0.1	0.1	0.1
30–40	0.25	0.1	0.1	0.1	0.1
40–60	0.0	0.2	0.2	0.2	0.2
60–80	0.0	0.2	0.2	0.2	0.2
80–100	0.0	0.2	0.2	0.2	0.2
		Optimized for SWB Model‡			
0–10	0.2	0.1	0.1	0.1	0.0
10–20	0.3	0.2	0.15	0.1	0.1
20–30	0.2	0.15	0.2	0.1	0.25
30–40	0.15	0.15	0.15	0.15	0.25
40–60	0.15	0.2	0.15	0.25	0.3
60–80	0.0	0.15	0.15	0.20	0.1
80–100	0.0	0.05	0.10	0.10	0.0

*Based on estimates from observed rooting patterns of desert plants. Estimates for annuals were taken from *Forseth et al.* [1984]; those for forbs, grass, and subshrubs were taken from *IBP* [1974]; and those for *Larrea* were taken from J. Brisson (personal communication, 1996).

†Based on the assumption that roots can remove water uniformly throughout the profile.

‡Based on an optimized fit to soil water with the SWB model.

with T_c given in (11) (2DSOIL) or (17) (SWB). The specific distribution of the roots with depth ($Rootfr_{ij}$; unitless) in layer j were specific for each guild i and initially assigned on the basis of generalizations from published studies in warm desert regions [*Cannon*, 1911; *International Biological Program* (*IBP*), 1974; *Moorhead et al.*, 1989] (see Table 3). Later simulations involved changing distributions to achieve better fits of modeled soil water distribution to the data sets (Table 3 and the results and discussion sections).

Evapotranspiration

In mesic environments, evapotranspiration is strongly controlled by radiation, turbulence, and vapor pressure deficit [*Monteith and Unsworth*, 1990] and by the nature of the overall plant canopy, primarily as it relates to the planetary boundary layer [*Jarvis and McNaughton*, 1986]. In deserts and semiarid regions, water loss is likely to be controlled more by plant and soil factors than by the atmosphere. In fact, actual daily ET is usually much less than the potential (atmospheric) ET in arid regions [*Sammis and Gay*, 1979; *Parton et al.*, 1981; *Nichols*, 1992]. Plant factors, including species composition and cover [*Cable*, 1980], phenology [*Kemp*, 1983; *Donovan and Ehleringer*, 1994], stomatal response [*Schulze*, 1986a, b], and rooting patterns [*Cable*, 1977; *Moorhead et al.*, 1989], as well as soil factors such as texture [*Alizai and Hulbert*, 1970; *Noy-Meir*, 1973], interact with rainfall and vapor pressure to produce particular patterns of evaporation and transpiration.

The specific methods for calculating transpiration, evaporation, and water uptake varies for each model. However, each uses a surface energy budget approach that treats transpiration and evaporation as separate, noninteracting entities [e.g., *Ritchie*, 1972]. In plant communities with low plant cover, an alternative approach is to treat them as coupled, interacting entities [e.g., *Shuttleworth and Wallace*, 1985]. Using SWB, we investigated both approaches and found small quantitative differences in predicted transpiration and soil evaporation amounts, which were related to plant cover (primarily) and soil evaporative loss (slightly). The different approaches did not result in qualitatively different predictions and we deemed the simpler, uncoupled approach reasonable for the plant covers encountered in this study (peak cover varying from 30% to 90% and peak leaf area index (LAI) from 0.9 to 1.9). *Stannard* [1993] also found that a simple energy budget model performed as well as a more complex, interactive one when used to calculate evapotranspiration in a semiarid rangeland.

Transpiration. There is considerable uncertainty in our understanding of the extent and mechanisms by which transpiration is reduced under conditions of aridity [*Kramer*, 1988; *Passioura*, 1988]. There is debate as to whether the plant is directly sensitive to the wettest layers (as indicated by *Ritchie* [1972] and *Waring and Schlesinger* [1985]) or responds to the average water potential of the soil [*Johnson and Norton*, 1979; *Fahey and Young*, 1984; *Fonteyn et al.*, 1987]. These two positions were addressed, to a degree, by the different approaches taken in the models. The approach in PALS-SW was essentially that of a "wettest-layer response." In this case, plant transpiration (conductance) from any soil layer is directly related to the water potential of that layer, thus allowing water to be extracted from one layer (in proportion to the amount of roots in the layer) irrespective of the water potentials of the other layers. The approaches in 2DSOIL and SWB is an "average soil water response," in which canopy conductance is related to the average water potential of the soil layers, weighted by the distribution of roots. In this case, water extraction from the soil slows when any portion of the soil dries and lowers the overall canopy conductance. However, regardless of the way in which plants sense soil water, it is observed that soil water deficits as well as atmospheric vapor deficits restrict water loss through reduction of stomatal conductance [*Schulze*, 1986a, b; *Kramer*, 1988; *Franco et al.*, 1994]. In PALS-SW, transpiration is a direct function of individual leaf stomatal conductance, which in turn is a function of soil water deficit (water potential) and atmospheric vapor deficit (VPD). Atmospheric water deficits also reduce stomatal conductance [*Bunce*, 1981; *Franco et al.*, 1994]. In 2DSOIL and SWB, transpiration is primarily a function of the canopy energy budget and secondarily of canopy conductance, which in 2DSOIL is a function of soil water and which in SWB is a function of soil water and VPD. Thus differences among the models in the simulation of the soil water profile, particularly deep soil water, would be expected to stem from these differences.

In PALS-SW, transpiration of guild i (T_{ij}, in centimeters per day) represents the fractional part associated with each soil layer via $Rootfr_{ij}$. This fractional part is a function of stomatal conductance (G), leaf area per unit ground area (LAI_i), and the leaf-to-air vapor pressure difference (VPD), which is assumed to be the same for all guilds and equal to the difference between saturated vapor pressure at ambient temperature and ambient vapor pressure at mean daytime temperature divided by air pressure (P). T_{ij} is partitioned into fractions directly related to uptake from each soil layer j via $Rootfr_{ij}$:

$$T_{ij} = \mathrm{VPD}/P \times G_{ij} \times \mathrm{LAI}_i \times \mathrm{Root}fr_{ij} \quad (5)$$

This equation does not include a boundary layer conductance, which could be somewhat different among the various guilds. However, such a conductance is large relative to G, for most desert plants which have small leaves, and incorporating various boundary layer conductances in series with G reduced T_{ij} by less than 1% (except for wind speeds below 0.5 m/s). Total canopy transpiration (T_c, in centimeters per day) is

$$T_c = \sum_{i=1}^{5} \sum_{j=1}^{J} T_{ij} \quad (6)$$

In this approach, stomatal conductance of each plant guild is partitioned into fractions that are associated with each soil layer, G_{ij} (centimeters per day). This is based on the assumption that stomatal conductance is determined by the local soil–root water potential gradients and relatively independent of soil–root water potential gradients in other soil layers. This could occur if root hydraulic conductivity is relatively low compared to moist soil [*Passioura*, 1988] and if root shrinkage or other mechanisms uncoupled roots from soil as it dries [*Nobel and Cui*, 1992a, b]. Values of G_{ij} were determined separately for each guild based on empirical responses for the principal species. For guilds 1–4 we assume an exponential relationship between G_{ij} (moles per square meter per second) and the water potential of soil layer j, Ψ_j (in kilopascals) [*Depuit and Caldwell*, 1975; *Kemp and Williams*, 1980; *Ehleringer*, 1983], and a linear relationship between G_{ij} and VPD [*Bunce*, 1981]:

$$G_{ij} = ae^{(b\Psi_j)}[1 - 0.1 \times \mathrm{VPD}] \qquad i = 1, \cdots, 4 \quad (7)$$

For annuals, $a = 1.2$ and $b = 0.0011$ [*Ehleringer*, 1983]; for forbs and grasses, $a = 1.0$ and $b = 0.00125$ [*Kemp and Williams*, 1980]; and for subshrubs, $a = 0.56$ and $b = 0.00085$ [*Depuit and Caldwell*, 1975]. For guild 5, represented by the shrub *Larrea*, we used a multiple linear regression to predict stomatal conductance from soil water potential and VPD based on data of *Franco et al.* [1994]:

$$G_{5j} = 0.52 + 0.081 \times \Psi_j - 0.064 \times \mathrm{VPD} \quad (8)$$

In 2DSOIL, potential canopy transpiration (PE_c) is calculated using an energy budget approach (from the model GLYCIM [*Acock and Trent*, 1991]). Transpiration of guild i (T_i) is calculated from the LAI_i and cover of guild i, and, to account for the effect of reduced soil moisture, a scaling factor, A_i:

$$T_i = \mathrm{PE}_c \times \mathrm{LAI}_i \times \mathrm{Cover}_i \times A_i \quad (9)$$

A_i relates transpiration of the ith guild to the root-weighted average soil water potential of the jth soil layer containing roots of that guild [*Belmans et al.*, 1983]:

$$A_i = \sum_{j=1}^{J} (a_{ij} \times \mathrm{Root}fr_{ij}) \quad (10)$$

where a_{ij} [see *Timlin et al.*, 1996] varies linearly from a maximum of 1 in moist soil (−80 kPa and above) to 0 at limiting soil water potential for each guild: annuals, −1500 kPa; forbs and grasses, −4000 kPa; and subshrubs and *Larrea tridentata*, −5000 kPa [*Odening et al.*, 1974; *Depuit and Caldwell*, 1975; *Kemp and Williams*, 1980; *Ehleringer*, 1983]. The total canopy transpiration is

$$T_c = \sum_{i=1}^{5} T_i \quad (11)$$

In SWB we calculated the atmospheric-limited potential evapotranspiration (PE), which was partitioned into canopy and soil components, and then accounted for reduced plant cover, the effects of reduced soil moisture, and reduced stomatal conductance on canopy transpiration. From *Campbell* [1977, p. 140], PE is calculated as a simple function of maximum daily solar radiation ($S_{\max}$, in watts per square meter) and maximum air temperature (T_{air}, in degrees Celsius):

$$\mathrm{PE} = \{a(T_{\mathrm{air}} + b)S_{\max}\}/\mathrm{LE}, \quad (12)$$

where LE (in joules per gram) = 2501 − 2.4 × T_{air}, $a = 0.025°\mathrm{C}^{-1}$, and $b = 4.6$. Equation (2) works as well as more complex energy budget equations for the western United States [*Campbell*, 1977; *Jensen et al.*, 1990]. Using the maximum or midday values of solar radiation and temperature produce the least error in calculated PE [*Beven*, 1979]. The daily maximum PE (in grams per square meter per second) is then integrated over the day by assuming a sinusoidal course of energy and temperature (i.e., PE × $2/\pi$) and multiplied by photoperiod and 0.0001 $\mathrm{m}^2\ \mathrm{cm}^{-2}$ to convert to centimeters per day. The potential evapotranspiration is then partitioned between the canopy and soil surfaces according to Beer's law, which in effect partitions the solar radiation component of the energy budget (by far the largest) via interception by the canopy [see *Ritchie*, 1972; *Nichols*, 1992; *Stannard*, 1993]:

$$\mathrm{PE}_c = \mathrm{PE} \times (1 - e^{-k\times\mathrm{LAI}}) \quad (13)$$

and

$$\mathrm{PE}_s = \mathrm{PE} \times e^{-k\times\mathrm{LAI}} \quad (14)$$

where PE_c and PE_s are the potential evaporation amounts for the canopy and soil surface, respectively. The value of k that governs radiation extinction by the canopy varies with the sun angle, distribution of plants, and arrangement of leaves. For LAI between 0.2 and 2.0 and a single midday calculation, k varies between 0.5 and 0.75, for randomly arranged plants and leaves [*Nichols*, 1992]. Parameter fitting for this parameter indicated that a value of $k = 0.6$ yielded the best overall fit to water content data (see parameterization section). The effect of reduced canopy conductance (due to soil moisture deficit and vapor pressure deficit) on canopy transpiration in SWB was accounted for using the equation of *Campbell* [1977, p. 143]:

$$\frac{\mathrm{AE}}{\mathrm{PE}} = \frac{\Delta + \gamma \times r_{va}/r_e}{\Delta + \gamma \times (r_{va} + r_{vc})/r_e} \quad (15)$$

where AE is the reduced transpiration, Δ is the slope of the saturation vapor pressure–temperature curve, and γ is the psychrometric constant. The resistance r_e was calculated from the relationship of *Campbell* [1977, p. 103]; the resistance r_{va} was calculated from *Monteith and Unsworth* [1990, p. 248]. For r_{va} we assumed that $z_0 = 0.05$ m, $d = 0.5$ m, and $z = 1.5$ m. The resistance r_{vc} is the reciprocal of the bulk canopy conductance, G_c, calculated as

$$G_c = 2 \sum_i (G_i \times \mathrm{LAI}_i) \quad (16)$$

where G_i is the stomatal conductance of guild i, LAI_i is the leaf area index, and the factor 2 reflects the fact that nearly all of the plants have stomates on both surfaces. We explored using both linear (as in 2DSOIL) and exponential (as in PALS-SW) relationships of G_i to soil water potential. The exponential relationship provided a better fit to soil moisture data because it allowed extraction of soil moisture at low water potentials (necessary for water withdrawal from subsurface soils during dry periods) and resulted in less rapid soil water withdrawal at moderate soil water potentials. Thus (7) and (8) were used to calculate G_i as a function of soil water potential, except that only one value of G_i was calculated for the whole soil profile, so that Ψ_j was replaced by the root-weighted average soil water potential of the profile. From (3) and (5), total canopy transpiration (T_c) is given by

$$T_c = \frac{\mathrm{AE}}{\mathrm{PE}} \times \mathrm{PE}_c \tag{17}$$

Evaporation. In PALS-SW, soil evaporation, E_s, is calculated as

$$E_s = \frac{e_s - e_a}{P_a r_L} \tag{18}$$

where e_s and e_a are the vapor pressures of the soil surface and air, respectively, P_a is the atmospheric pressure, and r_L (in moles per square meter per second) is the resistance for vapor diffusion, which is a sum of the resistance of the soil boundary layer, r_b, and the surface, r_s. The boundary layer resistance r_b (in seconds per meter) is calculated as

$$r_b = \frac{d}{D_v} \tag{19}$$

where d (in meters) is the thickness of the soil boundary layer ($d = 0.004 \times$ (clod size/wind speed)$^{1/2}$, with clod size equal to 0.05 m over the entire transect), and D_v (in square meters per second) is the diffusivity of the vapor in the air:

$$D_v = 2.126 \times 10^{-5} + 1.48 \times 10^{-7} \times T_{\mathrm{air}} \tag{20}$$

The surface resistance r_s (in seconds per meter) is estimated according to the water status of the top 1 cm soil layer [*van de Griend and Owe*, 1994]:

$$r_s = 10 \qquad \theta_1 > 0.15 \tag{21a}$$

$$r_s = 10e^{[35.6\times(0.15-\theta_1)]} \quad \theta_1 < 0.15,\ \Psi_1 > -10000 \text{ kPa} \tag{21b}$$

$$r_s = \infty \qquad \Psi_1 < -10000 \text{ kPa} \tag{21c}$$

In 2DSOIL and SWB, E_s is related to a potential evaporation of a partially vegetated soil surface, PE_s, but must be further reduced to reflect soil surface drying. In the case of 2DSOIL the evaporation from bare soil is equal to PE_s until the supply of water to the soil surface node cannot keep pace with PE_s; at this time, E_s is equal to the water flux to the surface node. In SWB we calculated E_s from the simple relationship proposed by *Linacre* [1973] and employed by *Johns* [1982] for bare soil. If the water potential of the top layer is greater than −10000 kPa, then evaporation is extracted from that layer as

$$E_s = \mathrm{PE}_s \qquad (\theta_1/\theta_{1,\mathrm{sat}}) > (\mathrm{PE}_s/\varepsilon)^{1/2} \tag{22a}$$

$$E_s = \varepsilon(\theta_1/\theta_{1,\mathrm{sat}})^{\rho} \qquad (\theta_1/\theta_{1,\mathrm{sat}}) \leq (\mathrm{PE}_s/\varepsilon)^{1/2} \tag{22b}$$

where θ_1 and $\theta_{1,\mathrm{sat}}$ are the water content and saturated water content, respectively, of the top soil layer, and ε and ρ are parameters that depend primarily on depth of the soil layer considered and soil texture. If the water potential of the top layer falls below −10,000 kPa, then evaporation is extracted from layers 1 and 2 in proportion to their water contents:

$$E_s = \varepsilon(\theta_{1+2}/\theta_{1+2,\mathrm{sat}})^{\rho} \tag{23}$$

Johns [1982] evaluated this equation during all seasons and for different depths of the evaporative layer for bare soil and found that the calculated evaporation was not very sensitive to parameters ε or ρ, which varied from about 0.8 to 1.0 and 2 to 2.3, respectively, for a soil depth of 10 to 20 cm. We also found that soil evaporation was not greatly affected by changes in these parameters but did find that predicted evaporation was very sensitive to the value of saturated water content ($\theta_{1,\mathrm{sat}}$ or $\theta_{1+2,\mathrm{sat}}$). Most values that approximated actual saturated water content resulted in insufficient soil evaporation. This was because the maximum value of soil water content of the top two soil layers was field capacity and thus the denominator of (24) was always much greater than the numerator. We found that a water content near field capacity, namely VWC at −20 kPa, provided the best fit to the observed soil water potentials at depths of 5 and 15 cm. This value is somewhat greater than the more usual field capacity of −30 kPa and apparently accounted for the fact that some water was retained in the surface layers and available for evaporation while the profile was draining to field capacity. In order to have just one field capacity parameter in the SWB model for soil evaporation as well as previously calculated water holding capacity, we opted to use the value that gave best fit to soil water content over the entire soil profile from 30 to 90 cm, which was −25 kPa. It is likely that mean values of measurements of water content over the profile were more reliable than those of water potential near the surface. Using water content at field capacity in place of θ_{sat} in (23) and (24) also had the advantage of allowing for variation in evaporation with soil texture [see *Alizai and Hulbert*, 1970; *Hillel*, 1980], since water content at field capacity was highly variable with texture, whereas $\theta_{1,\mathrm{sat}}$ was not.

Model Parameterization

In Situ Moisture Retention

Two equations were used to describe the relationship between soil water potential and water content. PALS-SW and SWB employed *Campbell et al.*'s [1993] equation to describe soil moisture retention:

$$\theta = \theta_1\left(1 - \frac{\ln(-\Psi)}{\ln(-\Psi_0)}\right) \tag{24}$$

where θ_1 and Ψ_0 are empirically determined parameters. Parameter values were determined by least squares fitting of the equations to the water content–water potential data collected at each station (Table 4). For SWB the soil moisture retention relationship is used to calculate the water-holding capacity of the soil layer. The model predictions were relatively insensitive to the value chosen for residual water content but were highly sensitive to the value chosen for field capacity. We found that the best fit to observed soil water contents over the profile was obtained with a value for field capacity as the water content at −25 kPa and the residual water content as that at −10000 kPa. So for SWB only, field capacity was taken as −25 kPa.

The 2DSOIL model uses *van Genuchten*'s [1980] equation to describe moisture retention:

Table 4. Soil Texture, Calculated Saturated Hydraulic Conductivity, and Parameter Values for Moisture Retention Equations of *van Genuchten* [1980] and *Campbell et al.* [1993] for Soils Along the Transect at a 30-cm Depth

								Campbell et al. [1993] Equation		
	Percent			*van Genuchten* [1990]				One Parameter:	Two Parameter	
Station	Clay	Sand	K_s, cm d^{-1}	θ_s	θ_r	α, cm	n	θ_1	θ_1	ln $(-\Psi_0)$
10	14.98	74.57	99	0.381	0.090	0.00145	1.844	0.189	0.221	12.26
15	13.52	75.94	128	0.381	0.001	0.350	1.144	0.135	0.158	12.36
20	19.08	67.98	47	0.404	0.0	0.150	1.147	0.284	0.266	14.71
25	18.00	70.43	57	0.396	0.091	0.0295	1.678	0.266	0.246	14.68
30	18.54	67.36	52	0.408	0.092	0.00324	1.732	0.279	0.301	12.95
35	17.15	72.29	67	0.393	0.035	0.343	1.316	0.140	0.137	14.09
40	14.63	72.09	105	0.392	0.020	0.00465	1.328	0.234	0.248	13.24
45	9.33	82.35	272	0.359	0.0	0.139	1.251	0.124	0.165	11.46
50	7.83	82.28	355	0.359	0.008	0.0542	1.311	0.120	0.164	11.14
55	8.69	79.26	305	0.393	0.0	0.217	1.302	0.075	0.094	11.72
60	7.01	82.63	411	0.355	0.017	0.00064	2.346	0.072	0.136	10.27
65	9.52	79.74	262	0.385	0.045	0.00245	1.859	0.138	0.158	12.60
70	8.70	76.73	304	0.396	0.0	0.117	1.237	0.159	0.226	11.03
75	8.32	72.98	325	0.377	0.0	0.271	1.307	0.066	0.097	10.64
80	7.61	74.38	370	0.359	0.026	0.00063	8.531	0.078	0.092	12.20

For the one-parameter *Campbell et al.* [1993] equation, Ψ_0 is set to -10^6 kPa. In this case, regression of θ_1 against clay is $\theta_1 = -0.115 + 0.0107 \times$ percent of clay (which can be compared to Campbell et al.'s values: $\theta_1 = 0.03 + 0.007 \times$ percent of clay).

$$\theta = \frac{\theta_s - \theta_r}{[1 + (\alpha|\Psi|)^n]^m} + \theta_r \tag{25}$$

where θ_s is saturated water content, θ_r is residual water content; $m = 1 - 1/n$; and n, m, and α are empirical parameters. Saturated water content, θ_s, is assumed equal to total porosity, estimated from the bulk density (BD) [*Campbell*, 1985]:

$$\theta_s = 1 - (\mathrm{BD}/D) \tag{26}$$

where D is density of the solid phase ($\approx$2.65 g cm^{-3}). Bulk density values were not measured at each station, but were found to be correlated with the sand fraction (S) of the soil on the basis of an analysis near the center of the transect [*Wierenga et al.*, 1987]:

$$\mathrm{BD} = 1.009 + 0.835S \qquad r^2 = 0.64$$

Thus θ_s was indirectly determined from station sand content. The remaining parameters (α, n, θ_r) for the van Genuchten equation were determined using nonlinear least squares fitting to the in situ moisture retention data at each station.

Having obtained parameter values for moisture retention relationships, we then sought to ascertain if there was a relationship between these parameters and soil texture along the transect. Multiple correlation analysis revealed that the parameters for the *Campbell et al.* [1993] equation were strongly correlated with sand and clay of the soil, whereas those for the *van Genuchten* [1980] equation were not. Since sand and clay were highly interdependent, we used a simple regression to relate the parameter value(s) of the *Campbell et al.* [1993] relation to the clay content of the soil. The moisture-retention parameter values determined from soil clay content were thus used in all of the following model exercises using PALS-SW and SWB, whereas parameter values for the van Genuchten relationship in 2DSOIL were determined directly from each station's moisture retention curve.

Hydraulic Conductivity

Both PALS-SW and 2DSOIL use *Gardner*'s [1958] formulation to describe unsaturated soil hydraulic conductivity to capillary regions:

$$K = K_s/(1 + \Psi/\Psi^*)^p \tag{27}$$

with K_s being the saturated hydraulic conductivity, $p = 3$, and $\Psi^* \approx -26$ kPa (from the assumption that at field capacity $K/K_s = 0.1$ and $\Psi = -30$ kPa). The resulting function gave a larger K compared to the more commonly used approach of *Mualem* [1976] but resulted in a better fit to the soil water data following soil water recharge from rainfall and after evaporation during dry periods, as has been reported for other studies with coarse textured soils [e.g., *Vereeken*, 1992].

Estimates of saturated hydraulic conductivity are based on values measured nearby in several different soil layers, using different methods [*Wierenga et al.*, 1989]. We pooled the values from all estimates and calculated mean K_s for each soil layer. The mean values of K_s (in centimeters per day) were correlated with the clay fraction (C),

$$K_s = 1443e^{(-17.9C)} \qquad r^2 = 0.48 \tag{28}$$

and are comparable (Table 4) to those measured in other coarse textured soils [*Hills et al.*, 1992; *Rawls et al.*, 1992].

Since relationships relating water potential to water content [following *Campbell et al.*, 1993] and hydraulic conductivity were based on parameters determined from soil texture, we were able to vary the parameters over the profile according to texture. However, all stations had roughly similar texture values over the profile [*Nash and Daugherty*, 1990], justifying the use of the same water retention and conductivity parameters over the profile. One limitation of this uniform characterization is that there are calcic and argillic horizons (lenses) at various depths along most of the transect. They are relatively shallow (<50 cm) and poorly developed at the upper end of the transect and deeper (>50 cm) and more developed at the lower end. While calcic horizons can act as barriers to vertical water movement [*Hennessy et al.*, 1983], we observed soil water recharge below the horizons along the entire transect.

Sampling Stations

Initial model development and estimation of values for remaining undetermined parameters was based on fitting the

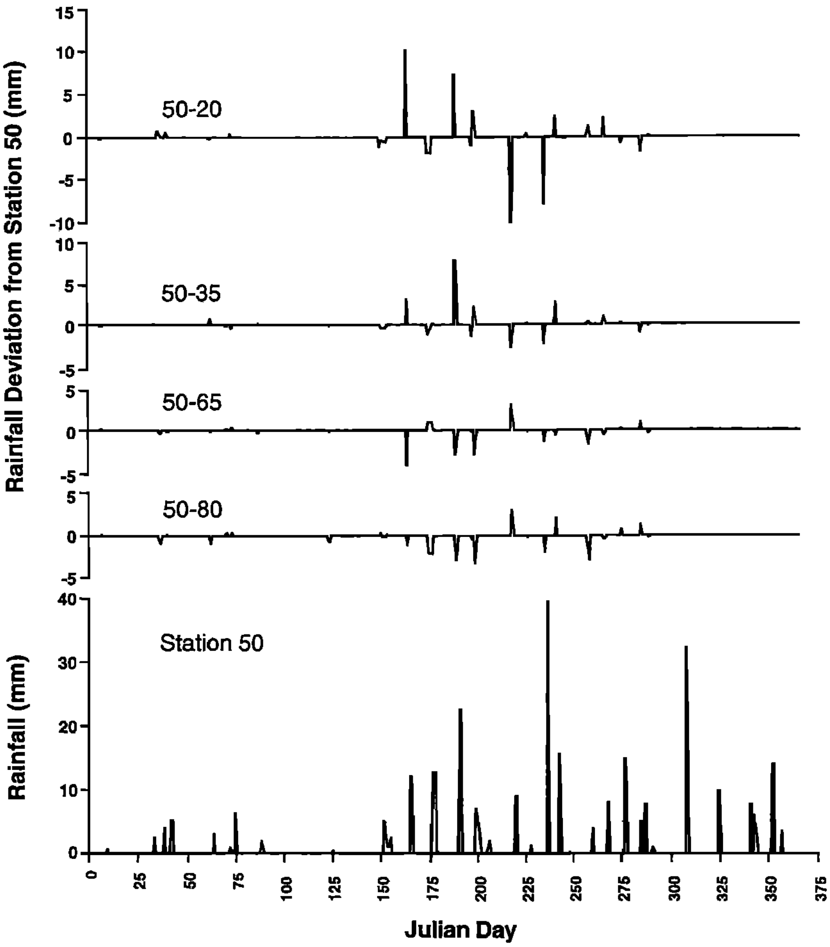

Figure 3. Rainfall at station 50 during 1986 and the deviation between station 50 and other stations used in study. After day 300, rainfall was measured only at station 50; thus no deviations could be determined, and rainfall used for simulations was the same.

models to water content data from station 50. Parameter values were estimated in an iterative fashion to obtain a minimum root mean square error (RMSE) between observed and simulated water contents at the three depths over the rooting zone (30, 60, and 90 cm). Once the parameter values were established for each model for station 50, they were left fixed for all simulations for the other stations presented in this paper.

For model validation and comparisons we used data from stations 20, 35, 65, and 80. These stations encompass a substantial amount of the variation in slope, soil texture, and vegetative cover along the transect. In addition, these stations received different amounts of rainfall in 1986 (Figure 3).

Results and Discussion

Soil Water Dynamics

Effects of root distributions. The results of simulations of soil water content at station 50, using the initial prescribed root distributions (see Table 3) are shown in Figure 4. The biggest discrepancy between observed and predicted values was with SWB, which failed almost completely to predict soil moisture decline at both 60 and 90 cm during spring and summer soil drying (Figure 4). Since there is no soil water redistribution in SWB, the simulated water distribution at the 60 and 90 cm depths depended upon the assignment of root fractions and plant water uptake from these depths. The initial root distributions used in SWB allowed for only minimal water extraction by roots at depths below 40 cm, and only *Larrea* and subshrubs were originally prescribed with roots below 80 cm. Thus to achieve significant water withdrawal from below 40 cm, we modified the root distribution for SWB. To do this, we changed the root distribution of one guild at a time to achieve a minimum RMSE in simulated versus observed water contents for 30, 60, and 90 cm. This yielded the "optimized" root distribution shown in Table 3.

Since SWB does not account for redistribution of soil water, it might be argued that the root distribution required to achieve a good fit would be unrealistic, since it must account for both root uptake and simultaneous redistribution. However, the amount of redistribution of water was found to be quite low when it was examined independently using 2DSOIL. This is because of the relatively low matric potentials and resultant low hydraulic conductivities that existed during most of the year at the 60- and 90-cm depths in both the modeled and real desert systems. Thus water loss from the deeper soils must be accomplished largely by root uptake.

Since PALS-SW and 2DSOIL also had errors in their predictions of soil water at the 60- and 90-cm depths using the original prescribed root distributions, we tested whether changes in root distribution assignments in these two models would improve predicted soil water content. In 2DSOIL we

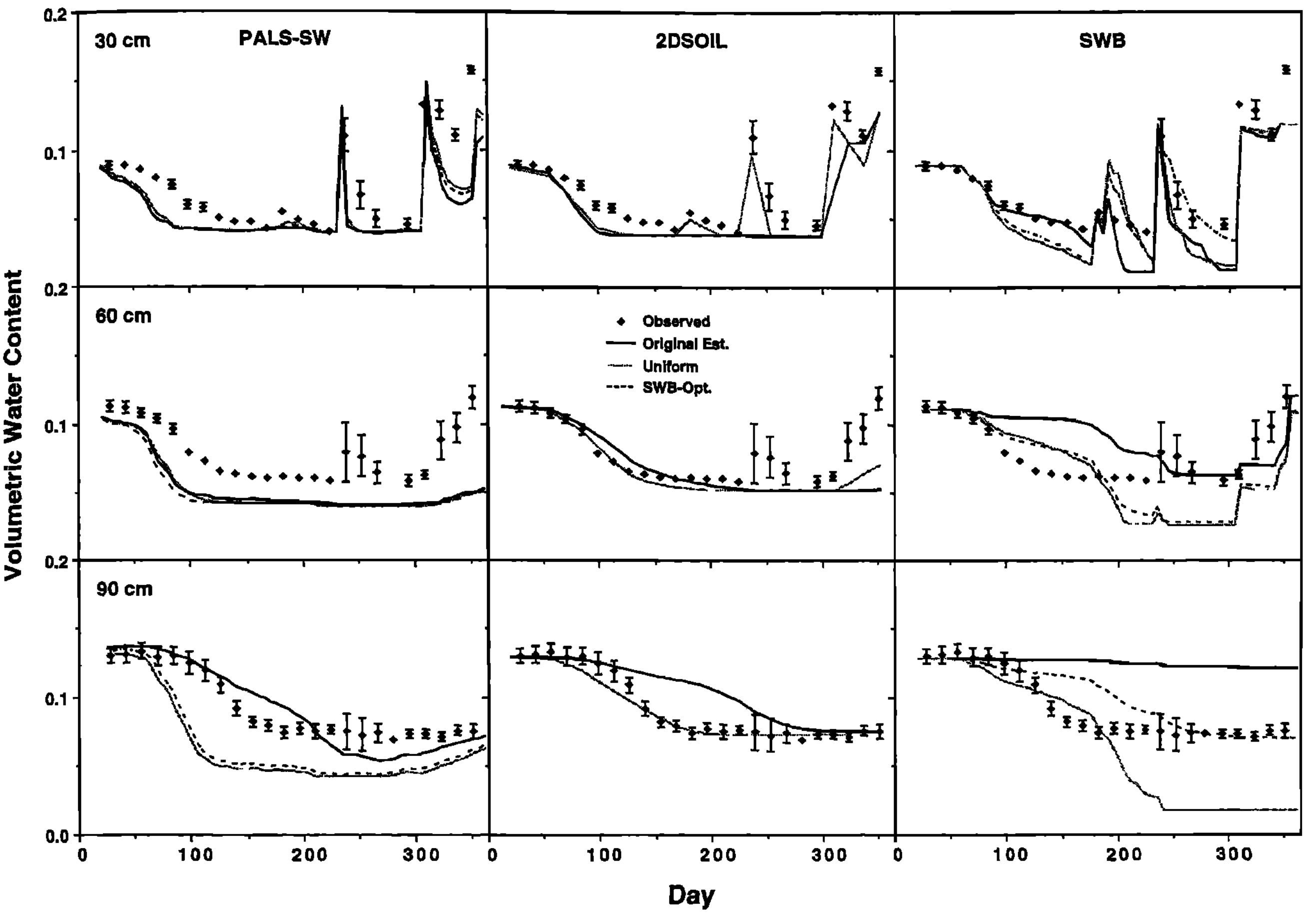

Figure 4. Effects of simulated root distributions upon predicted volumetric soil water contents for station 50 at three depths for the three models. Points are observed values.

found that a uniform distribution throughout the top meter (except for annuals) gave the minimum error. Since the actual root mass in the field is unlikely to be uniformly distributed, this suggests that the ability of roots to extract water in the desert system may be somewhat independent of the actual distribution of root mass, and, instead, is related to the maximum depth of rooting and the activity patterns of the roots. Root "activity" in the broad sense could mean uptake capacity of existing roots [*BassiriRad and Caldwell*, 1992] or, more likely, very rapid growth of ephemeral, fine feeder roots into areas of moist soil for arid-adapted plants [*Sala and Lauenroth*, 1982; *Nobel*, 1985; *Caldwell and Richards*, 1986]. The structure of PALS-SW did not allow for determining the optimum root distribution for each guild, so we tested the uniform root distribution and the SWB-optimized distributions but found that the original root distribution provided the best fit. Thus a different root distribution was utilized in each model for all further simulations: the original estimate for PALS-SW, uniform for 2DSOIL, and optimized for SWB (Table 3).

Although changes in root distribution were undertaken principally to explore how they affected water loss rates of the lower root profile, the effects of these changes also impacted the upper profile water content (30 cm) (see Figure 4). A closer examination of the surface water effects of root distribution changes can be seen in the changes in water potential recorded at 5, 15, and 30 cm. Changes in root distributions had little impact on predicted water potentials of the near-surface soils (Figure 5). Impacts on the predicted water potentials at 15 and 30 cm varied with the particular model. Changes in root distribution had virtually no effect on water potentials at 15 and 30 cm predicted by PALS-SW, whereas the water potentials at these depths were moderately effected in 2DSOIL and more so in SWB. Using optimal root distributions, the predictive errors were similar for PALS-SW and 2DSOIL. The models tended to overestimate water potentials in summer (i.e., did not predict sufficient drying) and underestimated water potentials in early spring and autumn (i.e., predicted excessive drying). The SWB model predicted early spring drying more accurately than the other two models but did not predict sufficient summer and autumn drying. Thus the RMSEs were roughly similar for all models (Tables 5 and 6).

The best fit to water content over all depths was obtained with 2DSOIL (Table 5). The principal error associated with 2DSOIL was the failure to predict recharge at 60 cm following days 240 and 300 (Figure 4). Failure to predict complete soil water recharge at 30 and 60 cm at the end of the season was made by the other models as well and is possibly related to runoff (not considered here), since most rainfall produces some runoff from these basin slopes (W. Schlesinger, unpublished data, 1996). However, it is also likely that most rainfall during the late season is from low intensity frontal storms where the runoff is compensated for by runon. The underestimate of soil water recharge at 60 cm and the differences among models may be partly an artifact of the way in which the model output and data were gathered and averaged and which shows up when recharge occurs roughly to the depth of the measure-

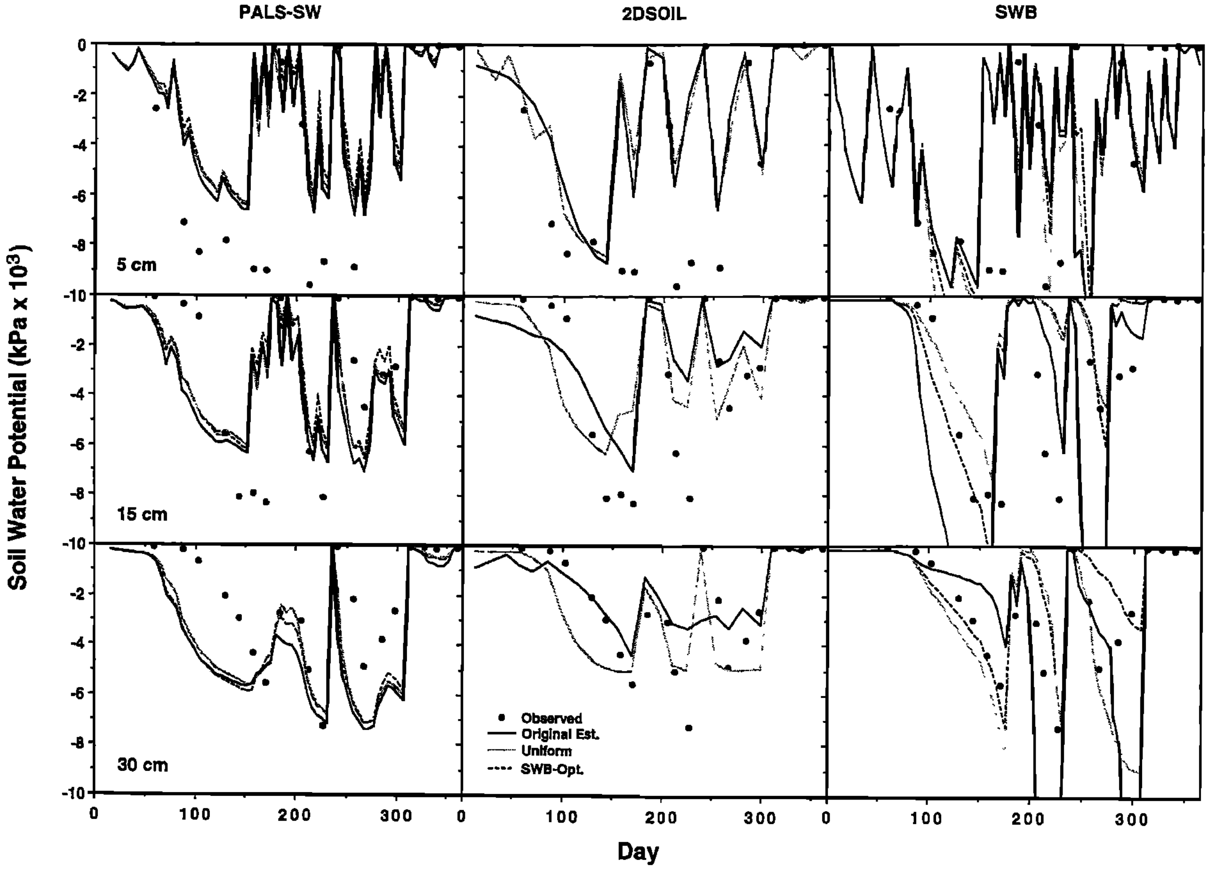

Figure 5. Effects of changing simulated root distributions upon predicted soil water potentials for station 50 at three depths using the three models. Points are observed water potentials.

ment node (in this case 60 cm). The observations represent an average water content over a roughly 30-cm thickness of soil that the neutron probe senses; the error bars associated with soil water at 60 cm at the end of the season are relatively large, suggesting that there was variation in actual recharge at that depth but that recharge was deep enough to be recorded by the neutron probe (at least 45 cm). The two soil water flux models report values for a small thickness of soil at that depth (5 cm). Since both of them essentially "missed" the year-end soil water recharge at 60 cm, it seems that the actual rainfall recharge must have penetrated no deeper than 55 cm (otherwise it would have been "recorded"). This suggests that recharge was to a depth of between 45 and 55 cm. This is substantiated by the output of the SWB model at 60 cm, which represents an average of 20 cm above and 20 cm below the 60 cm depth and which showed partial recharge of soil water at the 60 cm depth. Thus model predictions for soil moisture recharge during late summer and end of season may have less error associated with them than the RMSE terms or graphic comparisons would suggest.

Table 5. Root Mean Square Error Between Observed and Simulated Volumetric Soil Water Content for 25 Sample Dates at Three Depths and Five Different Stations for the Three Models

Model	Station 20	35	50	65	80	Mean
			30 cm			
PALS	0.051	0.035	0.022	0.030	0.027	0.033
2DSOIL	0.015	0.029	0.014	0.020	0.017	0.019
SWB	0.027	0.023	0.014	0.028	0.020	0.022
			60 cm			
PALS	0.025	0.031	0.033	0.021	0.013	0.025
2DSOIL	0.021	0.025	0.016	0.009	0.017	0.018
SWB	0.031	0.017	0.023	0.014	0.014	0.020
			90 cm			
PALS	0.021	0.012	0.013	0.008	0.007	0.012
2DSOIL	0.018	0.020	0.007	0.009	0.008	0.012
SWB	0.002	0.024	0.019	0.024	0.018	0.017
			Mean			
PALS	0.032	0.026	0.023	0.020	0.016	0.023
2DSOIL	0.018	0.025	0.012	0.013	0.014	0.016
SWB	0.020	0.021	0.019	0.022	0.017	0.020

Spatial variability. Results of the soil water distributions at different stations along the transect are shown in Figure 6. Although PALS-SW and 2DSOIL had similar errors associated with their predictions (Table 5), there were some consistent differences. PALS-SW generally predicted that soil water contents during spring (days 50–150) declined earlier at 30- and 60-cm depths than that which actually occurred. It also failed to adequately predict the complete amount of recharge in soil moisture that occurred at the end of the year at 30- and 60-cm depths. 2DSOIL fit the observed data quite well during this period. The differences in the models stemmed from differences in transpiration water loss (see below), with PALS-SW predicting greater transpiration. This result was unexpected, since the direct and exponential limitation of tran-

Table 6. Root Mean Square Error Between Observed and Simulated Soil Water Potentials for 21 Sample Dates at Three Depths and Five Different Stations for the Three Models

Model	Station 20	35	50	65	80	Mean
			30 cm			
PALS	2.7	3.17	3	2.81	3.57	3.050
2DSOIL	4.15	4.38	4.17	3.84	3.63	4.034
SWB	3.9	3.82	3.34	3.29	5.05	3.880
			60 cm			
PALS	2.34	3.31	2.75	3.31	2.92	2.926
2DSOIL	1.83	2.25	2.01	1.36	3.92	2.274
SWB	2.47	2.65	2.85	3.54	3.64	3.030
			90 cm			
PALS	2.39	2.39	2.3	2.58	2.29	2.390
2DSOIL	2.4	2.54	1.31	3.28	2.18	2.342
SWB	3.35	4.76	2.09	2.58	2.97	3.150
			Mean			
PALS	2.477	2.957	2.683	2.900	2.927	2.789
2DSOIL	2.793	3.057	2.497	2.827	3.243	2.883
SWB	3.240	3.743	2.760	3.137	3.887	3.353

Simulated soil water potentials: kPa × 10^2.

spiration by declining soil water, as well as limitation by vapor pressure gradients in PALS-SW, was expected to limit drying more than the linear response to soil water in 2DSOIL. The greater rate of transpiration in PALS-SW compared to 2DSOIL implies either that summing individual leaf transpiration rates for each life form in PALS-SW overestimates water loss via transpiration compared to the canopy-level transpiration limited by the surface energy budget as calculated in 2DSOIL or that in the spring the canopy transpiration is limited more by energy than by stomatal conductance [see *Schlesinger et al.*, 1990]. Whichever the case, the energy budget approach apparently yields more constrained estimates of transpiration and provided a somewhat better overall approximation of community water loss, particularly in spring.

The SWB model had slightly greater overall error in prediction of water contents than the other models (Tables 5 and 6). It failed to predict the general moisture depletion from the 60- and 90-cm depths in the fine textured soils (stations 20 and 35), and it predicted too much depletion, and shifted toward later in the year, from the coarse texture soils (stations 65 and 80). The differences in the SWB water content predictions at the lower soil depths compared to the other two models suggest that some of the error can be attributed to the lack of soil water redistribution in SWB. This is most likely the explanation for the fairly rapid change in water content at 60 and 90 cm early in the season. Water contents and resulting hydraulic conductivities were relatively high at these depths early in the season, which would have supported water redistribution toward the surface during this period.

Evapotranspiration

Predicted amounts of transpiration and soil evaporation along the transect are shown in Figure 7. With the exception of station 35, the models predicted that stations with high plant cover had high transpiration and low evaporation. Thus despite some quantitative differences in either evaporation or transpiration among the models, the ratio of transpiration to total evapotranspiration along the transect was similar among the models. While station 35 had the highest peak plant cover, it was predicted by PALS-SW and SWB to have low transpiration water loss, relative to other stations and relative to the evaporation loss; 2DSOIL, in contrast, predicted that transpiration at station 35 was second highest of all stations and higher than evaporative loss. All models predicted that station 65, with the lowest plant cover, had the lowest transpiration loss and highest evaporation loss.

In contrast to the relationship between evapotranspiration and plant cover, neither evaporation nor transpiration were strongly related to soil texture. For example, PALS-SW predicted similar evaporation and transpiration from stations 20 and 80, which had similar total plant cover, but different soil texture, 19% versus 8% clay, respectively, and PALS-SW predicted different total transpiration and evaporation for stations 20 and 35, which had similar soil textures although somewhat different cover. SWB predicted little difference in evaporation along the transect but predicted differences in transpiration that tended to be in concert with PALS-SW (relatively high transpiration for stations 20, 50, and 80, and low for stations 35 and 65) and thus not related to soil texture. The results of 2DSOIL were slightly more related to soil texture in that the two fine textured soils had highest transpiration rates and among the lowest evaporation rates.

Although we could not validate soil evaporation and transpiration predictions against actual measurements, we can indirectly evaluate the results on the basis of the degree to which the modeled soil water distributions fit the actual soil water patterns over the entire soil profile. A high degree of convergence over the entire profile suggests that the proportions of water extracted from the surface (primarily evaporation) versus deeper layers (primarily transpiration) is qualitatively correct. A second means of evaluating results is by comparing predictions among the models, which utilize different assumptions and formulations as to how water is extracted via soil evaporation and plant transpiration.

The three models differed little in their predictions of water distribution in the top 30 cm of the soil profile. This convergence was not due so much to similarity in the dynamics of the evaporation components per se, as to the fact that water is lost quickly from the surface, via liquid or vapor transport, and partly via transpiration. The SWB model predicted greatest evaporation of the three models, except for station 35, for which PALS-SW predicted a slighter higher evaporation. The high evaporation in the SWB model occurred despite the fact that evaporation was conservatively allowed only from the top 20 cm of soil. The simple relationship between evaporative loss and soil moisture content of the entire 10- or 20-cm surface layer may not adequately account for rapid surface drying and subsequent surface vapor barriers to water loss. This process would be accounted for in the mechanistic models, PALS-SW and 2DSOIL, which included soil surface nodes and water contents of the top 1 cm of soil. Although these two models were in general agreement with regard to evaporation, a significant departure was the prediction for station 35. Here PALS-SW predicted much greater evaporation than 2DSOIL. PAL-SW also predicted that evaporation for station 35 was much greater than for the similarly textured station 20; and concomitant transpiration much less. But because station 35 had greater plant cover than station 20, the opposite trend was expected. The only other obvious difference between the two stations was the lower initial moisture content in the top 60 cm of soil of station 35. Yet this would not account for the differ-

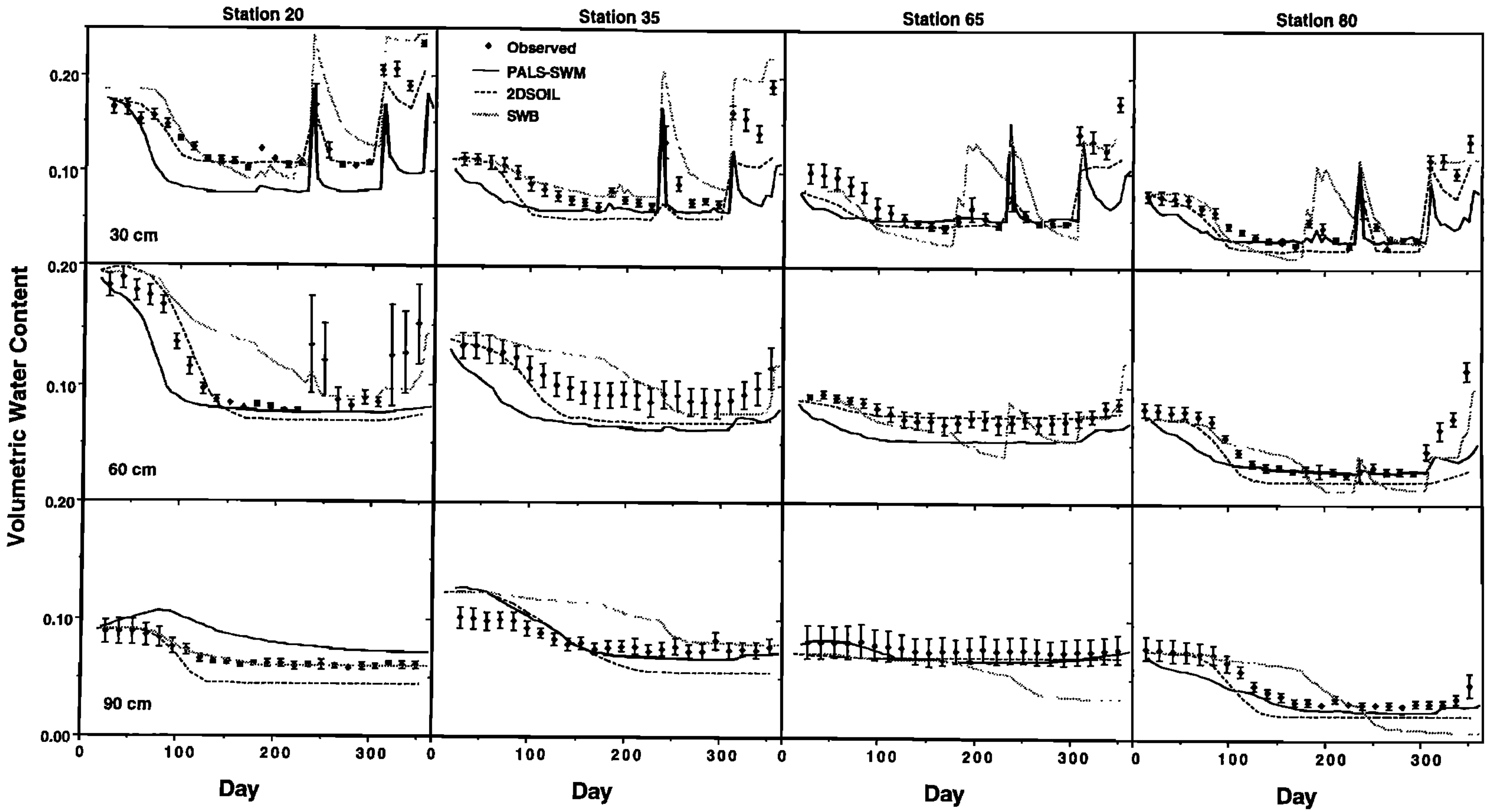

Figure 6. Predicted volumetric soil water contents at four stations along the transect and three depths using three models. Points are observed values.

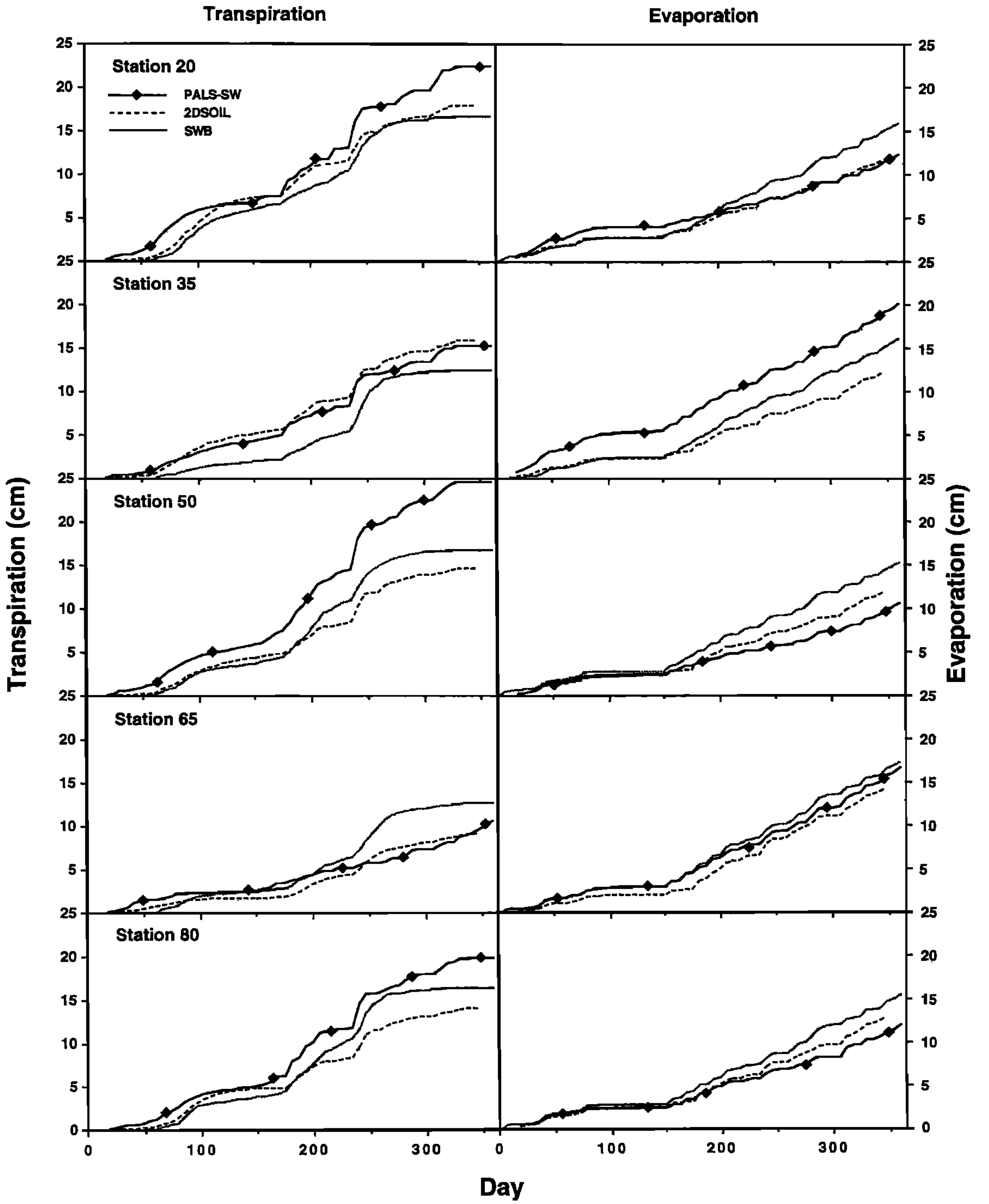

Figure 7. Predictions of transpiration and evaporation for five different stations along the transect by the three models.

ences in total evaporation either since the main departures in evaporation occurred in summer, long after the initial water content differences had disappeared (see Figure 7).

We suggest that the differences in evaporation relate to differences in transpiration between the stations, which, in turn, relate to the differences in life-forms. Station 20 had greater cover of subshrubs and lower cover of grass and annuals, compared to station 35. Rooting distributions for PALS-SW were such that subshrubs had 50% of their roots below 30 cm, whereas grasses had 30% of their roots below 30 cm, and annuals had none. Thus transpiration for station 20 resulted in greater fraction of water removal below 30 cm compared to station 35, as well as redistribution of surface water downward, with the net effect of relatively high transpiration and low evaporation. On the other hand, station 35, with greater rooting fraction in the surface soils, had reduced transpiration owing to the generally low water potentials of these surface soils. All models, in fact, predicted reduced transpiration for station 35, compared to station 20. Only PALS-SW predicted an increase in evaporation that offset the decline in

Table 7. Fraction of Total Annual Evapotranspiration Calculated as Soil Evaporation Versus Plant Transpiration for the Different Models and Different Stations

	PALS-SW			2DSOIL			SWB		
Station	Transpiration	Evaporation	*T*/ET	Transpiration	Evaporation	*T*/ET	Transpiration	Evaporation	*T*/ET
20	22.6	12.8	0.64	18.0	12.9	0.58	16.7	16.4	0.50
35	15.4	20.8	0.43	16.0	13.4	0.54	12.5	16.5	0.43
50	25.2	11.0	0.70	14.7	13.1	0.53	16.8	15.8	0.52
65	10.5	17.3	0.38	9.2	15.5	0.37	12.7	17.8	0.42
80	19.8	12.4	0.61	14.1	14.1	0.50	16.4	15.8	0.51
Mean	20.0	14.9	0.57	14.4	13.8	0.51	15.0	16.5	0.48

Evaporation and transpiration are the cumulative values for the year 1986. *T*/ET is the ratio of cumulative transpiration divided by cumulative ET.

transpiration; this was apparently because summer showers remained in the surface soils since there was not a strong subsurface transpiration sink to foster downward redistribution.

A lack of knowledge of the relative contributions of transpiration and evaporation to soil water loss in deserts greatly limits our general understanding of the interaction of plant, soil, and climate factors in controlling water loss in arid environments and thus our ability to predict soil water distributions [*Bailey*, 1981; *Evans et al.*, 1981; *Thames and Evans*, 1981; *Milton et al.*, 1994]. Empirical studies have produced rather conflicting results concerning the contributions of transpiration and evaporation to the loss of water from desert ecosystems. *Evans et al.* [1981] summarized ET studies in several desert plant communities and concluded that transpiration contributed very little to total soil water loss, except for communities that received the majority of precipitation during cold winters, which resulted in deep soil moisture recharge. *Sammis and Gay* [1979] found that only about 7% of the water lost from a Sonoran Desert creosote bush community was from transpiration. *Ross* [1977] also concluded that most water was lost as evaporation from the soil surface in arid Australian communities. However, others have concluded that transpiration represented a substantial proportion of the water loss in arid systems. *Lane et al.* [1984] found that transpiration accounted for 27% of the soil moisture loss at a site in the Mojave Desert with a low plant cover (25%). *Caldwell et al.* [1977] reported that shrub-dominated communities in the Great Basin Desert lost about equal amounts of soil water through transpiration and soil evaporation. *Schlesinger et al.* [1987] found transpiration accounted for the majority (72%) of water loss from a creosote bush/snakeweed community in the Chihuahuan Desert, while *Liu et al.* [1995] concluded that transpiration accounted for nearly 80% of the water lost from desert communities of southern Arizona. These divergent results underscore the need for higher-resolution studies of water distribution and loss in deserts.

Our results for evapotranspiration at station 65, which consisted of relatively low cover (30%) of *Larrea tridentata* and subshrubs, provide a direct comparison to the studies of *Sammis and Gay* [1979] and *Schlesinger et al.* [1987]. The three models predicted that the transpiration water loss for this community was about 40% of the total water loss (Table 7), very nearly midway between the 7% and 72% found in the above studies, respectively. The effect of plant cover is seen by comparing the results of station 65 with station 80, which had similarly coarse textured soil but about twice the plant cover (and with grass and annuals as dominants). The predicted transpiration fraction of water loss for this station was between 51% (SWB) and 61% (PALS-SW) of the total (Table 7). Experiments that study evaporation and transpiration in "isolation," such as measuring soil evaporation in a lysimeter devoid of roots [e.g., *Sammis and Gay*, 1979], are likely to reach erroneous conclusions because of the strong interdependency between these processes.

Our results emphasize the degree to which water is quickly lost from this desert system irrespective of the extent of plant cover, plant growth forms, or soil texture. Nevertheless, we found some differences along the transect. Transpiration was lowest (40% of total ET) for the creosote bush shrub community, which had the lowest plant cover (30% peak cover). The fraction of ET as transpiration increased with increasing plant cover: 2DSOIL predicted the highest transpiration (58% of total ET) for the mixed vegetation community (60% peak cover) on relatively fine textured soil, and PALS-SW predicted the highest transpiration (69% of total ET) for the mixed vegetation community (70% peak cover) on relatively coarse textured soils. The community type was found to have an effect on the amount of water lost as transpiration primarily via cover and secondarily via depth and distribution of roots. In this respect, PALS-SW predicted greatest differences among stations in terms of specific differences in plant community types. Since PALS-SW did not provide as good of a fit with the soil moisture data as did 2DSOIL, the differences in the morphology and physiology of the life-forms may be secondary to the overall control of water loss by the primary factors accounted for in 2DSOIL: vertical distribution of soil moisture, degree of canopy cover, and evaporative energy budget of the canopy. Soil texture did not play an important role in independently controlling soil evaporation but rather interacted with the amount and type of plant cover to affect both evaporation and transpiration.

Finally, although the models differed substantially in assumptions and methodology, they all predicted soil water distributions reasonably well. There also was general agreement among the models with regard to the relative contributions of soil evaporation and transpiration to soil water loss.

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J.-L. Chen, P. R. Kemp, and J. F. Reynolds, Department of Botany and Phytotron, Duke University, P.O. Box 90340, Durham, NC 27708-0340.

Y. Pachepsky, Remote Sensing and Modeling Laboratory, Bldg. 011A, Rm. 165B, BARC-W, Beltsville, MD 20705.

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