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Effect of the number of soil layers on a modeled surface water budget

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Abstract. A sensitivity analysis was performed to determine the effects of systematically increasing the number of soil layers in a land surface-atmosphere model on the components of the modeled water budget. The study was done for a forested location in central Oklahoma for a 65-day period in spring 1996 using the model called Simulator for Hydrology and Energy Exchange at the Land Surface (SHEELS). SHEELS is based on the Biosphere-Atmosphere Transfer Scheme (BATS), except that the subsurface hydrology was substantially changed to improve representation of the soil moisture profile. The soil profile was divided into zones of thickness 0.05 m (upper), 1.25 m (root), and 1.20 m (bottom). The two principal conclusions are that (1) the water budget is very sensitive to the number of layers in the soil profile under wet conditions and (2) the water budget is much more sensitive to the number of layers in the profile than to the range of 2 orders of magnitude in saturated hydraulic conductivity considered in this study. A result of the latter conclusion is that larger errors in modeled water fluxes can occur from using an insufficient number of soil layers than from using an incorrect value of saturated hydraulic conductivity.

1. Introduction

Because of the natural variability of the hydrologic properties of soil, land-atmosphere flux models vary widely not only in the representation of physical processes but in the detail in which soil properties are specified. Simple energy balance models use only one or two soil layers [*Sorooshian et al.*, 1993; *Hughes and Sami*, 1994; *Liang et al.*, 1994; *Lakshmi and Wood*, 1998]. In the two-layer model structure described by *Lakshmi et al.* [1997] the upper layer was chosen to be 1 cm thick to facilitate comparisons to satellite-observed soil moisture. *Boone et al.* [1999] studied certain limitations of a two-layer profile and showed that the addition of a third layer allowed root zone recharge during the drying season. Some models use as many as 20 layers with uniform layer thickness throughout the column [*Smith et al.*, 1994], while others allow use of a variable number of soil layers, with layer thickness specified by a geometric series [*Abramopoulos et al.*, 1988; *Viterbo and Beljaars*, 1995; *Wetzel and Boone*, 1995]. Since hydrologic properties can vary significantly throughout the vertical soil profile, specifying a large number of soil layers, rather than a small number, should yield, in general, an improved representation of the soil moisture profile and thus the vertical water flux. *Mahrt and Pan* [1984] performed a comparison between their simple two-layer soil hydrology model and a 100-layer model. They found that with respect to the high-resolution model the two-layer model significantly underestimated water flux between the layers owing to errors in specifying the vertical gradient of moisture and associated hydrologic properties. It was observed that the impact of these errors on water flux can be decreased by defining a sufficiently thin upper layer and by using the hydraulic conductivity corresponding to the water content of the wetter layer. *Koren et al.* [1999] analyzed output from 2-, 4- and 10-layer versions of the model of *Mahrt and Pan* [1984] and found that the number of layers had a substantial influence on evapotranspiration, runoff, and soil moisture change. While, in principle, soil water flux can be calculated for any number of soil layers, commensurate knowledge of the variation of the actual soil properties with depth is usually lacking. Therefore a relatively coarse accounting of the hydrologic soil properties must be invoked, as is the case in our investigation. Accordingly, the results of this and similar investigations must be viewed in this context.

The objective of this study was to determine the effects of varying the number of soil layers in a land surface-atmosphere model on the modeled water balance and associated water flux processes. The objective was accomplished through a sensitivity analysis of selected modeled water flux variables. Specifically, surface evaporation, transpiration, ponding, infiltration, and change in soil water storage were evaluated for a forested location in the Blue River Basin in south central Oklahoma over approximately a 2-month period in spring 1996. The approach we used was to vary the number of soil layers in two distinct profiles of soil properties on the basis of the highest and lowest values of saturated hydraulic conductivity in the basin. Because there are no observations of energy or water fluxes at this site, no adjustment or tuning of model parameters was performed. Even if observations had been available, model tuning is not necessary for the purpose of analyzing sensitivity to the number of soil layers.

The numerical model used for the analysis is called Simula-

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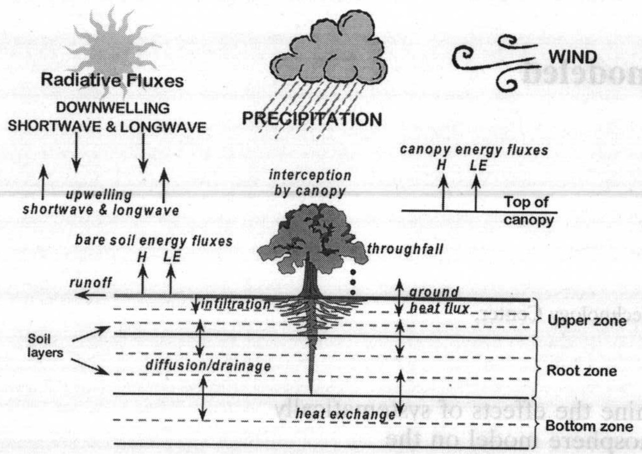

Figure 1. Model schematic for Simulator for Hydrology and Energy Exchange at the Land Surface (SHEELS). Model forcing variables are denoted by capital letters; model diagnosed variables are denoted by lowercase italicized letters. Zone thicknesses are not to scale.

tor for Hydrology and Energy Exchange at the Land Surface (SHEELS) [*Crosson et al.*, 2001]. SHEELS evolved from an earlier version called Experimental Biosphere-Atmosphere Transfer Scheme (Ex-BATS) described by *Smith et al.* [1993], which, in turn, was based on the original BATS model [*Dickinson et al.*, 1986, 1993]. The principle differences between SHEELS and earlier versions lie in the modeling of soil hydrology. These differences will be described in section 2.

A description of SHEELS and the model parameters is given in section 2. Data from the basin that were used as input into SHEELS are described in section 3. The sensitivity analysis is discussed in section 4, followed by a summary and conclusions in section 5. In Appendix A we explain the cause of an anomalous behavior in the change in soil water storage noted in section 4.

2. Description of SHEELS

The physics of SHEELS is based on that of BATS and Ex-BATS. Although the vegetation and surface flux schemes of the earlier models have been retained in SHEELS, the subsurface hydrology in SHEELS differs substantially. Major modifications have been made to improve the representation of the soil moisture profile. In particular, we have implemented the soil water flux algorithm developed by *Capehart and Carlson* [1994]. There are three nested soil zones in BATS and Ex-BATS. In SHEELS, on the other hand, there are three contiguous soil zones (upper, root, and bottom), each varying in thickness and each comprising a user-specified number of layers (Figure 1). The flexibility in number and thickness of layers can potentially improve the vertical soil moisture profiles, in turn leading to more accurate partitioning of the net radiation into its latent and sensible heating components. Soil properties used as input into SHEELS are specified only for each of the three soil zones, not individual layers. However, the energy and water fluxes are diagnosed with the vertical exchange of water occurring between adjacent soil layers due to gradients in water content, soil suction, and hydraulic conductivity. The temporal change in soil moisture content in each layer is determined by considering surface ponding, infiltration, runoff, evapotranspiration, vertical diffusion, and gravitational drainage.

The diagnosis of evapotranspiration in SHEELS is based on the concept of supply-limited and demand-limited evapotranspiration (ET) [*Federer*, 1982]. In applying this method, a maximum transpiration rate, which is a function of the upper and root zone water contents, defines the limiting value of the supply. A potential ET rate, formulated in terms of a conductance-weighted humidity gradient involving the ground, the canopy, and the above-canopy air, defines the demand. Actual ET is related to potential ET through a dimensionless fractional quantity that depends on aerodynamic and stomatal resistances. If transpiration exceeds the maximum rate, it is decreased and the resistance terms are adjusted accordingly.

The upper boundary conditions for the model simulations require seven atmospheric forcing variables: air temperature, relative humidity, surface pressure, downwelling shortwave and longwave radiation, wind speed, and precipitation. SHEELS then estimates several surface energy fluxes and water transport variables including sensible, latent and ground heat fluxes, surface runoff, and infiltration. Figure 1 shows a schematic of these forcing variables and the energy and water exchange processes that occur in SHEELS.

The thickness of each zone in SHEELS must be specified. The 0.05 m depth of the upper zone was chosen as an approximation to the shallow layer that, under strong drying conditions, can decouple from the deeper soil [*Santanello and Carlson*, 2001]. Results of experiments (to be discussed in section 4.1) showed little sensitivity to the number of layers in the upper zone for this shallow depth. Rooting depths for the various soil types found in the Blue River Basin were determined through an examination of the individual soil series descriptions. For each series the depth of the deepest layer containing fine roots was found, and the mean rooting depth for the basin was calculated to be 1.3 m, which is consistent with rooting depths for forests [*Dickinson et al.*, 1993]. A lower boundary condition was chosen such that we could treat the base of the soil column as an impermeable boundary across which there are no water losses. Thus an investigation was made to find the depth at which vertical soil moisture flux was essentially zero over the 2-month period. The appropriate depth of the total soil column was found to be 2.5 m.

3. Model Input Data

The sensitivity analysis in section 4 was performed in the Blue River Basin located in south central Oklahoma. An evaluation of the time series of rainfall and solar radiation in the basin during spring 1996 revealed several significant rain events with intervening periods of dry down. On the basis of the evaluation, a 65-day time period, from March 1 (day 61) through May 4 (day 125), was chosen for our investigation. An hourly time step was used for the model simulations. The location chosen for the soil layer sensitivity analysis is forested, with vegetation comprising deciduous trees, conifers, and shrubs, and received the greatest amount of rain (253 mm) in the basin over the 65-day period. This combination produces large surface energy and soil water fluxes. Although SHEELS is designed for use on a spatial grid, there is no interaction between grid points in the current application, so that selection of a single grid point is appropriate for this analysis.

Detailed sets of meteorological and soil property data from the Blue River Basin were used as input data to SHEELS.

Table 1. Soil Properties in the Upper Zone Associated With the Highest and Lowest Values of Saturated Hydraulic Conductivity Found in the Blue River Basin[a]

Soil Property	Units	Highest K_{sat}	Lowest K_{sat}
Soil texture class		loamy sand	clay loam
Saturated hydraulic conductivity K_{sat}	mm s^{-1}	4.2×10^{-2}	4.0×10^{-4}
Dry soil density	kg m^{-3}	1400	1325
Soil porosity	unitless	0.47	0.50
Saturated soil suction	mm	28	209
Clapp-Hornberger b parameter	unitless	4.4	5.4
Wilting point	unitless	0.06	0.10
Saturated soil albedo for visible radiation	unitless	0.11	0.14

[a]Upper zone thickness is 0.05 m; root zone thickness is 1.25 m; and bottom zone thickness is 1.20 m. Clapp-Hornberger b parameter is from *Clapp and Hornberger* [1978].

Meteorological and radiative flux data were obtained from three Oklahoma Mesonet [*Brock et al.*, 1995] sites (Sulphur, Tishomingo, and Durant) located just to the west of the basin. The Mesonet variables used in this evaluation include atmospheric pressure, air temperature and relative humidity measured at 1.5 m above the surface, wind speed at 2.0 m, and downwelling shortwave (solar) radiation. Hourly means of each variable were then averaged over the three stations to provide spatially uniform variables across the basin at each time step.

Compared to other meteorological variables, precipitation amounts can vary widely across a region. Therefore we decided to obtain rainfall data for the basin from spatially distributed Next Generation Weather Radar (NEXRAD) Stage III radar-estimated rainfall maps acquired from the Arkansas-Red Basin River Forecast Center. Thus the rainfall applied at the study grid point is representative of the local conditions, in contrast to the other meteorological variables that represent basin averages.

Hourly daytime downwelling longwave radiation was estimated outside SHEELS using the formula given by *Crawford and Duchon* [1999], which requires air temperature, water vapor pressure, and cloud fraction. This formula is an improved version of the physically based formula derived by *Brutsaert* [1975] for clear skies. Cloud fraction was defined as the ratio of measured downwelling shortwave irradiance to the clear-sky irradiance. Nighttime downwelling longwave radiation was similarly calculated except that hourly cloud fraction was estimated by linearly interpolating between cloud fraction at sunset and sunrise. In situations where air temperature increased or rainfall occurred, interpolated cloud fraction was increased as appropriate.

The slope of the land surface is normally calculated from elevation data input to SHEELS. However, in this study, the local slope angle was set to zero, eliminating surface runoff and thereby maximizing water ponding and infiltration and accentuating the exchange of water between soil layers. The absence of surface runoff and the inclusion of an impermeable lower boundary yield a simple water budget equation in which the change in water storage for the entire soil column equals precipitation minus evapotranspiration.

Several soil properties were obtained from the Natural Resources Conservation Service (NRCS) Map Unit Interpretation Record (MUIR) database. (Data for individual counties are available from http://www.statlab.iastate.edu/soils/muir.) The original source of the MUIR data was county-level soil surveys [e.g., *Watterson et al.*, 1984]. The MUIR soil properties include clay and sand fractions that were used to define soil texture classes on the basis of NRCS standards [*Soil Survey Staff*, 1999] and to approximate effective porosity, wetting front suction, and saturated hydraulic conductivity on the basis of the *Rawls and Brakensiek* [1985] parameterizations of the *Brooks and Corey* [1964] soil water retention variables. Bulk density was used to calculate total porosity and wilting point. Clay fractions of the upper zone also were used to estimate the *Clapp and Hornberger* [1978] parameter and saturated ground albedo at visible wavelengths. Additional soil properties for each zone were estimated from published values [*Dickinson et al.*, 1993] for the corresponding texture class.

The magnitude of saturated hydraulic conductivity varies by ~2–3 orders of magnitude from sand to clay soils [*Rawls et al.*, 1982; *Dickinson et al.*, 1993]. As indicated in section 1, the highest and lowest values of saturated hydraulic conductivity present in the various soils of the Blue River Basin were employed in this sensitivity analysis. We then associated other properties, such as porosity, bulk density, and saturated suction, with the extremes of saturated hydraulic conductivity to obtain two distinct sets of soil parameters. The soil properties are shown in Table 1, where it is seen that the saturated hydraulic conductivity in the upper zone varied by a factor of 100 between the two soils, which we refer to as loamy sand (high conductivity) and clay loam (low conductivity) soils. Both profiles of soil properties are consistent with the types of soils found within forested areas in the basin.

Another set of spatially varying input data used in SHEELS includes vegetation properties. Land cover data were obtained from county-level surveys that used aerial photographs to identify 99 classes of vegetation. In SHEELS these vegetation types were then grouped into seven categories. Vegetation properties such as leaf area index (4.5), canopy height (5.0 m), fractional vegetation cover (0.6), and stomatal resistance (minimum 150 s m^{-1}, maximum 5000 s m^{-1}), were defined on the basis of values used in the BATS model [*Dickinson et al.*, 1993].

An important consideration when utilizing a land surface flux model such as SHEELS is the initial value of soil moisture. Prior to the evaluation period in 1996 the area of study had experienced a drought for several months. Because of the dry soil conditions the upper zone soil moisture was initialized near the wilting point (volumetric water content with respect to saturation of 0.21) for both sets of soil properties, while the root and bottom zones were initialized at moderately dry moisture content (volumetric water content of 0.25 and 0.40, respectively).

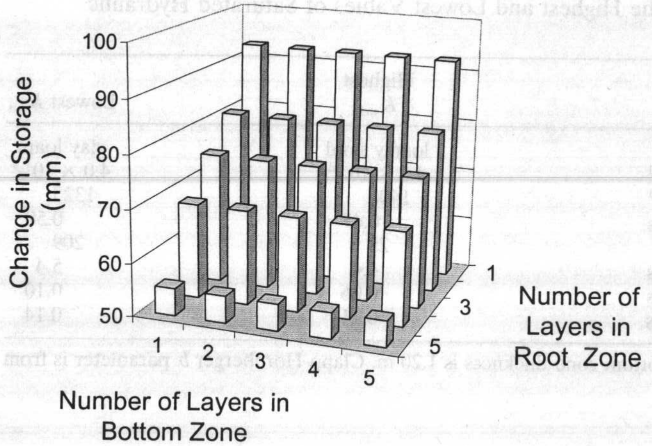

Figure 2. Change in (final minus initial) total soil water storage for the study period (March 1 through May 4, 1996) for the high saturated hydraulic conductivity case. The number of layers in the upper zone was held constant at one.

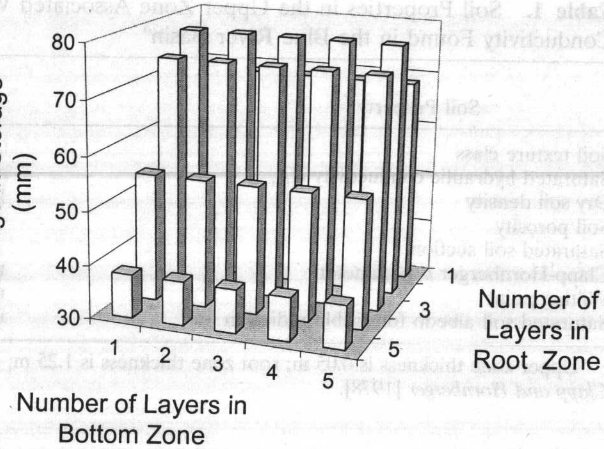

Figure 3. Same as Figure 2 except for the low saturated hydraulic conductivity case.

4. Sensitivity Analysis

4.1. Varying the Number of Layers in Each Soil Zone From One to Five

The relationship between the number of soil layers and the change in soil water storage was determined by performing a series of model runs for the high- and low-saturated hydraulic conductivity cases over the 65-day evaluation period and summing the various components of the water budget. These simulations were made by varying the number of layers in the upper (Nu), root (Nr), and bottom (Nb) zones from one to five. Figures 2 and 3 show the results for the high and low saturated hydraulic conductivities, respectively (hereinafter referred to as high-K and low-K), in which Nu was held constant at one and Nr and Nb varied from one to five. The change in storage is defined as the value of soil water storage at the end of the 65-day period minus the initial value. Figures 2 and 3 illustrate that in general, increasing the number of soil layers in the root zone results in a significant decrease in the change in total soil water storage. That is, the greater the number of root zone layers, the drier the soil column at the end of the study period. The exception to this generalization occurs in Figure 3, the low-K case. Here we see that the soil water change increases from Nr = 1 to Nr = 2 and then decreases as Nr is increased to 5. The source of this unexpected behavior is discussed in Appendix A. Suffice it to say here that the behavior is a consequence of interactions among soil water flux, ponding, and evaporation. Common to both Figures 2 and 3 is that an increase in the number of layers in the bottom zone has a very small effect on the change in total soil column water storage. This is because only a tiny fraction of the water transferred from the bottom zone to the root zone is consumed in evaporation and transpiration during the study period; thus the change in total soil water storage remains virtually the same.

Although not shown here, results similar to those shown in Figures 2 and 3 were obtained when the upper zone was divided into two to five layers, while the number of layers in the root and bottom zones was varied from one to five. From these results we concluded that an upper zone thickness of 5 cm and a bottom zone thickness of 120 cm, each comprising only one layer, are sufficient for use with SHEELS for the hydrometeorological conditions of this study. This conclusion is valid over the fairly wide range of soil properties used.

Selected components of the water budget for the 65-day period are shown in Table 2 for the high-K case and Table 3 for

Table 2. Selected Water Budget Variables for Different Numbers of Layers in the Root and Bottom Zones for the High Saturated Hydraulic Conductivity Case[a]

Number of Layers in Root Zone, Nr	Number of Layers in Bottom Zone, Nb	Total Evapo-transpiration	Canopy Evapo-transpiration	Ground Evaporation	Upper Zone Δwater	Root Zone Δwater	Bottom Zone Δwater	Total Column Δwater
1	1	160.3	123.8	36.4	−0.1	93.1	−0.1	93.0
1	5	160.5	123.9	36.6	−0.1	95.2	−2.4	92.7
2	1	172.5	123.3	49.2	0.8	81.2	−1.2	80.8
2	5	172.6	123.4	49.2	0.8	85.6	−5.7	80.7
3	1	178.4	123.0	55.4	0.9	74.9	−1.0	74.9
3	5	178.5	123.2	55.4	0.9	79.5	−5.7	74.7
4	1	185.2	122.5	62.7	1.6	67.3	−0.9	68.0
4	5	185.5	122.7	62.7	1.6	71.8	−5.6	67.8
5	1	198.1	121.8	76.3	1.6	54.4	−0.9	55.2
5	5	198.3	122.1	76.3	1.6	58.8	−5.5	54.9

[a]Number of layers in the upper zone was held constant at 1. Study period is March 1 through May 4, 1996. Change in water in a soil zone (Δwater) is defined as the water depth at the end of the time period minus the water depth at the beginning. Water budget variables have units of mm.

Table 3. Selected Water Budget Variables for Different Numbers of Layers in the Root and Bottom Zones for the Low Saturated Hydraulic Conductivity Case[a]

Number of Layers in Root Zone, Nr	Number of Layers in Bottom Zone, Nb	Total Evapo-transpiration	Canopy Evapo-transpiration	Ground Evaporation	Upper Zone Δwater	Root Zone Δwater	Bottom Zone Δwater	Total Column Δwater
1	1	184.6	112.2	72.4	2.1	66.6	0.0	68.7
1	5	184.7	112.3	72.4	2.1	66.7	−0.2	68.6
2	1	175.7	112.9	62.8	2.3	75.3	0.0	77.6
2	5	175.7	112.9	62.8	2.3	75.5	−0.3	77.6
3	1	179.4	112.8	66.6	2.7	71.2	0.0	73.8
3	5	179.5	112.8	66.6	2.7	71.4	−0.3	73.8
4	1	199.1	111.7	87.5	3.4	50.7	0.0	54.1
4	5	199.2	111.7	87.5	3.4	50.9	−0.2	54.1
5	1	215.1	111.0	104.1	3.9	34.4	0.0	38.2
5	5	215.1	111.0	104.0	3.9	34.5	−0.2	38.2

[a]Number of layers in the upper zone was held constant at 1. Study period is March 1 through May 4, 1996. Change in water in a soil zone (Δwater) is defined as the water depth at the end of the time period minus the water depth at the beginning. Water budget variables have units of mm.

the low-K case. In both Tables 2 and 3 the number of layers in the upper zone was held constant at one. Two conclusions can be immediately drawn from these results. The first is that the decrease in the change in total soil water storage (rightmost column in Tables 2 and 3) as the number of root zone layers increases is attributable mostly to the decrease in the change in root zone storage. The dominance of the root zone is not surprising because its thickness is 25 times that of the upper zone.

The second conclusion is that as Nr increases, the decrease in root zone water gain over the study period is nearly equal to the increase in ground evaporation. Canopy evapotranspiration remains nearly unchanged. The increase in ground evaporation is 40 mm in the high-K case (Table 2) and ~32 mm in the low-K case (Table 3). The increase in ground evaporation as the number of root zone layers increases is the result of an increase in evaporation from ponded water and saturated soil. The physical explanation is given in section 4.3.

Figures 2 and 3 show that soil water storage change has not yet stabilized with five root zone layers. Thus we decided to perform additional simulations with up to 15 layers in the root zone and a single layer in the upper and bottom zones.

4.2. Varying the Number of Root Zone Layers From 1 to 15

The relationship between the change in total soil water storage and the number of layers in the root zone for the high-K case is shown in Figure 4. Figure 4 demonstrates that a large decrease in the change in storage (from 93 to 81 mm) occurs as the number of root zone layers is increased from one to two. The change in water storage continues to decrease substantially until five layers in the root zone are used, after which there is a much slower decrease in storage gain. For Nr > 9 the decrease is insignificant. As dictated by the water balance equation, the cause of the decrease in the change in storage is the increase in evapotranspiration. Although not shown here, as the number of layers in the root zone is increased from 1 to 15, ET increases from 160 to 210 mm, resulting in a decrease in total soil water storage from 93 to 43 mm. As shown in Table 2, almost all the increase in ET is due, in fact, to an increase in ground evaporation.

The sensitivity of the modeled water storage to changes in the number of root zone layers for the low-K case is shown in Figure 5. Again, the model simulations were performed by setting Nu = Nb = 1, while varying Nr from 1 to 15. Except for the anomalous behavior from Nr = 1 to Nr = 2 (the cause of

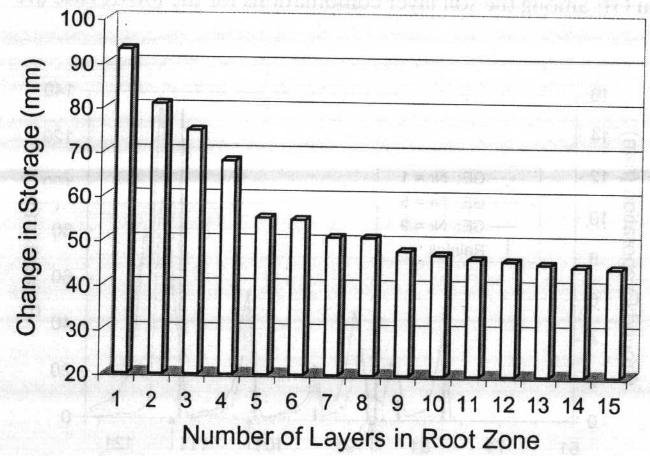

Figure 4. Change in total soil column water storage for the study period for the high saturated hydraulic conductivity case. The number of layers in the upper and bottom zones was held constant at one.

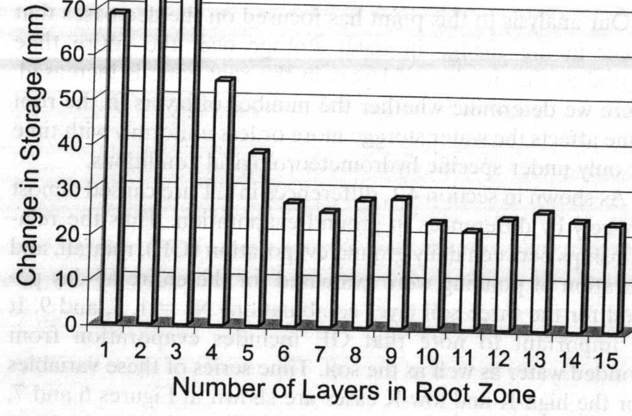

Figure 5. Same as Figure 4 except for the low saturated hydraulic conductivity case.

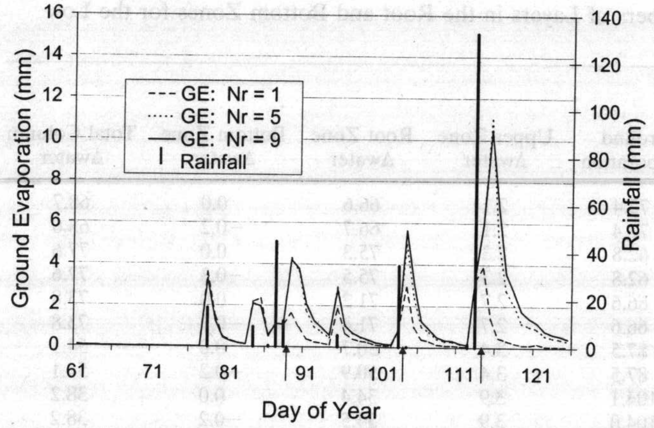

Figure 6. Daily ground evaporation (GE) for three soil layer combinations and daily rainfall for the high saturated hydraulic conductivity case. An arrow indicates the beginning of a period of ponding. The number of layers in the upper and bottom zones was held constant at one. The number of layers in the root zone (Nr) was one, five, and nine.

which is explained in Appendix A), there is a systematic, substantial decrease in the change in soil water storage through Nr = 6, followed by an oscillatory pattern of a few millimeters amplitude out to Nr = 15. We consider the differences beyond Nr = 10 to be insignificant. Although not shown here, the changes in soil water storage were also evaluated from Nr = 16 to Nr = 30, with the result that these oscillations damp out and converge toward a value of ~25 mm.

The most important conclusion from Figures 4 and 5 is that the components of the water budget are much more sensitive to the number of root zone layers than to saturated hydraulic conductivity. The basis for this conclusion can be explained as follows. The difference between the maximum and minimum change in storage for the two soil profiles (Figures 4 and 5) is in the neighborhood of 45 to 50 mm, which reflects the sensitivity to the number of soil layers. In comparison, the difference in the change in storage between the high-K and low-K cases, over which saturated hydraulic conductivity varies by 2 orders of magnitude, is ~20 mm for Nr ≥ 9 (equilibrium conditions). Thus knowledge of the exact saturated hydraulic conductivity is far less important than selection of an appropriate number of soil layers.

4.3. Time Series of Ground Evaporation and Rainfall

Our analysis to this point has focused on the decreases that occur in the change in water storage over the entire time period as the number of layers in the root zone is increased. Here we determine whether the number of layers in the root zone affects the water storage more or less uniformly with time or only under specific hydrometeorological conditions.

As shown in section 4.1, differences in ET are caused almost entirely by differences in ground evaporation. Thus the relationships between daily ground evaporation (GE), rainfall, and duration of ponding were evaluated for the entire 65-day period for the three soil layer combinations Nr = 1, 5, and 9. It is important to note that GE includes evaporation from ponded water as well as the soil. Time series of these variables for the high-K and low-K cases are shown in Figures 6 and 7, respectively.

The first observation from Figures 6 and 7 is that, in general, increases in GE occur as the number of root zone layers is increased. The explanation for these increases is that when a large number of root zone layers are used, the upper layers in the root zone become nearly saturated following a significant rain event. The upper zone stays close to saturation as long as it is in near equilibrium with the upper portion of the root zone, resulting in high GE. On the other hand, when Nr = 1, the thick root zone (1.25 m) never approaches saturation, and there is fairly rapid drainage into it from the upper zone. Consequently, the upper zone does not remain wet for as long a period as when Nr is large. Therefore, even when the soil properties for each individual layer are not known, simply changing the number of soil layers within a zone can significantly alter the soil moisture profile.

The second observation, also common to both Figures 6 and 7, is that large differences among GE for Nr = 1, 5, and 9 usually occur after a rain event, the largest of which follows the 135 mm rainfall on day 113. These differences can be understood by taking into account ponding of surface water. Ponding occurs when the rainfall rate exceeds the infiltration rate of the soil. Usually, as Nr increases, the ponding duration increases since the upper zone remains wet for a longer period. This results in a greater evaporation of ponded water. In addition, ponding duration is strongly affected by hydraulic conductivity. For the high saturated hydraulic conductivity soil, ponding lasted 15 hours for Nr = 1 and 25 hours for Nr = 9 after the heavy rain event on day 113. In comparison, for the low saturated hydraulic conductivity soil, ponding lasted 3 days for Nr = 1 and longer than 4 days for Nr = 9.

For a 6-day period that included the heavy rain event on day 113, we compared evaporation from ponded water with evaporation from soil for both the high-K and low-K cases. For the high-K case, evaporation directly from the ponded water was 1.1 mm for Nr = 1 and 3.2 mm for Nr = 9. After the ponded water evaporated, evaporation from the soil was 8.1 mm for Nr = 1 and 27.0 mm for Nr = 9. These values clearly show that for the high-K case most of the differences in GE among the soil layer combinations are caused by differences in evaporation from the soil, not from ponded water. For the low-K case, evaporation directly from ponded water was 20.4 mm for Nr = 1 and 41.6 mm for Nr = 9. Evaporation of water from the soil after ponding ended was 11.8 mm for Nr = 1 and 3.1 mm for Nr = 9. In contrast to the high-K case, most of the differences in GE among the soil layer combinations for the low-K case are

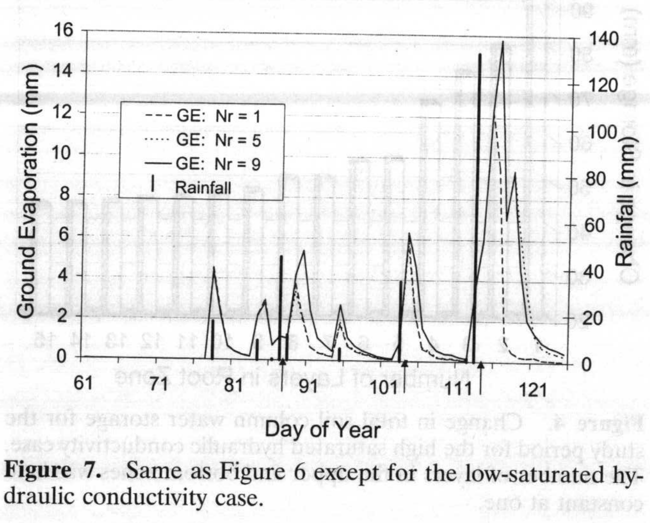

Figure 7. Same as Figure 6 except for the low-saturated hydraulic conductivity case.

due to differences in evaporation of ponded water. Thus while GE increases with an increasing number of soil layers in the root zone, the dominant source of the evaporated water depends on the magnitude of saturated hydraulic conductivity.

5. Summary and Conclusions

The objective of this study was to determine the effects of varying the number of soil layers on modeled water balance variables in SHEELS, a surface energy flux model. We performed a sensitivity analysis over a 65-day period in south central Oklahoma for two distinct soil property profiles: one characterized by a high-saturated hydraulic conductivity, and the other characterized by a low-saturated hydraulic conductivity. The analysis was performed for a forested grid point. Because most of the differences in the water budget were attributed to differences in ground evaporation (GE), we expect similar results for other vegetated surfaces. The soil, vegetation, and meteorological inputs in this study were representative of the local scale (<10 km^2); therefore our conclusions may not apply to soil hydrology components of large-scale models operating on grid cells of the order of 10^3–10^4 km^2.

Our first step was to examine the sensitivity of the magnitude of change in soil water storage to the number of layers in the upper, root, and bottom soil zones. We found that a single layer in both the upper and bottom zones was adequate. We then evaluated time series of daily GE and rainfall to determine which hydrometeorological conditions have the greatest effect on GE as the number of root zone layers is increased. We also investigated the cause of the anomalous behavior in soil water storage that occurred for the low saturated hydraulic conductivity case. Understanding the anomaly is not central to determining our objective but is noteworthy in its own right. Thus the discussion of the anomaly is presented in Appendix A.

We draw the following conclusions from our sensitivity analysis:

1. The water budget components are very sensitive to the number of layers in the soil profile (see Figures 4 and 5) under wet soil conditions. The components of the water budgets were found to stabilize for ~9 or 10 root zone layers, depending on the soil properties. We believe that water flux estimates from models with similar soil physics and soil water flux formulations also will depend on the number of soil layers.

2. The water budget is much more sensitive to the number of layers in the soil profile than to the range of 2 orders of magnitude in saturated hydraulic conductivity considered in this study (comparison of Figures 4 and 5).

3. Practically all the effect of varying the number of soil layers on evapotranspiration is, in fact, on ground evaporation; that is, the effect on canopy transpiration is minimal (see Tables 2 and 3). The differences in ground evaporation occur in association with rain events and most noticeably when there is ponding of surface water (see section 4). In general, as Nr increases, ground evaporation also increases for both the high and low saturated hydraulic conductivity cases (see Figures 6 and 7).

4. For the case of low saturated hydraulic conductivity considered here, the change in soil water storage did not follow a monotonic relationship with the number of layers in the root zone. This is due to the nonlinear relationships between the dependent variables soil water potential and hydraulic conductivity and the independent variable volumetric water content. A comparatively simple illustration of this kind of anomalous behavior is given in Appendix A. In contrast, a monotonic relationship was observed for the case of high-saturated hydraulic conductivity.

Appendix A: Evaluation of the Anomalous Behavior in the Change in Soil Water Storage

Here we investigate the cause of the increase in soil water storage for the low saturated hydraulic conductivity case as Nr is increased from 1 to 2 (as shown in Figure 5) followed by the decrease in storage as Nr is increased from 2 to 3. Similar model behavior was shown, but not explained by *Koren et al.* [1999] for estimated evapotranspiration and soil moisture change using 2, 4, and 10 layers in the Oregon State University model [*Mahrt and Pan*, 1984].

The 65-day study period was divided into three time periods: (1) days 61–111, during which time four small rain events occurred, (2) days 112–116, when the largest rainfall event of the study period occurred, and (3) days 117–125, when no rain fell. Rainfall, change in water storage in each of the three soil zones, and ET for three soil layer combinations (Nr = 1, 2, and 3) for each of these periods, and for the entire study period, are shown in Table A1. It is seen that the increase of 8.9 mm in the change in total column water between the Nr = 1 and the Nr = 2 cases over the entire time period is mainly due to the 7.2 mm change in water depth during the major precipitation event on days 112–116.

Time series of hourly GE for three soil layer combinations for the 5-day period (days 112–116) are shown in Figure A1. Note that the number of layers in the upper and bottom zones was held constant at one for all figures in Appendix A. Explaining how the differences among the three GE time series could occur during days 115 and 116 as Nr is increased is tantamount to explaining the anomaly in Figure 5. This explanation is the goal of the remainder of Appendix A.

Figure A1 shows that both prior to and for almost 2 days after the rain on day 113, GE is the same for all three soil layer combinations. However, late on day 115, it is seen that GE peaks earlier for Nr = 2 than for Nr = 1 or Nr = 3, then decreases rapidly. Throughout most of day 116, GE continues to be lowest for Nr = 2 and highest for Nr = 1. By the end of day 116, GE is once again nearly the same for all three soil layer combinations and remains so to the end of the study period. The very large GE of 1.3 mm h^{-1} occurring late on day 115 is due to the high wind speeds (nearly 10 m s^{-1}), a large dew point depression (around 17.5°C) in the above-canopy air during this period, and the presence of ponded water.

To determine the physical processes responsible for the anomalous decrease in GE for Nr = 2 as compared to Nr = 1 and Nr = 3, infiltration for the three soil layer combinations was examined, the time series of which can be seen in Figure A2. The prominent features in Figure A2 are that after the heavy rain event the infiltration rate of ponded water varies for each of the three soil layer combinations and, consequently, infiltration ends at different times. Infiltration ends first for Nr = 2, then for Nr = 3, and ~6 hours later for Nr = 1. The shorter duration of ponding for Nr = 2 results in evaporation at the potential rate for a longer time period relative to the other two cases.

Why do infiltration rates vary as the number of layers in the root zone is increased? Once saturation is reached in the upper soil zone, as is the case after the heavy rain on day 113, further

Table A1. Comparison of Water Budget Variables for the Entire Study Period and Three Subperiods for the Low Saturated Hydraulic Conductivity Case[a]

Number of Layers in Root Zone	Upper Zone Δwater	Root Zone Δwater	Bottom Zone Δwater	Total Column Δwater	Total Evapo-transpiration
Entire Study Period, Days 61–125, Rainfall of 253.3 mm					
1	2.1	66.6	0.0	68.7	184.6
2	2.3	75.3	0.0	77.6	175.7
3	2.7	71.2	0.0	73.6	179.5
				8.9[b]	
				−4.0[c]	
Days 61–111, Rainfall of 119.6 mm					
1	2.1	0.4	0.0	2.5	117.1
2	2.2	2.0	0.0	4.2	115.4
3	2.3	2.4	0.0	4.7	115.0
				1.7[b]	
				0.5[c]	
Days 112–116, Rainfall of 133.7 mm					
1	3.9	85.6	0.0	89.5	44.1
2	3.4	93.3	0.0	96.7	37.0
3	3.4	89.5	0.0	92.9	40.5
				7.2[b]	
				−3.8[c]	
Days 117–125, Rainfall of 0 mm					
1	−4.0	−19.4	0.0	−23.4	23.4
2	−3.3	−20.1	0.0	−23.3	23.3
3	−3.3	−20.7	0.0	−24.0	24.0
				0.0[b]	
				−0.6[c]	

[a]Number of layers in the upper zone was held constant at 1. Study period is March 1 through May 4, 1996. Change in water in a soil zone (Δwater) is defined as the water depth at the end of the time period minus the water depth at the beginning. Water budget variables have units of mm.
[b]Difference in total column Δwater between one and two layers in root zone.
[c]Difference in total column Δwater between two and three layers in root zone.

infiltration is limited by the amount of water moving from the upper zone to the root zone. A time series for days 112–116 of the water depth in the root zone for the three soil layer combinations is shown in Figure A3. Prior to the rain on day 113 the three water depths in the root zone are very similar. After the rain begins, all of the soil layer combinations begin to gain water in the root zone. However, the Nr = 2 and Nr = 3 curves diverge from the Nr = 1 curve, indicating that water flux from the upper zone to the root zone is greater for the combinations with more root zone layers. At first, the water depth in the root zone increases more rapidly for Nr = 3 than for Nr = 2, as indicated by the slightly steeper slope. However, by late day 114 the opposite occurs. The greater rate of water flux from the upper zone to the root zone allows a higher infiltration rate for Nr = 2 relative to the other cases, as shown in Figure A2.

The observed differences in water fluxes between the upper and root zones can be explained in terms of the soil moisture

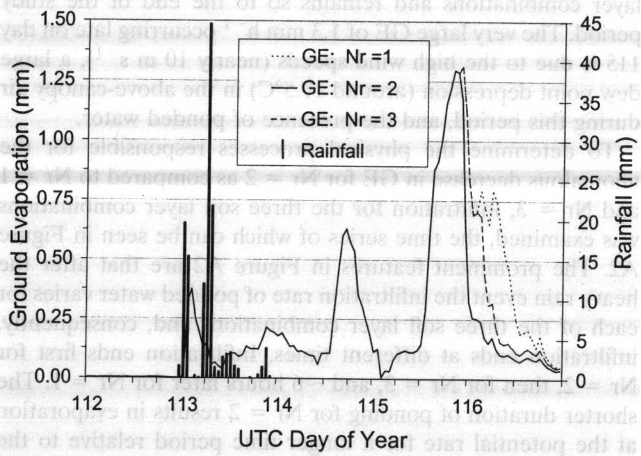

Figure A1. Hourly ground evaporation (GE) for three soil layer combinations and hourly rainfall from days 112 through 116 for the low saturated hydraulic conductivity case. The number of layers in the root zone (Nr) was varied from 1 to 3.

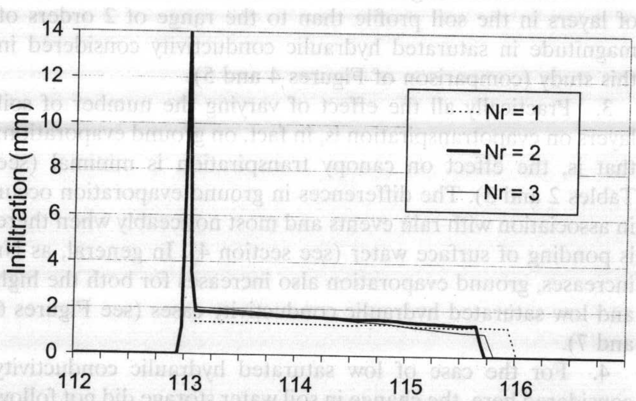

Figure A2. Hourly infiltration for three soil layer combinations from days 112 through 116 for the low saturated hydraulic conductivity case.

profiles. One of the variables that water flux depends on is the gradient of water content between layers. Volumetric water content for the upper zone and the top layer of the root zone for the three soil layer combinations is shown in Figure A4. We see here that the volumetric water content in the top layer of the root zone increases at a greater rate as Nr is increased. This is due primarily to the fact that as Nr is increased, the thickness of the root zone layers decreases. If water flux from the upper to the root zone depended solely on the moisture gradient, it is apparent from Figure A4 that as Nr is increased, less water flux into the root zone would occur and, consequently, infiltration would decrease. However, we have shown in Figures A2 and A3 that this relationship does not hold under all conditions; that is, water flux is not solely a function of the moisture gradient.

To demonstrate the complex relationships between soil water fluxes and the moisture profile, a simple three-layer model based on Richard's equation [*Jury et al.*, 1991] was used. In this model a fixed volumetric water content (near saturation) was used in the upper layer (layer 1), a varying volumetric water content (from 0.1 to 1.0) in the middle layer (layer 2), and a fixed, relatively dry volumetric water content (0.17) in the bottom layer (layer 3). These values were used to approximate the conditions that existed on day 115.

In the three-layer model, net water flux (mm s^{-1}) into the middle layer was estimated as follows:

$$\text{net water flux} = \frac{\partial w}{\partial t}\phi_{\text{tot}}\Delta z \tag{A1}$$

with

$$\frac{\partial w}{\partial t} = \frac{1}{\phi_{\text{tot}}}\left[\frac{\partial}{\partial z}\left(K\frac{\partial \psi}{\partial z}\right) + \frac{\partial K}{\partial z}\right], \tag{A2}$$

where

w volumetric water content relative to saturation (water volume/pore volume), mm^3 mm^{-3};
ϕ_{tot} total soil porosity, mm^3 mm^{-3};
z depth, mm;
t time, s;
K hydraulic conductivity, mm s^{-1}, equal to $K_{\text{sat}}w^{(2b+3)}$;
K_{sat} hydraulic conductivity at saturation, mm s^{-1};

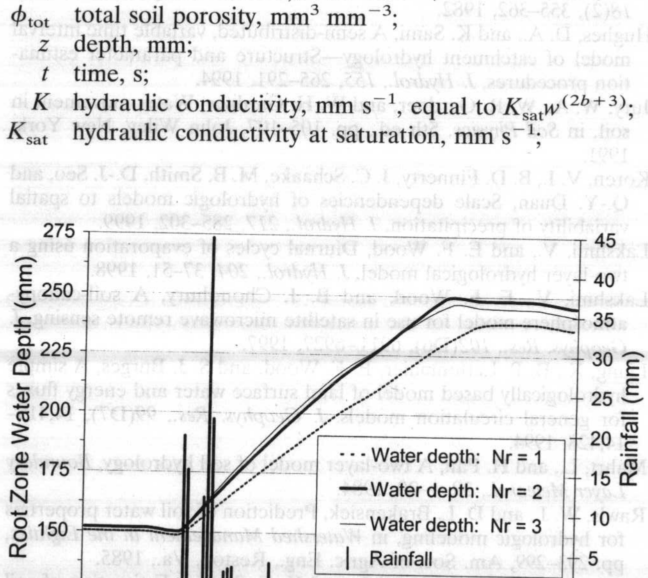

Figure A3. Hourly root zone water depth for three soil layer combinations and hourly rainfall from days 112 through 116 for the low saturated hydraulic conductivity case.

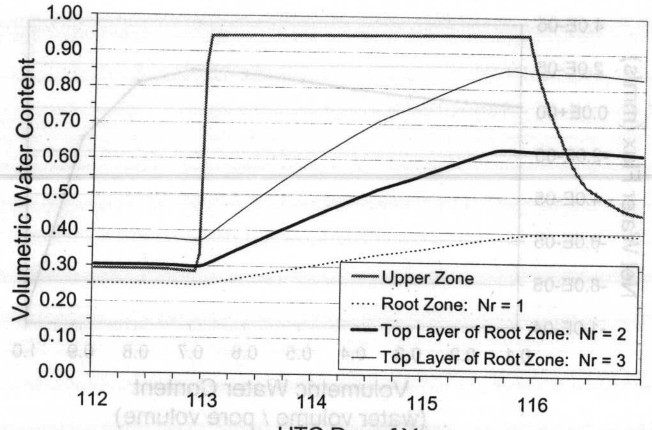

Figure A4. Hourly volumetric water content (w) in the upper zone and top layer of the root zone for three soil layer combinations for the low saturated hydraulic conductivity case. The upper zone w curve shown is for the case using one layer in the root zone; the upper zone w curve for the other two soil combinations is very similar.

ψ soil water potential (negative soil suction), mm, equal to $\psi_{\text{sat}}w^{-b}$;
ψ_{sat} soil water potential at saturation, mm;
b *Clapp and Hornberger* [1978] parameter, unitless.

The formulations for K and ψ above are based on those of *Clapp and Hornberger* [1978]. Applying mass continuity and expressing terms of (A2) as functions of w,

$$\frac{\partial w}{\partial t} = \frac{1}{\phi_{\text{tot}}}\left[\frac{\partial}{\partial z}\left(D(w)\frac{\partial w}{\partial z}\right) + G(w)\frac{\partial w}{\partial z}\right], \tag{A3}$$

where

$$D(w) = K\partial\Psi/\partial w = -bK\Psi/w$$

and

$D(w)$ is diffusion coefficient, mm^2 s^{-1}, and

$$G(w) = \partial K/\partial w = (2b+3)K_{\text{sat}}w^{(2b+2)} = (2b+3)K/w$$

and $G(w)$ is gravitational coefficient, mm s^{-1}.

Evaluating the first term in brackets of (A3) yields equation (A4) in which subscripts 1, 2, and 3 denote the upper, middle, and bottom layers and 1.5 and 2.5 denote the interfaces:

$$\left[\frac{\partial}{\partial z}\left(D(w)\frac{\partial w}{\partial z}\right)\right]_2 = \frac{D(w_{2.5})(\partial w/\partial z)_{2.5} - D(w_{1.5})(\partial w/\partial z)_{1.5}}{\Delta z}$$

$$= \frac{D(w_{2.5})\dfrac{(w_3 - w_2)}{\Delta z} - D(w_{1.5})\dfrac{(w_2 - w_1)}{\Delta z}}{\Delta z}$$

$$= \frac{D(w_{2.5})(w_3 - w_2) - D(w_{1.5})(w_2 - w_1)}{(\Delta z)^2}, \tag{A4}$$

where the value of w at the interfaces is determined as the average for the two layers:

$$w_{1.5} = \frac{w_1 + w_2}{2}.$$

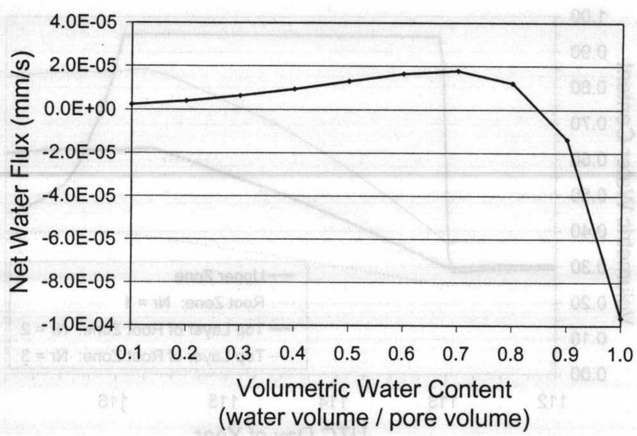

Figure A5. The relationship between the net water flux into and the volumetric water content (w) of the middle layer for a simple three-layer model. The upper layer w was held constant at 0.95, and the lower layer w was held constant at 0.17.

Evaluating the second term in brackets of (A3) (for layer 2) yields

$$\left[G(w)\frac{\partial w}{\partial z}\right]_2 = G(w_2)\left(\frac{w_3 - w_1}{2\Delta z}\right). \quad \text{(A5)}$$

For the Clapp-Hornberger parameter values [*Clapp and Hornberger*, 1978] used in this study (4.4 for loamy sand and 5.4 for clay loam) the exponents in the expressions for soil water potential and hydraulic conductivity make these functions very nonlinear. Thus the functional relationship between the moisture gradient and net water flux into the middle layer is not obvious. Figure A5 illustrates that for this simple model, when the volumetric water content of the upper layer is near saturation (0.95) and the water content of the bottom layer is relatively dry (0.17), the maximum water flux into the middle layer occurs at an intermediate value of middle layer volumetric water content (~0.7 in this case). The maximum is due to the nonlinear soil water dynamics and illustrates that vertical water fluxes are not necessarily monotonic functions of the moisture gradient. Instead, the water flux depends in a complex manner on both the gradient and the degree of saturation. These results, obtained from a simplified version of the SHEELS soil water dynamics, are for heuristic purposes only and do not exactly parallel the more complex processes in SHEELS. However, Figure A5 demonstrates how, for the soil moisture profile conditions during the period of ponding, the moisture profile for Nr = 2 could produce more water flux from the upper zone into the root zone than either Nr = 1 or 3, thus allowing water to infiltrate at a higher rate into the upper zone. These high infiltration rates, in turn, affect the ground evaporation rates through their control on ponding duration and postponding moisture conditions. As we have shown, these evaporation rates correspond exactly to the differences that occur in soil water storage as the number of layers in the root zone is increased.

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