

# Incorporating water table dynamics in climate modeling: 1. Water table observations and equilibrium water table simulations

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[1] Soil moisture is a key participant in land-atmosphere interactions and an important determinant of terrestrial climate. In regions where the water table is shallow, soil moisture is coupled to the water table. This paper is the first of a two-part study to quantify this coupling and explore its implications in the context of climate modeling. We examine the observed water table depth in the lower 48 states of the United States in search of salient spatial and temporal features that are relevant to climate dynamics. As a means to interpolate and synthesize the scattered observations, we use a simple two-dimensional groundwater flow model to construct an equilibrium water table as a result of long-term climatic and geologic forcing. Model simulations suggest that the water table depth exhibits spatial organization at watershed, regional, and continental scales, which may have implications for the spatial organization of soil moisture at similar scales. The observations suggest that water table depth varies at diurnal, event, seasonal, and interannual scales, which may have implications for soil moisture memory at these scales.

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# 1. Introduction

[2] The water cycles in the land and the atmosphere make up a fundamentally coupled system, with complex interactions among the various reservoirs (Figure 1) over a range of space and timescales. This coupling involves nonlinear dynamics in the atmosphere, land surface, and subsurface, leading to feedbacks that modulate the time-evolution of the system. The focus of this work is the role of the groundwater reservoir: its dynamic interaction with streamflow and its impact on soil moisture simulations at continental scales.

[3] With the exception of deserts where it can be deep and disconnected from the land surface, groundwater receives the surplus during wet periods and supplies the deficit during dry periods (arrows between boxes 2 and 4, Figure 1). As such, it can influence the near-surface and root-zone soil water content and potentially affect the energy and water fluxes between the land and the atmosphere. The groundwater reservoir also interacts with the

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rivers (box 3) by sustaining base flow in humid and subhumid climates, and by receiving river seepage in arid climates. It links the vertical fluxes near the land surface (infiltration) with the lateral fluxes across the landscape (river flow) and below the surface (groundwater flow). This lateral component not only helps close the water cycle, but also redistributes soil water in space. Therefore the groundwater reservoir is likely an important link in the coupled evolution of the hydrologic system over land.

[4] The role of water table in climate has been implicitly accounted for in several studies [e.g., *Koster et al.*, 2000; *Ducharne et al.*, 2000; *Walko et al.*, 2000; *Chen and Kumar*, 2001; *Seuffert et al.*, 2002; *Gedney and Cox*, 2003; *Yang and Niu*, 2003; *Niu and Yang*, 2003]. These studies attempt to improve the subgrid representation of soil moisture in general circulation models and regional climate models (GCMs and RCMs). TOPMODEL [*Beven and Kirkby*, 1979], a topography-based, watershed-scale formulation of equilibrium water table depth and soil water deficit, is often used to represent spatial variability. This approach recognizes the role of topography in controlling soil water, although the precise mechanisms, i.e., lateral groundwater flow and discharge to streams, are not dynamically represented.

[5] Water table dynamics have been explicitly accounted for in several studies aiming to improve the land surface schemes of GCMs. Most notably, *Abramopoulos et al.* [1988] formulated the NASA Goddard Institute for Space Studies' land surface scheme, which, in recognition of the potential role of the water table, provided two options for soil water drainage based on bedrock depth; if the bedrock is deep, then gravitational drainage is adopted, and if the

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Figure 1. A simplified view of the terrestrial water cycle with storages (boxes) and fluxes (arrows).

bedrock is shallow, no water escapes at the bottom of the soil, hence raising the water table. Habets et al. [1999] coupled a land surface scheme with a hydrology model (including the water table and groundwater-stream interaction) to simulate the water budget and river flows in the Rhone River basin in an off-line run with prescribed atmospheric forcing. Gusev and Nasonova [2002] incorporated a simple model of water table dynamics and its interaction with rivers into their land surface scheme and simulated the water and energy budgets in the boreal grassland of Valdai Hills in Russia (where the shallow water table may play an important role in controlling surface fluxes). Liang et al. [2003] demonstrated that introducing the water table into the Variable Infiltration Capacity (VIC) model significantly modified the near surface soil moisture distribution, resulting in a wetter bottom layer and drier top layer in two small Pennsylvania watersheds. Maxwell and Miller [2005] showed that, by coupling a land surface model with a groundwater model, the estimated soil moisture in the deeper layers is much closer to observations than without groundwater. Recently, Yeh and Eltahir [2005a, 2005b], through observations and model experiments in Illinois, demonstrated that the commonly used free-drain or no-drain soil bottom conditions can significantly bias the estimation of soil water flux and stream flow, and that without an explicit representation of the water table, the land surface water budget cannot be closed.

[6] Water table dynamics have also been directly coupled with the atmosphere [*Gutowski et al.*, 2002; *York et al.*, 2002] where all four reservoirs in Figure 1 are dynamically linked over a small watershed: a single-column atmospheric model interacts with a detailed land surface model that calculates water and heat fluxes through the unsaturated soil and the vegetation, which is in turn connected to a detailed groundwater model of shallow subsurface flow and interaction with stream segments. Their findings point to the conceptual advantage of direct atmosphere, land surface, and subsurface coupling by demonstrating potential feedbacks among the reservoirs.

[7] In this study, we contribute to this larger community effort by examining the observed water table dynamics and

by simulating the soil moisture at continental scales with explicit representation of the water table. Our work has two objectives. The first objective is to gain a better sense of the spatial-temporal characteristics in water table depth, based on observations in the United States. A big-picture view of salient spatial-temporal variability is useful for identifying regions and periods where and when the water table may play a role in near surface fluxes. It will also shed light on the following issues regarding the source of soil water variability at large scales and long times:

[8] 1. It is well understood that, at watershed scales and in humid climates, gravity-driven, lateral groundwater flow can result in wetter valleys and drier hills, affecting subsequent vertical fluxes. This has motivated the above-cited, TOPMODEL-based formulations of subgrid variability. It is also known that groundwater flow occurs at a range of scales; the flow regime of a small watershed is nested in the regional flow [*Toth*, 1963]. Are there systematic spatial patterns in water table depth at regional scales, in addition to watershed scales? For example, in a dry climate, the water table is below local drainage, and instead of reflecting local topography, its shape may follow the regional gradient. If so, are these large-scale features linked to large-scale soil moisture variability?

[9] 2. It is understood that soil moisture has long-term memory, which has large impacts on low-frequency climate variability over midlatitudes [e.g., Koster and Suarez, 2001]. In addition, van den Hurk et al. [2005] pointed out that most regional climate models predict warm-season soil that is too dry and runoff that is too sensitive to anomalies in precipitation minus evapotranspiration. They concluded that much of this is due to insufficient water storage in the land in these models, leading to less land surface memory and faster response to hydrologic events. Thus we wish to understand the role of the water table in controlling this memory. Specifically, what are the observed temporal structures in water table fluctuations? Vinnikov et al. [1996] reported that the decay time of top-1 m soil water is about 2-3 months in the Valdai Hills of Russia. Is this long-term persistence related to the year-round shallow water table (< 2 m below the land surface)?

[10] Our second objective is to implement the groundwater processes in an existing climate model and use it as a tool to address a number of science questions regarding the linked evolution of soil moisture, water table, and river flow at continental scales. We ask the following questions:

[11] 1. Will including the water table dynamics in climate simulations impact the soil moisture? Will it introduce new spatial structures into the latter, such as enhanced spatial organization since the water table reflects the structured topography and stream systems?

[12] 2. Will it introduce new temporal structures into the soil moisture? Will the long residence times of the ground-water reservoir anchor greater soil moisture memory?

[13] Our work will be reported in two companion papers. In this first paper, part 1, we address the first objective. As a means to interpolate the sparse and geographically biased observations, we use a simple groundwater model to derive an equilibrium water table that reflects the long-term balance between the climate and the geology. In part 2 [Miguez-Macho et al., 2007], we discuss the formulations that link the groundwater with soil and river flow, followed by parameterization and validation. Then we present the simulated soil moisture fields in North America, with and without water table dynamics, hence addressing the second objective. We stress that our work is exploratory and our findings are preliminary. As will be seen later, much of the needed hydrologic data are lacking. For example, to calculate groundwater flow, the hydraulic conductivity (K) is needed at greater depth, but it is yet to be compiled for large regions. It is our hope that the results presented here, based on less-than-desirable data, will help underscore the relevance of groundwater processes in understanding the terrestrial water cycle and the need for extending our hydrologic database to greater depths into the Earth's crust.

# 2. Observations of Water Table Depth in the United States

[14] We examine the observed water table depth in the lower 48 states of the United States to assess its spatial and temporal characteristics. Water table depth were compiled from the U.S. Geological Survey (USGS) database (http:// nwis.waterdata.usgs.gov/usa/nwis/gwlevels). From the entire record (1927-2005), we compiled 549,616 sites satisfying the following: the well is opened within 100 m of the land surface; it is not opened in a confined or mixed aquifer (code C and M); it is not a pumping well (code P), or injection well (I), or obstructed (O), damaged (W), plugged (M), discontinued (N), dried (D), or flowing (F) well. We limit our analysis to wells within 100 m of the land surface to examine the position of the water table in the surfacial, unconfined aquifers that are hydraulically linked to the land surface and the atmosphere. The aquifer-type code in the database flagged some but not all of the confined and mixed-confined aquifers, forcing us to adopt a cutoff depth. This leaves out the deep water table aquifers and includes the shallow confined aquifers, but it is a compromise between focusing on the shallow subsurface and including as many observations as possible. In the following we will use the term "water table" to refer to the water level in these wells. Such simplification and ambiguity is necessary given the data constraints, but they are sufficient for producing a

first-order, large-scale picture of water table conditions across the continent. Many of the sites are affected by pumping nearby as their time series reveal long-term trends of water level decline. A well-known case is the Ogallala Aquifers in the High Plains [Weeks et al., 1988] where largescale pumping began in the 1930s and continues to date. These affected wells are included here for a better observational coverage. Pumping also affected many other parts of the nation (see, e.g., http://pubs.usgs.gov/fs/fs-103-03/#pdf). Of the 549,616 sites, 542,281 (about 99%) have water table depth, and 7,335 have water table head (elevation) measurements, and not all sites have land or gage elevation for conversions between the two. The best monitored site has 18,444 (daily) observations, but 81% of the sites have only one reading, taken sometime over the record period (1927-2005). This snapshot nature of the groundwater sampling raises concerns as to how well this data set represents the long-term water balance at the sites. Nonetheless, they are the best continental data available, and we consider them sufficient for assessing large-scale properties.

#### 2.1. Spatial Characteristics of Water Table Depth

[15] Figure 2 is a map of temporally averaged (albeit only one observation at most sites), point observations of water table depth over the lower 48 states. At many sites in the eastern part of the country, it is within 5 m of the land surface. The deep water table in parts of Florida is caused by groundwater withdrawal [Solley et al., 1998]. The water table is very deep under the High Plains because of the high permeability of the Ogallala Aquifer System (fast drainage), as well as decades-long groundwater pumping [Weeks et al., 1988]. Over the western states, the water table is also deep, due both to the dry climate and heavy usage [Solley et al., 1998]. Even in semiarid regions, however, the water table can be shallow over some fraction of the landscape. For example, in the closed basins of Nevada and Utah, the water table is within 1 m of land surface in the broad valleys all year round [e.g., Fan et al., 1997]. One important observation from Figure 2 is that the water table depth varies greatly across the continent, which makes it difficult to adopt one soil depth everywhere for modeling purposes and hope to represent the correct soil drainage conditions.

[16] The observations have several limitations; they are biased toward river valleys and coastal regions where dense human settlements occur; they are biased toward where large and productive aquifers occur; they are snapshots taken at different times at different places; and they contain large pumping effects. These limitations prevent us from systematically assessing the inherent spatial structure that result from fundamental driving forces such as climate and geology. For this reason, we will use a simple groundwater flow model to simulate the long-term position of the water table, as a means to interpolate and synthesize the scattered observations, as discussed in section 3.

## 2.2. Temporal Characteristics of Water Table Depth

[17] To examine the timescales at which the water table may vary, we analyze the observed time series at 22 sites across the lower 48 states of the United States. In the first set of analyses, we focus on seasonal and interannual timescales at 20 sites over a 10-year period (1990–1999). The locations of the sites are marked on Figure 2, numbered



**Figure 2.** Point observations of water table depth at 549,616 sites in the lower 48 states of the United States, averaged over the record period at each site. Numbered sites are where 10-year time series analysis is performed.

from west to east. We attempted to choose one site in each state, and within a state, a well that is shallow, open in surfacial deposits, with no indication of pumping effect, and with observations frequent enough to detect at least the seasonal cycle. In many states, such a site cannot be found. Table 1 lists the sites chosen. Figure 3 plots the time series of water table depth and the power spectrum, obtained using the algorithm for unevenly sampled data from *Press et al.* [1992]. We observe that, first, a strong seasonal cycle is present at most sites, and second, there is a significant

 Table 1. Site Information for the 10-Year Time Series Analyses

| State          | USGS Site ID    | Ave <i>wtd</i> , | Land z, | Well<br>Depth_m | Formation                 | Number of<br>Observations | Observation<br>Frequency day |
|----------------|-----------------|------------------|---------|-----------------|---------------------------|---------------------------|------------------------------|
| Suite          |                 | 2.215            |         | Deptil, III     | 1 Officiation             | 50                        | Trequency, day               |
| Oregon         | 453514122575801 | 3.217            | 62      | 18              | unknown                   | 50                        | /3                           |
| California     | 385501121361904 | 1.926            | 10      | 16              | unknown                   | 108                       | 34                           |
| Idaho          | 433852116244801 | 2.587            | 784     | 13              | unknown                   | 634                       | 6                            |
| Montana        | 461518114090802 | 3.570            | 1,085   | 12              | Quaternary alluvium       | 64                        | 57                           |
| Utah           | 403158109372201 | 8.022            | 1,826   | 13              | outwash deposits          | 680                       | 5                            |
| New Mexico     | 350138106395503 | 3.100            | 1,502   | 15              | Santa Fe formation        | 80                        | 46                           |
| Texas          | 315712106361201 | 3.237            | 1,151   | 16              | Rio Grande alluvium       | 120                       | 30                           |
| Nebraska       | 413813103334501 | 3.593            | 1,256   | 24              | unknown                   | 565                       | 6                            |
| Kansas         | 390006095132301 | 5.960            | 255     | 16              | Newman terrace deposits   | 779                       | 5                            |
| Louisiana      | 320958091425501 | 3.192            | 20      | 23              | Miss. R. alluvial aquifer | 729                       | 5                            |
| Arkansas       | 335258091152301 | 6.381            | 45      | 29              | Quaternary alluvium       | 69                        | 53                           |
| Tennessee      | 350857089591401 | 10.804           | 83      | 18              | terrace deposits          | 116                       | 31                           |
| Wisconsin      | 440345089151701 | 0.860            | 266     | 4               | sand                      | 3,316                     | 1                            |
| Georgia        | 331507084171801 | 4.653            | 290     | 9               | surfacial aquifer         | 87                        | 42                           |
| Florida        | 271757081493001 | 1.936            | 23      | 5               | Norartesian sand aquifer  | 78                        | 47                           |
| Ohio           | 411137081172100 | 0.812            | 324     | 4               | outwash deposits          | 75                        | 49                           |
| North Carolina | 335629078115407 | 2.246            | 9       | 5               | post Miocene rocks        | 65                        | 56                           |
| Virginia       | 363928076332901 | 2.803            | 12      | 5               | Quaternary system         | 105                       | 35                           |
| New York       | 421556075281602 | 2.775            | 298     | 4               | sand and gravel           | 760                       | 5                            |
| Connecticut    | 415956073241501 | 3.727            | 248     | 9               | till                      | 476                       | 8                            |



Figure 3. Time series of (left) water table depth (m) and (right) its power spectrum (m<sup>2</sup>).



Figure 3. (continued)



**Figure 4.** (a) Hourly precipitation at Bound Brook; hourly water table at (b) Morrell and (c) Readington, with (d and e) 180th–270th day enlarged; and (f) the location of observations.

variability at interannual scales, the latter reflecting multiyear wet or dry spells that are preserved in the subsurface.

[18] In the second set of analyses, we focus on event and diurnal scales at two sites in New Jersey for the year 2003. We chose these sites because of our familiarity with the hydrology of the state. Figure 4f gives the location of the sites. Morrell 1 Well is 3.35 m deep and opened to a sandy aquifer, and Readington School 11 Well is 15.24 m deep and opened to a fractured shale aquifer underlying clay soil. We compiled hourly water table depth at the two sites and hourly precipitation at a location in between the two. We chose 2003 because there is no observation gap for that year. The time series plots in Figure 4 reveal the following features. First, at both depths, the water table responds to rainfall events in hours, a timescale for vertical drainage from the land surface to the water table, but the decay of each pulse takes many days, a timescale determined by lateral subsurface flow toward local streams. It points to the "quick recording" and "slow forgetting" nature of the water table response to atmospheric forcing. This suggests that the water table stays high many days after the events have passed, which may dampen soil moisture variability at event scales. Second, there is a strong diurnal cycle in the shallow well (Morrell 1) from mid May to mid October. The magnitude of the cycle is about 0.1 m, with the water table being the highest around 0800 and lowest around 1900 local time. This cycle is likely linked to plant transpiration, since

it is apparent only in the growing season, the water table resides in the root zone, and the timing coincides with the period of photosynthesis. Diurnal change in barometric pressure is ruled out as a cause because the well is shallow and in a sandy aquifer and therefore directly connected to the atmosphere [*Rasmussen and Crawford*, 1997], and this cycle is largely absent in the deeper well which should exhibit a stronger response to barometric change. This diurnal cycle implies that the soil moisture may have less diurnal variability because of a source so nearby. Hence including the water table dynamics in climate simulations can potentially reduce event and diurnal scale variability and enhance seasonal and interannual variability, as shown by the 10-year time series earlier.

#### 3. Equilibrium Water Table

[19] The water table observations discussed above have several limitations (biased toward river valleys and coastal regions; biased toward large and productive aquifers; snapshots at different times at different places; contain large pumping effects), which prevent us from systematically assessing the inherent spatial patterns that result from fundamental drivers such as climate and geology. As a means to interpolate and synthesize the scattered observations presented earlier, we build a simple groundwater flow model to simulate the long-term position of the water table.



Figure 5. Model domain over North America, color scheme giving land surface elevation in m.

[20] We introduce the concept of an equilibrium water table, defined as the climatologic mean water table, a result of long-term mass balance between two fluxes: the vertical, atmospherically induced flux across the water table, and the lateral, topographically induced flow below and parallel to the water table. The result is a smooth and undulating surface beneath the land topography, occasionally appearing at the land surface as wetlands, rivers, and lakes in a humid climate, and deep below the land surface in an arid climate. Shorter timescale climatic fluctuations, such as at interannual, seasonal, diurnal, and event scales, will cause the water table to rise and fall, small wavelength signals riding on top of the equilibrium surface.

[21] This equilibrium water table is important for two reasons. First, it is useful for portraying the first-order spatial variability in the water table height as a function of its primary controls: the climate and the geology. Roughly speaking, the climate at a location determines the vertical flux across the water table, and the geology determines the lateral flow below the water table. This first-order spatial variability will help identify regions where the water table may contribute to the water balance near the land surface. Second, it is useful for estimating the hydraulic parameters needed for modeling groundwater flow. One such parameter is the hydraulic conductivity (K) of sediments and bedrocks at greater depths beyond the existing soil database. It depends on the sediment-bedrock profile that reflects tectonics, weathering, and erosion-sedimentation in the past. Another parameter, discussed later, is the river hydraulic connection to the groundwater, which establishes the rivers as primary drainage for groundwater in humid regions. It

depends on stream network morphology such as drainage density and valley slope [e.g., Troch et al., 1995; Wood, 2002]. Both parameters are the result of long-term landscape evolution from complex interactions among climate, geology, and biota, and they are likely "tuned" to balance the drainage needs of the land. A humid climate plus a gentle relief likely produces a deeply weathered soil mantle, favoring groundwater flow and sustaining streamflow as its primary drainage; an arid climate plus steep terrain will likely lead to a shallow regolith (loose material), favoring surface runoff. Thus these hydraulic parameters are likely linked to the hydrologic equilibrium in a given climatic and geologic setting. The equilibrium water table, once found, can be used to estimate these parameters, a common proposition in solving inverse problems. Thus we will make an attempt to establish this equilibrium water table, based on physical principles of groundwater flow and constrained by the large number of observations discussed earlier, as a means of estimating these hydraulic parameters as well as portraying the inherent spatial patterns in the water table position.

# 3.1. Groundwater Mass Balance and River Flow

[22] We use a two-dimensional (lateral flow only) and steady state (equilibrium) groundwater flow model to estimate the equilibrium water table, hereafter referred to as EWT, over North America. The model domain is shown in Figure 5. The land elevation spans a range from below sea level to over 4,000 m, and the climate from superhumid (annual rainfall > 4500 mm) to arid (annual rainfall < 100 mm). This wide range of geologic-climatic conditions,



**Figure 6.** (a) Groundwater store,  $S_g$ , and associated fluxes, cross-section view; (b) plan view of lateral groundwater flow to neighboring cells,  $Q_n$  (n = 1...8); (c) replacing the square grid cells with octagons to calculate the width (w) of flow cross section between two cells; and (d) calculating flow transmissivity (T) assuming exponential decay in hydraulic conductivity (K).

and the lack of data on hydraulic parameters at depth, makes it difficult to represent the whole continent using simple and uniform parameterization schemes. Nonetheless we present our preliminary efforts, hoping to capture the first-order spatial variability in the hydrologic equilibrium over the continent.

[23] Refer to Figures 6a and 6b; for a model grid box, (i, j), of size  $\Delta x$  by  $\Delta y$ , the groundwater mass balance can be written as,

$$\frac{dS_g}{dt} = \Delta x \Delta y R + \sum_{1}^{8} Q_n - Q_r, \qquad (1a)$$

where  $S_g$  [L<sup>3</sup>] is groundwater storage in the column, R [L/T] is net recharge or the flux between the unsaturated soil and the groundwater,  $Q_n$  [L<sup>3</sup>/T] is lateral flow to/from the *n*th neighbor, and  $Q_r$  [L<sup>3</sup>/T] is groundwater-river exchange. At equilibrium, the left-hand side vanishes, and we have,

$$\Delta x \Delta y R = -\Sigma Q_n + Q_r; \tag{1b}$$

that is, the recharge balances the lateral groundwater flow to adjacent cells plus flow to the rivers within the cell. If the model grid cells are fine enough so that a cell is either a hillslope or a river cell, but not both, then  $Q_r$  may be dropped for a hillslope cell, so that,

$$R\Delta x \Delta y = -\Sigma Q_n \tag{1c}$$

and recharge R may be dropped for a river cell, so that,

$$Q_r = \Sigma Q_n$$
 (1d)

Thus at a hillslope cell the atmospherically induced recharge is dissipated laterally to the neighbors, and at a river cell, lateral groundwater convergence from the neighbors is discharged into the rivers.

[24] In continental-scale models, a feasible grid cell size is normally too coarse for resolving hillslopes and river valleys; a cell size on the order of 10 m is often considered adequate for such purposes [*Zhang and Montgomery*, 1994]. In this work, we use a grid size of 1.25 km, although fully aware of the limitations introduced to the results; we will simulate the equilibrium water table at this resolution for the time being and refine it in the future when it becomes feasible. [25] The lateral flow  $(Q_n)$  is calculated with Darcy's Law,

$$Q_n = wT\left(\frac{h_n - h}{l}\right) \tag{2}$$

where  $Q_n [L^3/T]$  is positive if flow enters the cell, w is the width of flow cross section [L], T is flow transmissivity  $[L^2/T]$ ,  $h_n$  is water table head in the *n*th neighbor [L], h is the head in the center cell (i,j), and l is the distance between cells:  $l = \Delta x$  along x or y, and  $l = \Delta x \sqrt{2}$  along the diagonal. To give equal chance of flow in all 8 directions from a cell, we assume equal width of flow (w) in all 8 directions, by replacing the square cells  $(\Delta x = \Delta y)$  with octagons of same surface area, as shown in Figure 6c, which gives the width of flow cross section,

$$w = \Delta x \sqrt{0.5 \tan(\pi/8)} \tag{3}$$

To obtain flow transmissivity, we examine two cases (refer to Figure 6d), the water table above or below the 1.5 m depth, since reliable soil data is only available to that depth. In case a, the water table is above 1.5 m depth and the transmissivity is,

$$T = T_1 + T_2 \tag{4a}$$

$$T_1 = \sum K_m \Delta z_m \tag{4b}$$

$$T_2 = \int_0^\infty K dz' = \int_0^\infty K_0 \exp\left(-\frac{z'}{f}\right) dz' = K_0 f \qquad (4c)$$

where m is the number of layers between the water table and the 1.5 m depth, and K is the hydraulic conductivity which is assumed to decay exponentially with depth as,

$$K = K_0 \exp(-z'/f), \tag{5}$$

where  $K_0$  is the known value in the bottom layer of the 1.5 m column, z' is the depth below 1.5 m, and f is the e-folding length, all discussed in detail later. In case b, the water table is d below 1.5 m and the transmissivity is,

$$T_2 = \int_d^\infty K dz' = \int_d^\infty K_0 \exp\left(-\frac{z'}{f}\right) dz' = K_0 f \exp\left(-\frac{z-h-1.5}{f}\right)$$
(6)

where d = z - h - 1.5 m, and z is the land surface elevation of the center cell. To ensure that flow from cell A to B is the same as from B to A under the same hydraulic potential, T is calculated for both cells involved and the average of the two is used.

#### 3.2. Hydraulic Conductivity (K) Profile

[26] To calculate groundwater flow, the hydraulic conductivity (K) for the geologic material is required at depth. Large-scale databases exist for the top meters for land surface modeling (e.g., http://ldas.gsfc.nasa.gov/), but parameters for continental-scale groundwater modeling are at best scattered in local government, academic, and industry archives. The USGS Regional Aquifer-System Analysis, known as RASA (http://water.usgs.gov/ogw/rasa/html/ introduction.html), provides a fundamental geologic framework for characterizing large-scale groundwater flow, yet efforts are needed to translate RASA findings into a hydraulic parameter database. Such a database must be supplemented by local aquifer tests routinely required by state regulatory agencies and archived in the states. These local data are important for quantifying RASA and for characterizing local aquifers and formations not included in RASA. Similar efforts must be made in other parts of North America. This effort has just begun as a part of our work and it will take many years to complete. Although we fully recognize this fundamental data deficiency, we will proceed with our modeling effort using commonly accepted assumptions on the vertical distribution of these hydraulic parameters.

[27] Porosity and permeability of geologic materials generally decrease with depth because pressure-heat release and weathering processes initiate at the land surface. Over scales of kilometers, the rate of decrease seems to follow a linear trend in competent rocks like sandstones, and an exponential trend in less competent rocks such as shale and mudstone [e.g., Deming, 2002]. Over scales of tens of meters, such trends are less obvious. One example is the fractured mudstones in the Early Mesozoic Basin Aquifers (http://capp.water.usgs.gov/gwa/ch 1/L-text4.html) in northeastern America, where the clay-rich soil is less permeable than the fractured bedrocks below. It underscores the high degree of complexity in local permeability fields and the need for observation support. It also poses a challenge for our attempt to represent groundwater flow over a continent using simple parameterization schemes.

[28] Over scales of meters, and for the purpose of watershed modeling, it is widely assumed that permeability decreases exponentially with depth [e.g., *Beven and Kirkby*, 1979], in the form of equation (5). *Decharme et al.* [2006] found that an exponential profile improved simulated discharge in a land surface model with river routing. Hence we adopt the exponential function in this study, in the absence of actual observations.

[29] The magnitude of f in equation (5), reflecting sediment-bedrock profile at a location, depends on factors that modulate the balance among tectonics, in situ weathering, and erosion-deposition processes. It is a complex function of past climate and geology. However, it is generally recognized that the balance between erosion and weatheringdeposition, leading to a particular regolith, depends strongly on terrain relief or slope (e.g., Ahnert [1970], Summerfield and Hulton [1994], and a synthesis by Hooke [2000]). Generally speaking, the steeper the terrain slope, the thinner the regolith; the transition from flat to steep terrain often marks the transition from transport-limited to weatheringlimited regimes. Climate also plays an important role, but the relationship between regolith and climate is more complex [Walling and Webb, 1983; Hooke, 2000]; for example, low rainfall produces low sediment runoff, leading to sediment accumulation and deep regolith; high rainfall leads to deeper percolation and denser biota, both enhancing in



**Figure 7.** Functional forms that are considered for representing the e-folding depth, f, as a function of terrain slope. The hyperbolic functions, for regolith and bedrock, are adopted here.

situ weathering and leading to deeper regolith as well [e.g., *Langbein and Schumm*, 1958; *Dendy and Bolton*, 1976]. Other factors, such as rainfall intensity and bedrock lithology, also play a limited role [*Hooke*, 2000]. In this exploratory work, we strive for simplicity and only consider the first-order control, the terrain slope, in determining the e-folding depth.

[30] The functional relationship between f and terrain slope is determined by trial and error. For regolith, the unconsolidated material, the best results are obtained with the following hyperbolic equation,

$$f = \frac{a}{1 + b\beta}$$
, for  $\beta \le 0.16$ ;  $f = 5$ m, for  $\beta > 0.16$  (7)

where *a* and *b* are constants, and  $\beta$  is the terrain slope. The best fit estimates are *a* = 120 m and b = 150, that is, *f* = 120 m at slope  $\beta$  = 0, and *f* = 5 m at and above  $\beta$  = 0.16 (relief = 200 m over a cell of 1.25 km), which are plotted in Figure 7. A map of terrain slope is shown in Figure 8a.

[31] For bedrock the constants are a = 20 m, and b = 125, i.e., f = 20 m at  $\beta = 0$ , and f = 1 m at or above  $\beta = 0.16$ . In the soil database, bedrock data are reliable only when bedrock is encountered above the 1.5 m depth (http:// www.soilinfo.psu.edu/index.cgi?soil data&conus&data cov&dtb&methods). For this reason, we use the parameters at this depth as the starting point for the exponential function. A map of the soil type at the 1.5 m depth is shown in Figure 8b, which reveals the large regions of shallow bedrock on the west coast, over the mountain ranges in the Great Basin, the Colorado Plateau aquifer system (sandstone) in the Rockies, the Edwards-Trinity aquifer system (sandstone and carbonate rocks) in Texas, the Ozark Plateaus aquifer system (carbonate rocks) in Missouri and Arkansas, and the extensive Pennsylvanian aquifer system (sandstones) underlying the Ohio and Tennessee river valleys. The resulting map of f is shown in Figure 8c, where the effect of terrain slope, Figure 8a, and bedrock distribution, Figure 8b, can be readily discerned.

# 3.3. Lateral Hydraulic Conductivity $(K_L)$

[32] To calculate lateral groundwater flow from cell to cell, the lateral conductivity values are needed. The LDAS soil database provides vertical conductivity for calculating vertical soil water fluxes for land surface modeling, but not the lateral. Lacking observations, we rely on the general concept of anisotropy, which relates the lateral conductivity,  $K_L$ , to the vertical conductivity,  $K_K$  through the anisotropy ratio,  $\alpha = K_L/K_V$ . It is well understood that soils and bedrocks of sedimentary origin exhibit stratified structure, leading to significant anisotropy, with a ratio in the range of 1-1000, and that the greater the lithological difference among the strata, the greater the anisotropy. We apply a crude rule for assigning the anisotropy ratio based on the clay content of the soil, because the presence of clay has a strong effect on anisotropy due to its platy mineral form and its low permeability as a unit. The set of values we chose, shown in Table 2, are well within the range observed in nature

[33] If bedrocks are encountered at the 1.5 m depth, then the soil class immediately above is used for the anisotropy ratio because we have little information on bedrock type and age. The soil type above is often indicative of the parent material below (although erosion-sedimentation can disrupt this connection). For example, a sandstone bed will likely weather into a sandy soil, both having low anisotropy; a shale or mudstone bed will likely produce clay-rich soil, both known to have high anisotropy.

## 3.4. Climatologic Mean Water Table Recharge (R)

[34] In the absence of direct observations, the recharge term *R* in equation (1a),(1b),(1c),(1d), is obtained from the archived results of a 50-year (1950–2000) integration of VIC model for the lower 48 states of the United States, on a  $0.125^{\circ}$  grid, by *Maurer et al.* [2002]. Recharge is calculated as the 50-year mean precipitation (observation-based) minus surface runoff and evapotranspiration (both model-estimated). For northern Canada, we use the results of the 13-year (1980–1993) VIC simulations for the globe [*Nijssen et al.*,



**Figure 8.** (a) Terrain slope, (b) soil type at 1.5 m depth, (c) the e-folding depth of K in m, and (d) water table recharge in mm/yr.

2001] on a 2° grid. Since the 13-year product does not separate out surface runoff from total runoff (surface runoff + base flow), we calculate recharge as precipitation minus evaporation, which is greater than actual recharge. A map of R is shown in Figure 8d, where a boundary is apparent between the United States and northern Canada because of these differences. As shown later, this boundary in recharge, plus the boundary in soil type (Figure 8b), resulted in a faint boundary in the simulated equilibrium water table. Thus our subsequent analyses, as described in part 2, will be limited to the southern side of this boundary.

#### 3.5. Equilibrium Water Table Depth

[35] Starting with the initial guess of water table being at the land surface, we solve equation (1c) iteratively. In humid river valleys, lateral convergence causes river cells to appear naturally. At these cells, the convergent flow is removed as groundwater discharge to the rivers (the term  $Q_r$ in equation (1d)) by keeping the EWT at the land surface. The resulting water table depth over the model domain is shown in Figure 9a, with details over the Rockies (Figure 9b) and the mid-Atlantic (Figure 9c).

#### 3.6. Comparison With Observations

[36] Figure 10a plots the simulated versus the observed head, the latter as the mean of point observations (if >1)

within a 1.25 km model cell, giving a total of 261,449 cells with observations. The water table head, instead of depth, is plotted because the head measures the potential energy that drives flow, and is therefore physically meaningful and can be calculated from physically based flow models. We focus on the residual, defined as the simulated minus the observed head, which should follow a Gaussian distribution with a zero mean.

**Table 2.** Anisotropy Ratio Used in This Study for the 12 Soil

 Classes in LDAS Database

| LDA    | AS Soil Class   | Vertical K | Anisotropy |  |
|--------|-----------------|------------|------------|--|
| Number | Name            | m/day      | Ratio      |  |
| 1      | sand            | 15.2064    | 2          |  |
| 2      | loamy sand      | 13.5043    | 3          |  |
| 3      | sandy loam      | 2.9981     | 4          |  |
| 4      | silt loam       | 0.6221     | 10         |  |
| 5      | loam            | 0.6048     | 12         |  |
| 6      | sandy clay loam | 0.5443     | 14         |  |
| 7      | silty clay loam | 0.1210     | 20         |  |
| 8      | clay loam       | 0.2160     | 24         |  |
| 9      | sandy clay      | 0.1901     | 28         |  |
| 10     | silty clay      | 0.0864     | 40         |  |
| 11     | clay            | 0.1123     | 48         |  |
| 12     | peat            | 0.6912     | 2          |  |



Simulated Equilibrium Water Table Depth (m from Land Surface)

**Figure 9.** Equilibrium water table depth (m) at 1.25 km resolution, obtained from a 2-D groundwater flow model that simulates the equilibrium water balance between vertical flux (recharge or stream discharge) and lateral flux (groundwater flow): (a) the entire model domain, (b) details over the Colorado Plateau, and (c) details over the mid-Atlantic coast.

[37] Figure 10b plots the histogram of the residual. The histogram is after subtracting 2 m from the simulated head over the entire domain. This 2 m shift in the EWT serves two purposes, first, to center the residual histogram so that the peak occurs at zero, and second, to compensate for the biashigh in the simulated head due to allowing EWT to rise to the land surface at river cells. At the 1.25 km resolution, the mean river elevation, the drainage level for the groundwater, is likely below the mean land surface elevation. This bias is propagated upland, causing a bias-high at high-elevation cells. Because it is difficult to derive an observation-based

mean river level (also scale-dependent) for the continent, we resort to a simple shift of the EWT as suggested by observations, so that the residual histogram is centered at zero. The maps in Figure 9 already reflect the shift. Clearly this bias cannot be 2 m everywhere, but it is the simplest way to correct the bias in the right direction.

[38] With the shift, 12% of the cells have a simulated head within 1 m of observations, 24% within 2 m, 44% within 5 m, and 66% within 10 m. Comparisons with observations are also affected by the nature of observations. For example, at 81% of the 549,616 sites, only one



**Figure 10.** Comparison between simulated EWT ( $\Delta x = 1.25$  km) and point observations. (a) Simulated head (m) versus observed head (m), the latter as mean point observations (if more than 1) within a 1.25 km grid cell. (b) Histogram of residuals, defined as simulated minus observed head. The histogram is shifted toward the left by 2 m, in order to center the peak at zero. (c) Residual (m) versus land surface elevation (m). (d) Residual (m) versus terrain slope.

measurement was made over the record period ( $\approx$ 75 years); that is, the observations were at different times. This snapshot nature renders the observations less reliable for validating equilibrium conditions. Also, of the 261,449 model cells with observation, 70% have only one point observation, making the comparison difficult in rough terrain or near pumping wells where the water table has a steep gradient.

[39] The shifted histogram is centered at zero, but skewed to the right. That is, there are more cells where the simulated water table is higher than observed. To examine where the biases occur over the wide range of conditions across the continent, we plot the residual versus land surface elevation in Figure 10c and terrain slope in Figure 10d. The negative correlations in both suggest that the bias-high mostly occurs at the lower-elevation range and flatter terrains. We made no attempt to correct this, recognizing that groundwater pumping can significantly lower the observed water table, leading to low bias in the observations. Since pumping tends to occur in agricultural and urban areas, which mostly occupy river valleys and coasts, its effect is likely stronger at this lower topographic end (see USGS report http://pubs.usgs. gov/fs/fs-103-03/#pdf on groundwater depletion in the nation).

[40] The largest residuals are found where the terrain is steep (slope>10%, Figure 10d), likely for the following reasons. First, the observations only include wells within 100 m of the land surface. This leaves out deep wells in regions of deep water table such as the high deserts in the southwest. In the model, an arid climate plus deep soil will result in a deep water table, and no cutoff depth is imposed. This mismatch may partially explain the large bias-low in the simulation. Second, in places with steep slope such as near a cliff side, the water table can be very different across a short distance. Model simulations based on mean topography will give very different results from point observations made either on cliff top or at cliff base. The third reason is the presence of local perched aquifers above the regional groundwater system. These small aquifers are important locally and are often characterized and monitored (hence in the USGS database), but they cannot be captured in our model which simulates large-scale flow. Finally, our parameterization of the hydraulic conductivity predicts a very deep soil if bedrocks are not encountered at 1.5 m depth (the deepest reliable bedrock data). However, with the case of thin alluvium in high mountains, bedrock is not far from the 1.5 m depth. This can cause fast drainage and deep water table in the model, much deeper than observed in an



**Figure 11.** Equilibrium soil water profile above a water table at 10 m, 5 m, 2 m, and 1 m depth below land surface.

intermountain valley. This again underscores the need for better quantification of subsurface hydraulic properties. For now we are content that such large deviation only occurs at a small fraction of the domain ( $\approx 0.1\%$  of the cells with residual  $\geq 100$  m).

[41] Although some residuals are alarmingly large, we note the difficulty in capturing detailed hydrologic conditions without observed hydraulic properties, and using simple and uniform parameterization across the continent (no local tuning), and coarse grid spacing (hydrologically speaking). More importantly, our goal is not to describe local groundwater conditions, but to capture the first-order spatial variability across a continent as the result of longterm and large-scale climatic and geologic forcing.

#### 3.7. Spatial Variability in Water Table Depth

[42] One can identify a few salient features from Figure 9 that are relevant to continental-scale water cycle. First, at a given location, it is the balance between the vertical flux (climate-induced) and the lateral divergence/convergence (geologically controlled), not one of them, that determines the hydrologic equilibrium. The water table can be shallow in a humid climate, such as in the southeast, as well as in an arid and semiarid climate, such as in the intermountain valleys of the west. In the latter case, large winter precipitation at high elevations, via lateral transport down the steep slopes, feed the groundwater in the valleys. Likewise, the water table can be deep in an arid and semiarid climate, such as in the northern Great Plains, as well as in a humid climate, such as over the hills of the Appalachians. In the latter case, efficient drainage out of the shallow soil into the dense river networks over hilly terrain can keep pace with the large amount of climate-induced input. Similarly, the water table can be deep and shallow in a given climatic setting, depending on the bedrock depth, such as in the Ohio River valley. Therefore a realistic portrait of the water table conditions over a continent must consider both the climatic and the terrain factors.

[43] To answer the question posed early, regarding the spatial structure in water table depth beyond the watershed scales, we note a few large-scale features in Figure 9. The water table becomes shallower as one travels from the higher drainage of the Mississippi River to the lower drainage (because of topographic gradient), and from the western side to the eastern side of the drainage (because of climatic gradient). Over the Atlantic seaboard, the water table becomes shallower as one travels toward the ocean. These regional trends in the water table depth may seem obvious, but they lead to the following question: do these trends introduce large-scale features in the soil moisture fields, which in turn may introduce mesoscale patterns in land-atmosphere fluxes?

# 4. Water Table Depth and Soil Moisture

[44] We conclude this paper with a brief discussion of the linkage between the water table depth and the soil moisture profile. We numerically solve the equations of water flux in an unsaturated soil column, using the water table as the lower boundary condition. Vertical flux in an unsaturated column can be described by the Richards' Equation,

$$q = K_{\eta} \left( \frac{\partial \psi}{\partial z} - 1 \right), \quad K_{\eta} = K_{f} \left( \frac{\eta}{\eta_{f}} \right)^{2b+3}, \quad \psi = \psi_{f} \left( \frac{\eta_{f}}{\eta} \right)^{b} \quad (8)$$

where q is water flux between two adjacent layers, K is hydraulic conductivity at given volumetric water content  $\eta$ ,  $\psi$  is soil capillary potential, b is soil pore size index, and subscript f denotes the quantity at saturation. The above equation is solved with zero flux through the column, for three soil types, and at four water table depths (10, 5, 2, and 1 m). At the water table, saturation is prescribed (neglecting capillary fringe). The resulting equilibrium soil moisture profiles are given in Figure 11.

[45] The effect of the water table on soil moisture is different in different soils. For the clay loam, where capillarity is strong, a water table depth of 10 m can still be "felt" in the root zone and near the surface. For the sandy loam, the water table has little role as a source if it is below the root zone, and it will function as a receptor for rapid drainage of rainfall events. This simple result points to the potential link between soil moisture and the water table: by influencing the soil water content from below, the water table acts as the lower boundary condition of the soil column, much in the same sense that the atmosphere acts as its upper boundary. Both can drive the soil water flux in the unsaturated zone at their own characteristic spatial and temporal scales. In the context of Figure 9, the simulated equilibrium water table depth, this simple result leads to the hypothesis that the water table can be shallow enough (< 5 m) deep) to influence soil moisture over large regions of the continent, particularly in the humid southeast and the intermountain valleys of the west. This hypothesis will be tested in part 2.

#### 5. Summary

[46] In this paper, we used USGS observations to examine the spatial and temporal characteristics of water table depth in the lower 48 states of the United States. We find that, at many sites in the eastern part of the country, as well as in closed basins and mountainous valleys in the West, the water table is shallow, lying within 5 m of the land surface. Thus there is a potential for the water table to anchor the soil moisture patterns in these regions. In other parts of the country, primarily in the West, the water table can be very deep and thus relatively disconnected from soil moisture. This large spatial variation in water table depth across the continent underscores the difficulty, from a modeling perspective, of correctly representing soil drainage with a single uniform soil depth.

[47] Temporally, we observe a strong seasonal cycle and significant interannual variability, at most sites. In addition, upward and downward fluctuations in water table depth at event and diurnal timescales reflect the balance between vertical drainage, lateral subsurface flow to local streams, and the upward flux to feed evapotranspiration. The longer timescales of the lateral processes, as well as the inertia provided by the large groundwater reservoir, have a potential for increasing soil moisture memory.

[48] While providing the best available direct observational coverage, water table observations are still scattered and sparse in most areas. Our findings point to the lack of, and hence the need to improve, large-scale and long-term water table observations, as well as the need to extend our hydrologic database deeper into the Earth's crust. For now, we use a simple two-dimensional groundwater flow model, constrained by the USGS observations, to construct an equilibrium water table as a means for synthesizing and interpolating between the measurements. This equilibrium water table is a useful conceptual and practical tool, for two reasons. First, it brings out more clearly the first-order spatial variability in water table depth across the continent that results from the long-term balance between large-scale climatic and geologic forcing. It illustrates how the water table depth at any location is a function of both the vertical,

climate-induced flux and the lateral, geologically controlled divergence/convergence of surface and subsurface flow. Any realistic portrait of the water table conditions, and hence soil moisture conditions, over a continent must consider both these climatic and terrain factors. Second, the equilibrium water table provides a means of estimating the parameters needed for modeling groundwater flow, such as hydraulic conductivity and the hydraulic connection between the groundwater and the rivers. These hydraulic parameters are linked to the hydrologic equilibrium in a given climatic and geologic setting, enabling us to estimate them from the equilibrium water table using inverse methods.

[49] In the next paper, part 2, we incorporate the water table dynamics in a climate model and explicitly investigate the role of the groundwater reservoir as a driver of soil moisture at continental scales.

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