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Potential groundwater contribution to Amazon evapotranspiration

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Abstract

Climate and land ecosystem models simulate a dry-season vegetation stress in the Amazon forest, but observations show enhanced growth in response to higher radiation under less cloudy skies, indicating an adequate water supply. Proposed mechanisms include larger soil water store and deeper roots in nature and the ability of roots to move water up and down (hydraulic redistribution). Here we assess the importance of the upward soil water flux from the groundwater driven by capillarity. We present a map of water table depth from observations and groundwater modeling, and a map of potential capillary flux these water table depths can sustain. The maps show that the water table beneath the Amazon can be quite shallow in lowlands and river valleys (< 5 m in 36% and < 10 m in 60% of Amazonia). The water table can potentially sustain a capillary flux of > 2.1 mm day⁻¹ to the land surface averaged over Amazonia, but varies from 0.6 to 3.7 mm day⁻¹ across nine study sites. Current models simulate a large-scale reduction in dry-season photosynthesis under today's climate and a possible dieback under projected future climate with a longer dry season, converting the Amazon from a net carbon sink to a source and accelerating warming. The inclusion of groundwater and capillary flux may modify the model results.

1 Introduction

The Amazon rainforest is the planet's largest and biologically richest ecosystem, and the threat of climate change and deforestation requires an understanding of its responses and feedbacks to its environment. One standing question is how well the forest fares in the dry season under the climate today, which is pertinent to how well it will fare under the projected future climate with a longer dry-season. Although annual rainfall is abundant, a large part of the Amazon experiences a multi-month dry season in the Austral winter. Soil water deficit and partial shut-down of photosynthesis are indeed simulated in state-of-art climate and ecosystem models (Kleidon and Heimann,

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2000; Werth and Avissar, 2004; Baker et al., 2008; da Rocha et al., 2009). However, a seminal paper over two decades ago (Shuttleworth, 1988) based on observational syntheses had shown that evapotranspiration (ET) in the dry season is no less than in the wet season. Subsequent flux-tower measurements at multiple sites and satellite images revealed the same (Saleska et al., 2000, 2007; Xiao et al., 2005; Huete et al., 2006; Ichii et al., 2007; Myneni et al., 2007; Juarez et al., 2007; Fisher et al., 2009), all suggesting that the Amazon forest does well or better in the dry-season.

A considerable literature exists proposing different mechanisms to explain the observed absence of water stress. First, soil water store is far greater in nature, which is filled in the wet season and sustains ET later in the dry season; in models, excess infiltration drains through the shallow soil and is removed as river outflow, no longer available for plant use later. Second, tree roots extend deeper than model soil-root system and can access deep water (Nepstad et al., 1994; Kleidon and Heimann, 2000; Ichii et al., 2007; Harper et al., 2010). However global syntheses of observations suggest that 95% of root mass resides in the top 2 m of soil for all major biomes of the world (Schenck and Jackson, 2002), and tracer studies in the field found that the root system does not seem to take water beyond 2 m depth (Sternberg et al., 2002; Romero-Saltos et al., 2005). These observations suggest that the role of deeper roots is not entirely clear, and they seem to justify the common practice of including only the top 2 m soil in models. Third, the small fraction of deep roots, albeit insignificant in mass, can be efficient water conduits via hydraulic redistribution (Dawson, 1993; Caldwell et al., 1998; Burgess et al., 1998; Oliveira et al., 2005; Lee et al., 2005; Amenu and Kumar, 2008); rooting depth can be tens of meters in plants relying on deep source as in arid climate (Canadell et al., 1996). A modeling study (Lee et al., 2005) found that incorporating HR significantly reduced, although far from eliminated, the model ET bias (still 50% less than observed). The modeling study by Baker et al. (2008) further shows that the combination of all above (deep soil, deep rooting depth, and hydraulic redistribution) performs better than any one alone. Fourth, upward soil water flux driven by capillary force in the dry season, from the deeper and wetter soil to the shallower and drier soil,

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may be an important mechanism (da Rocha et al., 2004; Romero-Saltos et al., 2005). Finally, the groundwater may be a source where it is directly accessible by roots as suggested by field studies (Poels, 1987; Vourlitis et al., 2008). Diurnal variations in water table depth that coincide with the period of photosynthesis in a Suriname rainforest are direct indications of water table contribution to forest ET (Poels, 1987).

We have not found any systematic investigations of the influence of groundwater on land surface fluxes across the wide range of hydrologic-ecologic conditions found in the Amazon. Observations of water table depth were reported at very few sites in the Large Scale Biosphere-Atmosphere Experiment in Amazonia (LBA), an internationally-coordinated research initiative led by Brazilian scientists. Nor did we find Amazon-wide modeling studies that incorporate the water table in simulating Amazon land surface fluxes. The objective of this study is to fill in this knowledge gap. First, we present water table observations compiled from all available sources. Second, we synthesize the sparse observations using a simple mechanistic groundwater flow model, to interpolate the observations and to elucidate process controls on water table depth from hillslope to continental scales. Third, we calculate the potential capillarity-driven soil water fluxes from the modeled water table. Finally, we compare this potential capillary flux to independent ET estimates at nine sites where they are available. Throughout the study, we attempt to keep the methodology as simple as possible.

2 Water table observations

The first issue to be examined is the depth of the water table. A deep water table will have little influence on root-zone soil moisture and surface fluxes, but a shallow water table (within or not far below rooting depth) will impede soil water drainage during rain periods and hence prolong the effect of rain, and moisten deep roots and shallow soil during dry periods via hydraulic redistribution and capillary flux.

We compiled observations of water table depth (WTD) at 34 351 sites over the South American continent (Fig. 1) from government archives and published literature. We searched the government database of each country in S. America and

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each province except for Brazil and Chile. Repeated emails were sent and phone calls were made where no data was found on the government website or the data is incomplete. Tens of published articles are also found which reported the water table depth. Data are mostly presented as plots or maps, from which we read the approximate values. The Brazilian Geological Survey (<http://siagasweb.cprm.gov.br/layout/index.php>) is the single largest data source (98%) with observations at 33 570 wells in unconfined aquifers. Unfortunately they are concentrated in the developed eastern and southeastern Brazil and clustered over large metro regions, because these wells were drilled for groundwater exploitation, not for observation and research; groundwater is considered cleaner and has replaced surface water to be the major source for municipal supply. About 95% of the wells in the dataset report high pumping rates. In Amazonia, all metro regions are situated on principle aquifers and supplied by groundwater; regional water table decline >20 m has been seen in recent years (http://www.ana.gov.br/pnrh_novo/documentos/01%20Disponibilidade%20e%20Demandas/VF%20DisponibilidadeDemanda.pdf). This introduces a low bias in the observations, an issue to be kept in mind when validating model simulations that cannot incorporate groundwater pumping. However, this is the best dataset available at this time, and it offers a lower bound to the natural water table depth.

In the Amazon, WTD varied from land surface to 159 m (distribution shown in Fig. 1 inset). Even with widespread pumping, the water table is relatively shallow: 34% of the sites with WTD < 5 m, 57% of the sites with WTD < 10 m, and the peak of the histogram occurs at 2–4 m. This suggests that shallow WTD is widely observed. At least at these observation sites, it has the potential to influence the land surface.

However, these observations are far too sparse to resolve the spatial variations in water table at scales relevant to surface and groundwater convergence (hilltops to valleys) and to discern patterns of climate and terrain control on groundwater regimes across the continent. There is a need to first, interpolate the observations to fill in the gaps, and second, to synthesize the observations within the framework of hydrologic

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processes, so that the hydrologic reasons of deep vs. shallow water table can be revealed. This is the subject of the next section.

3 Groundwater flow simulation

We use a groundwater model to simulate the hydrologic equilibrium water table depth over the continent. This equilibrium water table depth reflects the long-term balance between the climate-induced vertical flux ($P-ET$) and terrain-induced lateral divergence from high grounds and convergence into discharge zones. The model was developed and tested in our recent work over N. America using >567 000 site observations and mapped wetlands (e.g., Fan and Miguez-Macho, 2010). The model concept is simple (Fig. 2a); climate forcing is precipitation (P) minus ET and surface runoff (Q_s), giving the net flux across the water table, or recharge (R): $R=P-ET-Q_s$. The latter is redistributed by lateral groundwater divergence under high grounds (Q) and convergence under lowlands which feeds rivers and wetlands (Q_r). The water table position is constrained by the sea level along the coast, the ultimate baseline for continental drainage. The resulting water table is an undulating surface beneath land topography, occasionally seen at the surface as springs, wetlands and rivers, and merging with the sea level along the coastline.

At the hydrologic equilibrium, mass balance dictates that in hillslope cells, recharge (R) balances lateral divergence (Q) to the lower neighbors (Fig. 2a):

$$R = \sum Q \quad (1)$$

And in valley cells, lateral convergence (Q) balances discharge into rivers and wetlands (Q_r):

$$\sum Q = Q_r \quad (2)$$

Equation (2) also applies to coastal cells where groundwater must exit before the sea. River-wetland cells appear naturally in the simulation where water table rises to the

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land surface as dictated by mass balance. At these cells, the water table is reset at the surface, mimicking river and ET removal in nature (Q_r). The lateral groundwater flow between cells (Q) is calculated with the Darcy's law and the Dupuit-Forchheimer Approximation (lateral flow only) (see, e.g., Freeze and Cherry, 1979), which relates the water table slope to groundwater flow rate:

$$Q = wT \left(\frac{h - h_n}{l} \right) \quad (3)$$

where Q is the flow between the center cell and neighbor n , w the width of cell interface, T the transmissivity, h the water table head above sea level, h_n the head in neighbor n , and l the distance in between. To obtain T (integration of hydraulic conductivity over depth), we examine two cases (Fig. 2b): water table above (case-a) or below (case-b) the depth (d_0) with known hydraulic conductivity K_0 . The distinction is necessary because global soil datasets do not include information below the top 2–3 m of land surface, and hence they need to be treated separately. In case-a, the water table depth d_1 is $< d_0$ and we have,

$$T = T_1 + T_2, \quad T_1 = K_0(d_0 - d_1), \quad T_2 = \int_0^{\infty} K dz' = \int_0^{\infty} K_0 \exp\left(-\frac{z'}{f}\right) dz' = K_0 f \quad (4)$$

where z' is depth below d_0 , with K assumed to decrease exponentially from K_0 ,

$$K = K_0 \exp(-z'/f) \quad (5)$$

where f is the e-folding depth (more below). In case-b, the water table is d_2 below the known K ,

$$T = \int_{d_2}^{\infty} K dz' = \int_{d_2}^{\infty} K_0 \exp\left(-\frac{z'}{f}\right) dz' = K_0 f \exp\left(-\frac{z - h - d_0}{f}\right) \quad (6)$$

where z is land surface elevation of the center cell.

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To calculate groundwater flow, hydraulic conductivity K for the geologic material must be known at tens of meters of depth, but global soil datasets do not go below the top 2–3 m of land surface. Lacking actual measurements, we adopt common assumptions on its vertical distribution. Permeability of the Earth's crust generally decreases with depth (Manning and Ingebritsen, 1999). At the scales of tens of meters, it is widely assumed that the decay is exponential (e.g., Beven and Kirkby, 1979; Jiang et al., 2009), in the form of Eq. (5). The e-folding depth, f , reflecting sediment-bedrock profile at a location, depends on the balance among tectonics, in-situ weathering, and erosion-deposition, a complex function of climate, geology and biota. But the balance depends strongly on terrain slope; the steeper the land, the thinner the soil. Climate plays an important role but the mechanisms are more complex; e.g., low rainfall produces low sediment runoff, leading to sediment accumulation and deep soil; high rainfall leads to deeper percolation and denser biota, enhancing in-situ weathering and leading to deeper soil as well. For simplicity with only the first order control, we consider the terrain slope only. The function of f (in m) with slope s is determined by calibration to best reproduce water table and wetland observations in N. America (Fan and Miguez-Macho, 2010) and takes the form:

$$f = \frac{75}{1 + 150s}, \quad f \geq 4 \text{ m} \quad (7)$$

Since there are no continental-scale observations of water table recharge (R in Eq. 1 and Fig. 2a), it is obtained from four global land surface models as $R=P-ET-Qs$ (Fig. 3) where P is observation-based, but ET and Qs (surface runoff) are model simulated. The four models give different estimates of recharge forcing due to inherent differences in flux parameterization and soil configurations (see Table 1 for soil depth and layer information). We note that CLM gives the lowest recharge estimates in the Amazonia which will result in the deepest water table simulation; everything else equal, a higher R leads to a higher or shallower water table. To obtain the range of water table depths due to uncertainties in recharge, we use R from all four models and choose the result that best agrees with the WTD observations in Fig. 1, and that gives

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a conservative (or deeper) estimate of the water table depth and its potential impact on land surface.

Digital topography data, at 3 arc-second resolution, was obtained from the US Geologic Survey (<http://hydrosheds.cr.usgs.gov/>) and aggregated to 9 arc-second for the simulation in S. America (~157 m×274 m at the southern tip of the continent, to ~280 m×274 m at the Equator), totaling 248 103 766 cells over the continent and 84 184 468 in Amazonia. This grid size is a compromise between the need to resolve fine terrain features and computation feasibility. Terrain slope is shown in Fig. 4a, which is used to calculate the rate of decrease in permeability with depth (Eq. 7). It gives the first indication of land drainage efficiency; the flat (purple) Orinoco, upper and middle Amazon, and the vast area from Brazilian Pantanal to Argentina Pampas are regions of poor drainage and likely high water table.

Soil information is derived from UNESCO Food and Agriculture Organization (FAO) digital soil map of the world at 5 arcmin grids. Fractions of silt, clay, and sand are mapped to 12 soil-texture classes defined by the US Department of Agriculture (<http://soils.usda.gov/education/resources/lessons/texture/>). The 12 soil classes (Fig. 4b) are assigned soil hydraulic parameters based on established and commonly adopted procedures (Clapp and Hornberger, 1978). The dominant soil types in the Amazonia are clay-loam (class 8) and clay (class 11), both fine-textured and conducive to strong capillary fluxes.

Lateral boundary condition for the water table head is set at sea level along the coast. Although our primary interest is in the Amazonia, the simulation needs to include the entire continent so that the sea level, the true physical boundary condition for water table head, can be used to constrain the model. Starting the initial water table at the land surface, we solve the flow equations iteratively until the mass balance error is less than 1 mm year⁻¹. The simulations, forced by four recharge estimates, are shown in Fig. 5 at the 9 arc-sec resolution (~274 m).

We note the broad features in WTD distribution common to all four simulations. A shallow water table is found in four types of settings. The first is the humid lowlands

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of the Orinoco and Amazon basins with high rainfall over a land of poor drainage. The unusually low elevation of the interior Amazon, at such a great distance from its outlet, was marveled at by early explorers (e.g., Wallace, 1889) describing the region as a flat plate with a steep rim (the Andes). The Pastaza-Marañón basins in the Peruvian Amazon, no more than 200 m above sea level, are down-warped basins as the Andes rose (Clapperton, 1993) and are being actively filled in by sediments derived from the Andes. Drainage in these continental depressions simply cannot keep pace with the high annual rainfall. The second setting is the flat lowlands in semi-arid to arid climate, but nonetheless receiving large-scale groundwater convergence, as in the vast region from Mato Grosso of Brazil to the Pampas of Argentina, where the Pantanal, the world's largest freshwater wetland, has developed; in the model, this is a direct result of allowing lateral redistribution of recharge by groundwater flow at inter-cell to continental scales. The third setting is the coastal belt because the water table cannot drop below the sea level; in the model, this is a direct result of setting the sea level as the lateral boundary condition. The fourth is the river valleys (details in Figs. 9 and 11) dissecting the plateaus of Guyana and Brazilian Highlands where the shallow water table along river corridors is known to support gallery forests in otherwise dry grasslands (Whitmore and Prance, 1987; Clapperton, 1993); in the model, this is a direct result of hillslope to catchment-scale groundwater convergence, a process that hydrologists know very well.

It is seen here that a primary function of the groundwater flow system is to reorganize the land surface surplus (i.e., recharge R) according to the terrain structure at hillslope to continental scales, with respect to the sea level control on coastal drainage. This terrain structure and sea level control may over-ride climate control in many cases in maintaining a high water table, such as in the Pampas region of Argentina, the river corridors in the Cerrados of eastern Brazil, and the arid valleys and coastal zones in Peru and Chile.

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4 Validation of WTD simulations

We validate the simulations with all available observations, first over the continent, second over the Amazonia, and third at nine research sites reported in the literature.

The 34 351 site observations (Fig. 1) fall into 27 947 model cells over the continent.

5 We examine the model residual (simulated – observed head). Without systematic biases, the residual should follow a zero-mean Gaussian distribution with no dependence on climate or terrain but allowance for random deviations due to coarse grids (~274 m cell vs. well/point observations), generalization in geology (neglecting aquifer heterogeneity) and the temporal noise (observations taken at different times). Table 1 summarizes the residual statistics, and Fig. 6 plots the residual histogram (a), dependence of residual on annual precipitation (b), land elevation (c) and slope (d), from the four recharge forcings. The simulation using HTESSSEL recharge has the smallest continental mean residual (–0.55 m, i.e., simulation 0.55 m too low compared to observations); Mosaic has the largest residual (–4.36 m, simulation 4.36 m too low); CLM and NOAH both have a positive residual (1.91 m and 2.66 m too high, respectively). Because the observations likely contain a low bias due to pumping, the CLM and NOAH forced simulations may be closer to the natural water table conditions. Common among the four is a negative residual correlation with terrain slope, that is, the positive model-bias occurs mainly in flat areas, consistent to the notion of low bias in observations due to pumping
20 which occurs at flat lands where cities, industries and agriculture are found.

Validation statistics over the Amazonia (outline in Fig. 1), where 2511 grid cells have observations, is given in Table 2 and Fig. 7. The CLM-forced simulation has the smallest mean bias, but even CLM, the lowest recharge in the Amazonia (Fig. 3), produces a water table that is 2.31 m too high compared to the observations. Groundwater pumping, clustered over large metro regions in Amazonia is again thought to be the cause; large cities such as Manaus, Belém, Santarém, Rio Branco, and Boa Vista, situated on major aquifers, are partially, and the city of Vilhena is entirely, supplied by groundwater. In the state of Maranhão (capital being São Luís), 70% of the water supply for
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its cities comes from the groundwater. The fact that these observations are in dense clusters over major cities further enhances their influence on the statistics because any pumping well in the cluster would affect all the other wells nearby. The mostly eliminated groundwater recharge over urban pavements also contributes toward the low bias (http://www.ana.gov.br/pnrh_novo/documentos/01%20Disponibilidade%20e%20Demandas/VF%20DisponibilidadeDemanda.pdf). However, it is difficult to remove these effects without some degree of arbitrary manipulation of the data, and thus we use them here but recognize the biases.

We note that the large residual standard deviation in Table 1 and 2 have several causes. First, the mean water table depth over the grids of ~ 274 m are compared with well observations taken at a point, and depending on whether the point is on the hilltop or at the valley bottom, the observed water table can vary. For example, in the Ducke Reserve near Manaus (Pineda, 2008), the water table head can differ by >30 m within the span of a grid cell. Second, groundwater pumping can lower the local water table by tens of meters, rendering the observations greatly different from the natural conditions which the model attempts to simulate. Third, heterogeneity in local permeability is neglected due to the lack of actual permeability data across the continent. Instead, a uniform parameterization is applied which considers only the influence of terrain slope on soil depth. While this may work well over the continent statistically, it can misrepresent local conditions and causing the large deviation between model and observations. Fourth, while the model attempts to simulate a hydrologic equilibrium condition, the observations are taken at various times over several decades; nearly all of the sites have only one reading, that is, each observation point has a different time stamp. This introduces a large temporal noise that also causes a large deviation between model and observations. Thus the large standard deviation is a result of inherent deficiencies in the observations (point nature, groundwater pumping, and temporal noise) and the simple parameterization of permeability in the absence of actual measurements. These issues cannot be resolved by this study alone without fundamental improvement in the observational dataset, detailed characterization of local aquifers

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over the continent, and greater computational power. However, our goal is to provide a first-order, unbiased assessment of the water table position and its potential influence on land surface fluxes. The result may illustrate a need to improve groundwater observations and subsurface datasets in support of better large-scale groundwater models.

As a third step of validation, we zoom into various study sites reported in the literature and examine how well the simulated water table compares with local observations where they are available. In particular, we assess how well the model reproduces the observed spatial patterns along the topographic gradient. A search of literature found nine sites in Amazonia with WTD reports (Fig. 8). These research sites are not affected by groundwater pumping based on the descriptions given in the article. The reported WTD are compared with the simulation forced by CLM (smallest bias, lowest recharge and deepest water table in the Amazon). The results (Fig. 9) suggest that the CLM simulation is accurate in the valleys, but it is too deep under high grounds in seven out of the nine cases (Table 3).

In summary, the validations with point observations over the continent and the Amazonia suggest that the simulations in general have no systematic biases along climate and terrain gradients, but all four give a WTD that is shallower than observed in the Amazonia. It is at least partially caused by the low bias in the observations from pumping. For a conservative estimate of the water table's role, we will use the simulation from the lowest recharge in the Amazon (CLM) for the remainder of the study. However, the CLM-forced WTD, when compared with detailed observations at nine sites, appears accurate in the valleys but too deep under high grounds, which may under-estimate its potential influence on the land surface in upland ecosystems.

Based on the CLM-forced (deepest and most conservative) water table simulation, 36% of the area in Amazonia has a $WTD < 5$ m, and 60% < 10 m. This is close to the observations where 34% has a $WTD < 5$ m, and 57% < 10 m. Next we calculate the upward soil capillary flux such a WTD distribution can potentially support.

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5 Calculating upward soil capillary flux from the water table

Soil water movement occurs in 3 vertical zones (Fig. 10a). Above the water table, the soil is unsaturated, and both capillary (C , can be upward or downward) and gravity forces (G , downward) drive the flux; below the water table, the soil is fully saturated and gravity drives the lateral exchange with the neighboring area depending on the water table slope. Above the water table is a zone of saturation termed capillary fringe (or tension-saturated-zone). The physics of soil water movement in a column is described by the Richards Equation, solved in most land surface component of climate and ecosystem models. We use the Clapp-Hornberger (1978) soil water retention relation and solve the Richards Equation numerically for the two dominant soil types in the Amazonia (clay-loam and clay, Fig. 4b), to examine the influence of water table depth on shallow soil moisture. The boundary conditions are prescribed wilting point in the top layer (all layers are 0.05 m thick) and saturation at the water table (not the top of capillary fringe, to be conservative and to avoid another parameter). The resulting soil moisture profiles and capillary fluxes are shown in Fig. 10b for the two soil types. It shows that the water table keeps the shallow soils above their wilting point for all water table depths examined here (1–20 m). Upward capillary flux from the water table can be $>4 \text{ mm day}^{-1}$ with WTD $<2 \text{ m}$ but decreases rapidly as the water table drops deeper.

This simple method can be used to assess the potential contribution of this flux to Amazonian dry-season ET at locations where independent ET estimates are available. Figure 11 gives the simulated water table depth forced by CLM-recharge in the Amazonia, with details at nine sites with ET estimates. Using this WTD map as the lower boundary condition to solve the Richards Equation for all grid cells, we obtain a map of potential capillary flux (Fig. 12), aggregated from 9 arcsec (WTD grid) to 1 arcmin ($\sim 2 \text{ km}$, footprint of flux towers). We placed a cap of $<5 \text{ mm day}^{-1}$ on the flux before aggregating because it exceeds the potential evaporation rate (Fig. 10) at shallow water table depth. The $<5 \text{ mm day}^{-1}$ cap is below the maximum observed ET rate; globally,

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closed canopy tropical forests can evaporate up to 2 m year^{-1} , or (5.5 mm day^{-1}) (Gordon et al., 2005); estimates of ET from a wet forest in Costa Rica, based on two methods (Penman-Montieth and Priestly-Taylor), are 5.2 and 6.3 mm day^{-1} in 1998 and 1999, respectively (Loescher et al., 2005). Thus a cap of 5 mm day^{-1} is reasonably conservative. At the nine sites with ET estimates in Fig. 12, the capillary flux (upper number) and the ET (lower number) are given. Where the water table is shallow only in the valleys (Jarú, Guyaflux), the capillary flux over a 2 km grid at the site is $<1 \text{ mm day}^{-1}$, and where it is shallow over broader areas (Acre, Bananal Island), the flux can be $>3 \text{ mm day}^{-1}$. Weak capillarity in sandy soils in Kabu-Tonka (Poels, 1987) and Sinop (Vourlitis, 2008) reduces the water table influence. Across the nine sites, potential groundwater contribution varies from 13% to 106% of the estimated ET. At the Bananal Island site, capillary flux can exceed the total ET which is limited by atmospheric demand. Averaged over the Amazonia, the potential flux is 2.1 mm day^{-1} , a significant portion of estimated total ET (Fisher et al., 2009) of $\sim 3.8 \text{ mm day}^{-1}$.

The flux estimate above may be conservative for the following reasons. First, it is based on the water table simulation forced by CLM recharge which is the lowest among the four and gave the deepest water table. Validation suggests that it may be too deep under high grounds (Table 3); based on the other simulations (Fig. 5, Table 4, Fig. 13), the Amazon mean capillary flux is 2.2 (MOSAIC forced), 2.6 (NOAH forced), and 2.6 (HTESSEL forced) mm day^{-1} , respectively (vs. 2.1 with CLM forced). Second, it is based on equilibrium or long-term mean water table; field observations show that seasonality in water table can lag rainfall by three to four months so that the water table is the shallowest in the peak dry season (Tomasella et al., 2008; Jirka et al., 2007); that is, dry season water table and the capillary flux could be higher than the annual means. Third, the capillary fringe above the water table, a zone of full saturation reaching a few meters high in clay soils (e.g., Freeze and Cherry, 1979) is neglected. Finally, plant rooting depth is neglected in calculating the capillary flux by confining soil water uptake to the top 0.05 m only (wilting-point water content in the top layer). Consideration of the latter two factors would in effect bring the water table closer to the land surface.

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We note that the simulated water table depth and the capillary flux are not highly sensitive to recharge differences. The recharge ranged 411–839 mm yr⁻¹ over Amazonia among the four estimates (Fig. 3), with the resulting mean Amazonia WTD in the range of 9.47–7.36 m (Fig. 5) and the capillary flux of 2.1–2.6 mm day⁻¹ (Fig. 13, Table 4). This is because groundwater drainage is self-limiting (e.g., de Vries, 1994, 1995; Eltahir and Yeh, 1999; Marani et al., 2001); as recharge increases, the water table rises, steepening the hydraulic gradient from hills to valleys and expanding the channel network by groundwater seepage, both accelerating drainage and effectively bringing down the water table; as the recharge decreases, the water table falls, flattening the hydraulic gradient and bringing the water table below local streams, both reducing discharge and preserving the groundwater store. This negative feedback dampens the water table sensitivity to recharge uncertainties.

The simple analyses presented here suggest that in parts of Amazonia where the water table is shallow, capillary flux from the water table may be an important mechanism to keep the upper soil moist during the dry season when the source of water is from below only. Where the water table is deep, hydraulic redistribution via deep roots can be an important mechanism, and the presence of a water table can only enhance its significance by providing a deep water source. Our findings support the various mechanisms proposed to explain the observed lack of vegetation stress in the dry season (see Introduction); a water table and the upward capillary flux increases total soil water store (first mechanism) and provides deep water source for root and capillary actions (second to fifth). We emphasize that plant roots do not need to physically reach the water table to benefit from it; capillary flux from the water table can send water upward many meters to meet the roots in fine-textured soils.

The results may have important implications to modeling terrestrial water, energy, and carbon cycles. Groundwater is not yet routinely included in climate and land ecosystem models, but in the Amazon, the groundwater store may affect the land

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surface where it is within the reach of soil capillary action, buffering the ecosystem in the dry season. In terms of the physical climate simulated by models, inclusion of the water table has been shown to increase and stabilize soil moisture and ET, cooling the surface and enhancing local convective and downwind precipitation (York et al., 2002; Liang et al., 2003; Yeh and Eltahir, 2005; Yu et al., 2006; Niu et al., 2007; Maxwell et al., 2007; Fan et al., 2007; Miguez-Macho et al., 2007, 2008; Anyah et al., 2008; Yuan et al., 2009; Jiang et al., 2009; Yeh and Famiglietti, 2009). Current models tend to simulate a dry-season Amazon that is too dry and too hot (Kleidon and Heimann, 2000; Werth and Avissar, 2004; Delworth et al., 2006; Huete et al., 2006; Myneni et al., 2007; Juarez et al., 2007; Baker et al., 2008) and a large-scale dieback under projected future climate with a longer dry season (Cox et al., 2004), converting the Amazon from a net carbon sink to a net source and accelerating warming. Without the groundwater buffer below, the model land is wetted from above by rainfall only, which may cause it to be overly sensitive to rainfall shortages.

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Table 1. Comparison of biases in the simulated WTD using four recharge estimates, for the continental validation.

Model (no. layers, soil depth)	Residual histogram			Correlation with precipitation	Correlation with land elevation	Correlation with terrain slope
	Shift	Stand. dev.	Skew			
HTESSEL (4, 1.89 m)	-0.55	21.35	+0.10	+0.2458	-0.1282	-0.2800
CLM (10, 3.44 m)	+1.91	18.19	+1.10	+0.0370	-0.0530	-0.3681
MOSAIC (3, 3.50 m)	-4.36	22.38	-0.34	+0.2447	-0.1376	-0.2814
NOAH (4, 2.0 m)	+2.66	18.19	+1.11	+0.0846	-0.0592	-0.3484

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Table 2. Same as Table 1, but over the Amazonian ecosystem (outline in Fig. 1).

Model (no. layers, soil depth)	Residual histogram			Correlation with precipitation	Correlation with land elevation	Correlation with terrain slope
	Shift	Stand. dev.	Skew			
HTESSEL (4, 1.89 m)	+4.36	8.59	+1.63	−0.1073	+0.0902	−0.0955
CLM (10, 3.44 m)	+2.31	8.71	+1.52	−0.0779	+0.0454	−0.1827
MOSAIC (3, 3.50 m)	+2.75	8.80	+1.37	−0.0040	−0.0076	−0.1755
NOAH (4, 2.0 m)	+4.11	8.65	+1.74	−0.0825	+0.0679	−0.1136

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Table 3. Information on the 9 sites for detailed validation and comparison between observed and simulated WTD (forced by CLM).

Site and Source	Location		Annual <i>P</i> (mm)	Soil	Vegetation	Observed WTD (m)			Simulated WTD* (m)			
	Latitude °	Longitude °				Elevation, m	valley	midslope	plateau	valley	midslope	plateau
Kabo/Tonka (Poels, 1987)	5.29	-55.65	30–50	2116	sandy loam	dense evergreen forest	0–1	2–5	4–8	0–2.5	2.5–10	10–20
White Sand (Bongers et al., 1985; Coomes and Grubb, 1996)	2.49	-66.14	99–105	2600–3600	bleached sand	palm forest to short caatinga	0–0.4	0–0.9	0.4–1	0–2.5	2.5–5	5–10
Manaus (Pineda, 2008; Hodnett et al., 1997a, 1997b)	-3.13	-60.12	30–120	1800–2800	clay loam	dense evergreen forest	0–0.5	0.5–5	20–40	0–2.5	2.5–20	20–40
Satarem (Nepstad et al., 2002)	-2.897	-54.952	~160	2000	Oxisol rich in kaolinite clay	dense evergreen forest			>12	0–2.5	2.5–10	10–40
Redenção (Grogan and Galvao, 2006)	-7.83	-50.27	220–250	1700–1900	varied sand shallow rock	transitional evergreen forest	0–4	0–8	2–10	0–2.5	2.5–5	5–20
Acre (Selhorst et al., 2002)	-10.0831	-67.6236	~220	1640	Oxisol	evergreen forest		3–9	6–10	0–2.5	2.5–10	10–20
Bananal Island (Borma et al., 2009)	-9.82	-50.15	170–182	1656	hydromorphic sand	flooded savanna and forest	0–2	0–2	0.6–4.5	0–2.5	0–2.5	2.5–10
Juruena (Jirka et al., 2007)	-10.50	-58.50	240–280	2200	clay loam	ecotone rainforest –savanna	0–1	0.5–3	3–7	0–2.5	2.5–10	10–40
Sinop (Vourlitis et al., 2008)	-11.4125	-55.325	335	1857	quartzarenic neosol-sandy	tropical semi-deciduous forest			3–3.5	0–2.5	2.5–10	10–20

* Corresponds to the center area of the images in Fig. 9.

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Groundwater and
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G. Miguez-Macho**Table 4.** WTD and capillary flux (mm day^{-1}) over the flux-tower footprint (~ 2 km) from four simulations at nine sites with independent ET estimates and the mean over the Amazonia.

Site	CLM ($R=411$ mm)		MOSAIC ($R=476$ mm)		NOAH ($R=787$ mm)		HTESSEL ($R=839$ mm)		Independent ET Estimate (mm day^{-1})
	WTD (m)	Flux (mm day^{-1})	WTD (m)	Flux (mm day^{-1})	WTD (m)	Flux (mm day^{-1})	WTD (m)	Flux (mm day^{-1})	
Acre	5.59	3.1	6.49	2.6	3.99	3.6	5.07	3.3	3.0
Bananal Island	2.00	3.7	2.43	3.1	1.56	4.0	1.52	4.0	3.5
Caxiuana	11.44	1.5	10.71	1.5	8.44	2.0	7.95	2.1	2.9
Guyaflex	10.38	0.8	8.64	1.1	6.19	1.5	8.02	1.2	6.0
Jaru	15.43	0.6	16.71	0.5	10.67	0.9	9.56	1.0	2.7
Kabu-Tonka	5.83	2.0	5.91	2.0	4.74	2.2	4.01	2.4	5.1
Manaus (K34)	14.69	1.8	13.62	1.9	11.92	2.1	11.37	2.2	3.2
Santarem (K67)	8.83	1.2	9.60	1.0	6.86	1.5	5.45	1.9	3.0
Sinop	11.58	1.1	13.23	1.0	9.63	1.2	8.83	1.3	2.5
Amazonia Mean	9.16	2.1	9.47	2.2	7.37	2.6	7.36	2.6	3.8

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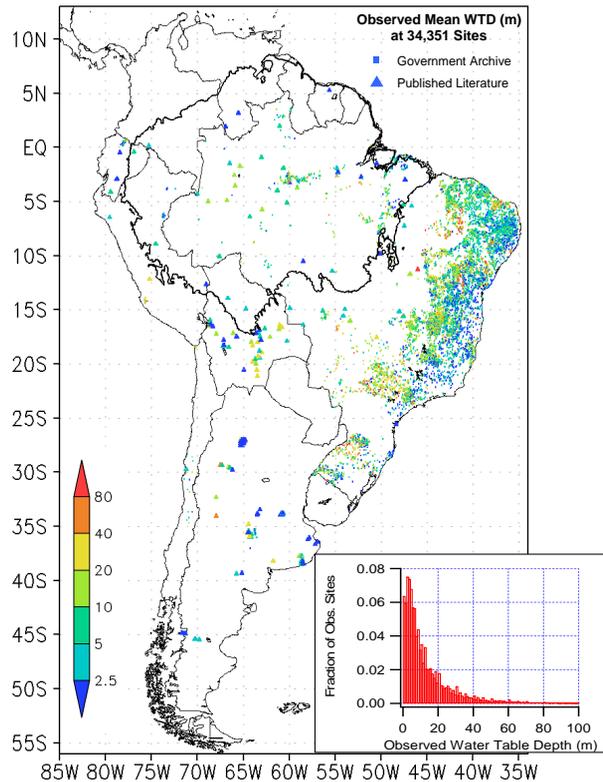
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Fig. 1. Observations of water table depth, WTD (m), compiled from government archives and published literature, with the outline of Amazonia ecosystem (inset gives the histogram of WTD).

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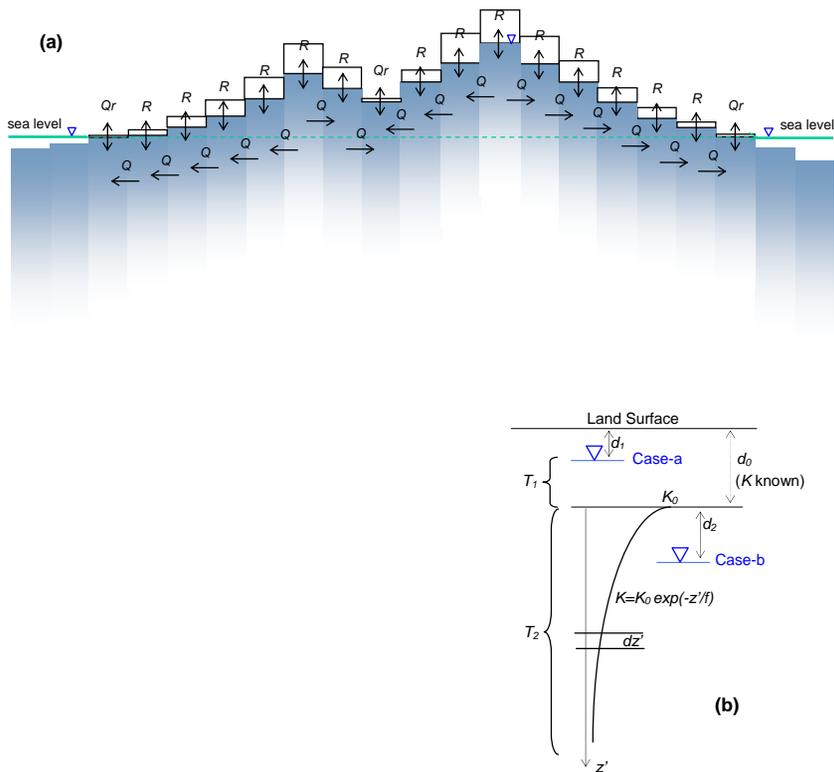


Fig. 2. (a) Schematic of the 2-D groundwater model to simulate the climate (recharge R), terrain (lateral flow Q) and sea level (boundary condition) control on water table depth over a continent. In upland cells, recharge balances lateral groundwater divergence to lower neighbors. In valley or coastal cells, lateral groundwater convergence discharges into wetlands and rivers, **(b)** details of calculating flow transmissivity, T , for case-a, water table within the depth of known soil hydraulic conductivity (K), and case-b, water table below the known depth where K is assumed to decrease exponentially with depth.

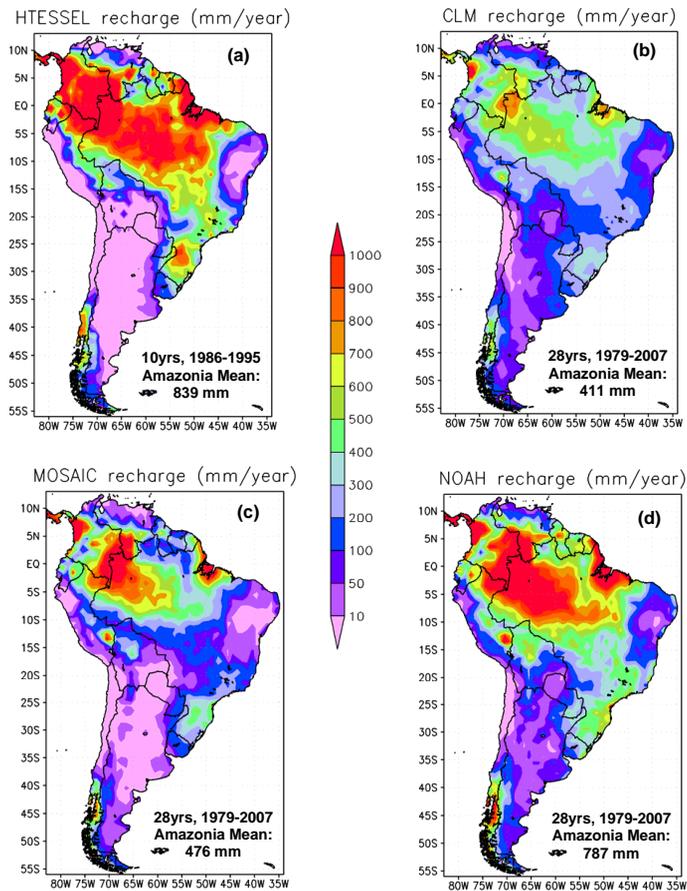


Fig. 3. Mean annual water table recharge (mm yr^{-1}) simulated by four global land surface models: (a) HTESSEL (Balsamo et al., 2009, data from B. van den Hurk), (b) CLM, (c) MOSAIC, and (d) NOAH (last three from GLADS, Rodell et al., 2004).

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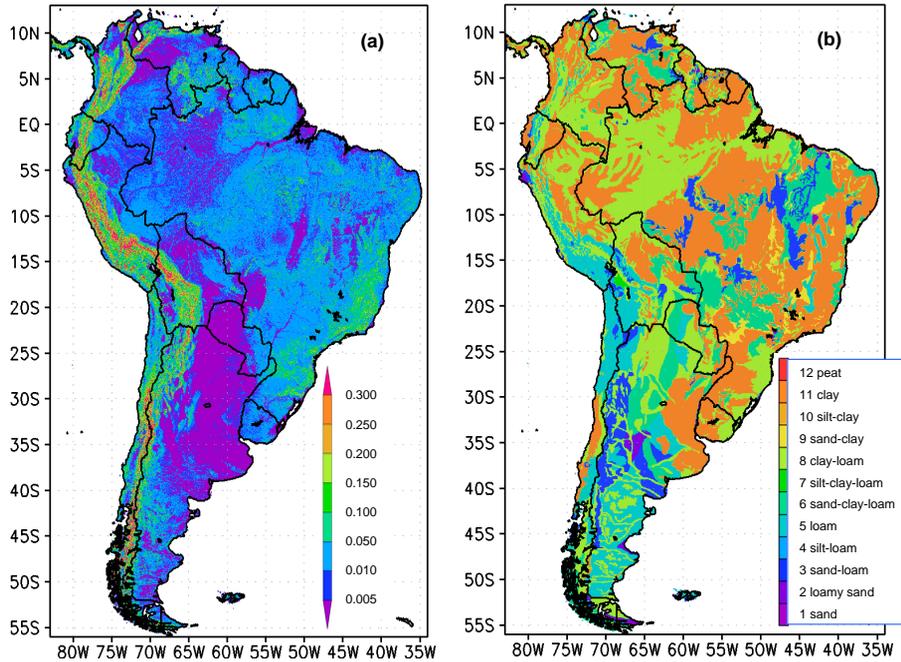


Fig. 4. (a) Terrain slope at 9 arcsec resolution, and (b) soil texture classes from GLDAS (Rodell et al., 2004).

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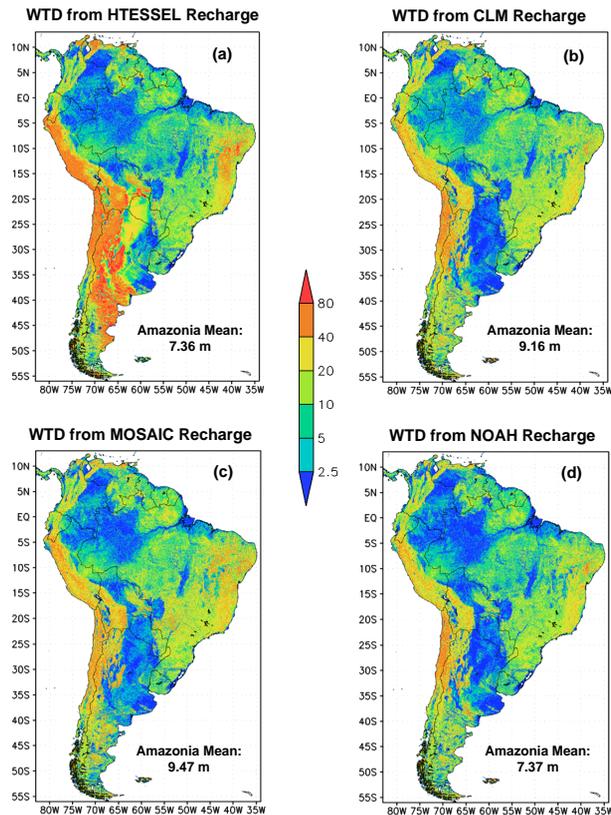


Fig. 5. Simulated equilibrium water table depth (m) forced by recharge from (a) HTESSEL, (b) CLM, (c) MOSAIC, and (d) NOAH.

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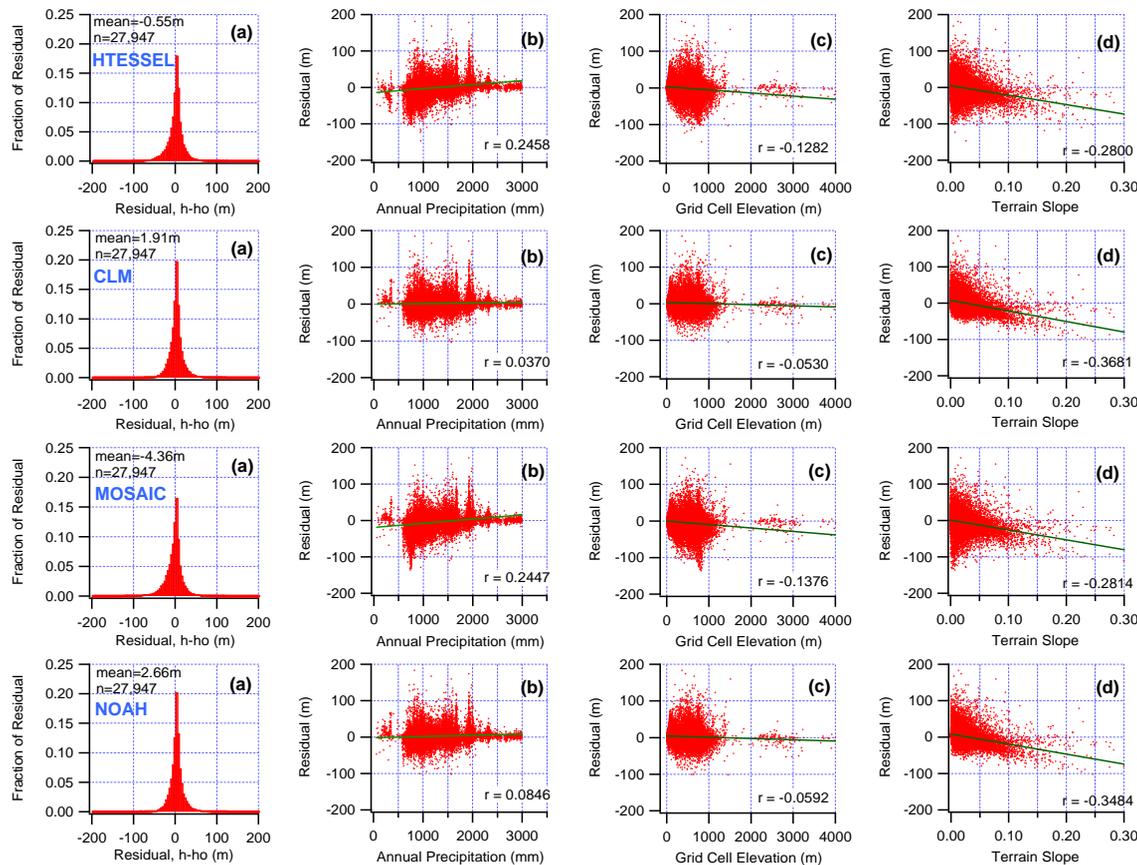


Fig. 6. Residual statistics for the four simulations: **(a)** histogram of the residual (modeled head – observed head), **(b)** residual vs. annual precipitation, **(c)** vs. land elevation, and **(d)** vs. terrain slope, with the Pearson correlation coefficient (r) given.

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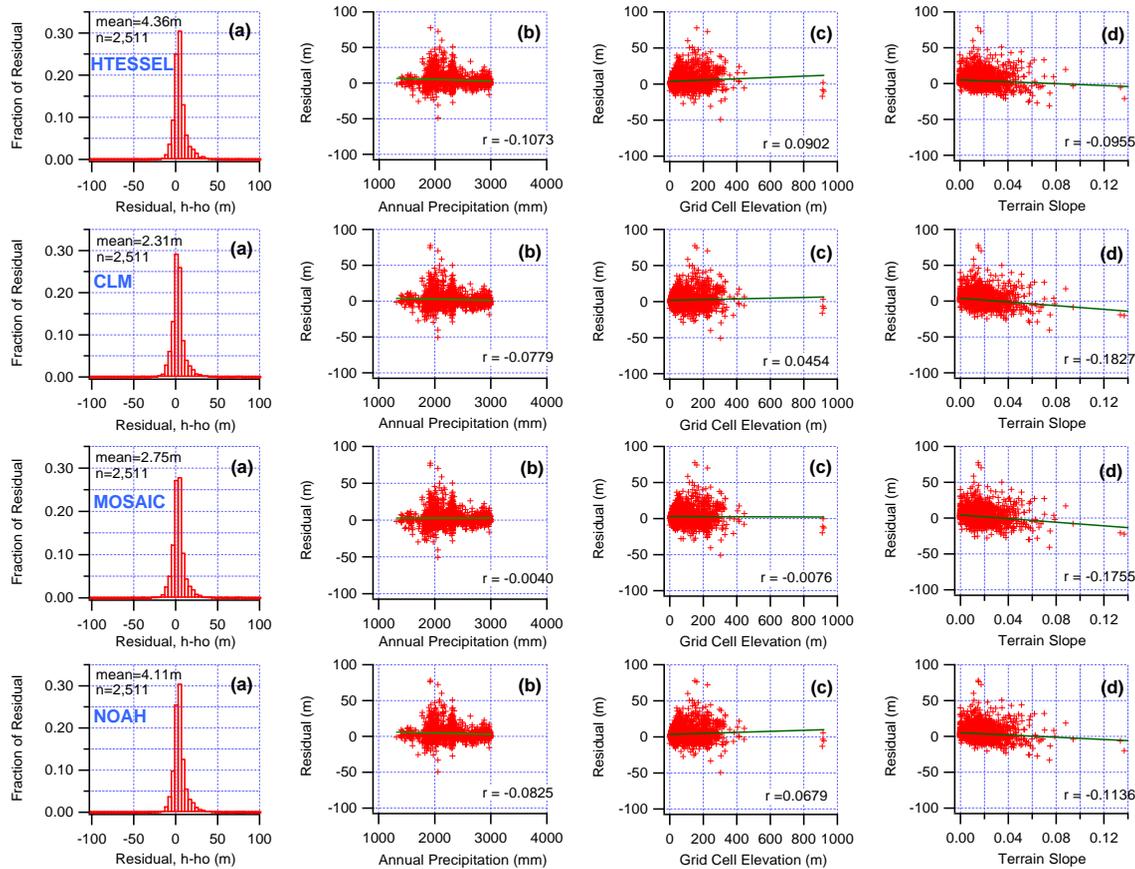


Fig. 7. Same as Fig. 6, but for validation over the Amazonia (outline shown in Fig. 1).



Fig. 8. Sites of detailed validation using water table observations from published literature (base-map: vegetation index from R. Simmon, NASA Earth Observatory).

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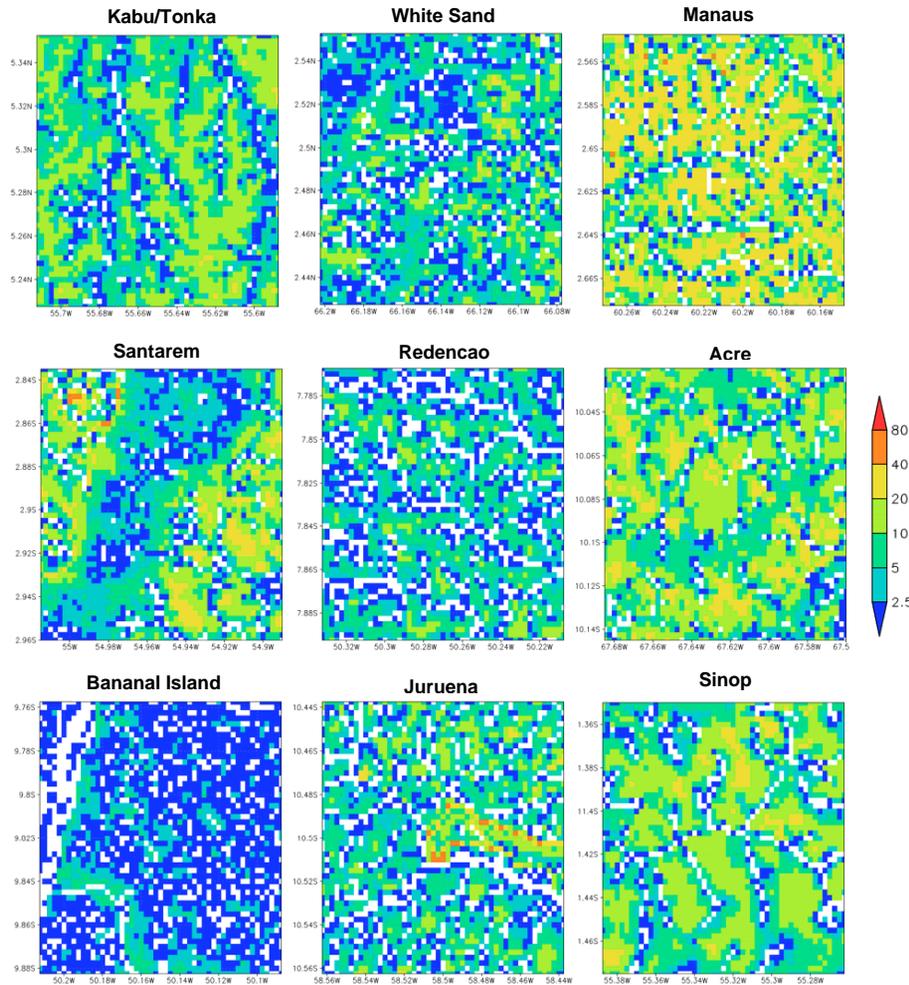


Fig. 9. Simulated WTD (m) surrounding the nine sites (Fig. 8) with WTD observations (white cell=WTD at land surface).

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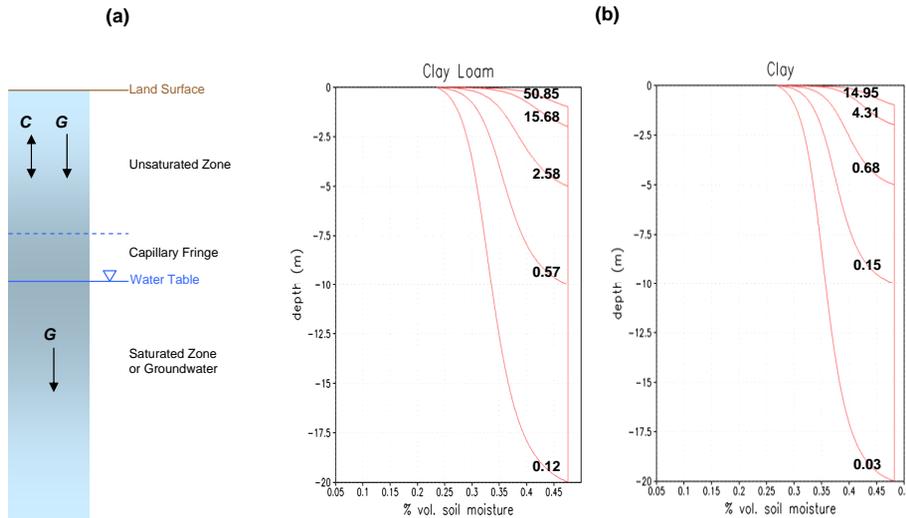


Fig. 10. (a) Typical soil water zones. Upward and downward capillary fluxes (*C*) and downward gravity flux (*G*) drive soil water movement above the water table, and gravity (*G*) drives the flow below the water table. (b) Soil moisture profile and upward capillary flux (mm day^{-1}) calculated from the Richards Equation, with wilting point prescribed in the top 0.05 m and saturation at the water table of various depths (1, 2, 5, 10, and 20 m), for the two most abundant soil types in the Amazon (clay-loam and clay).

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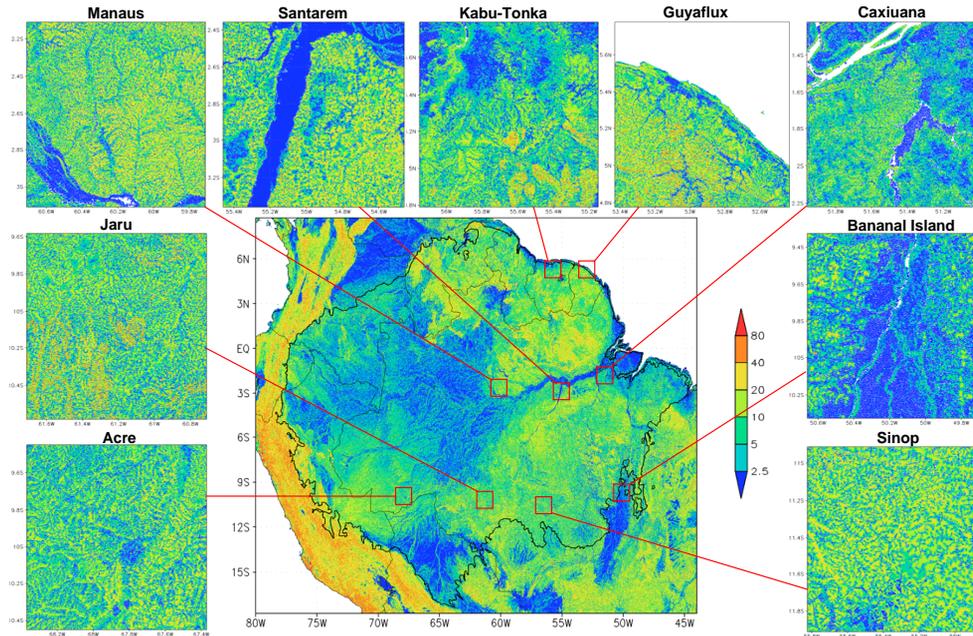


Fig. 11. Spatial details in the CLM-forced WTD simulation (m) in Amazonia and around the nine sites with independent ET estimates (white cells=WTD at land surface).

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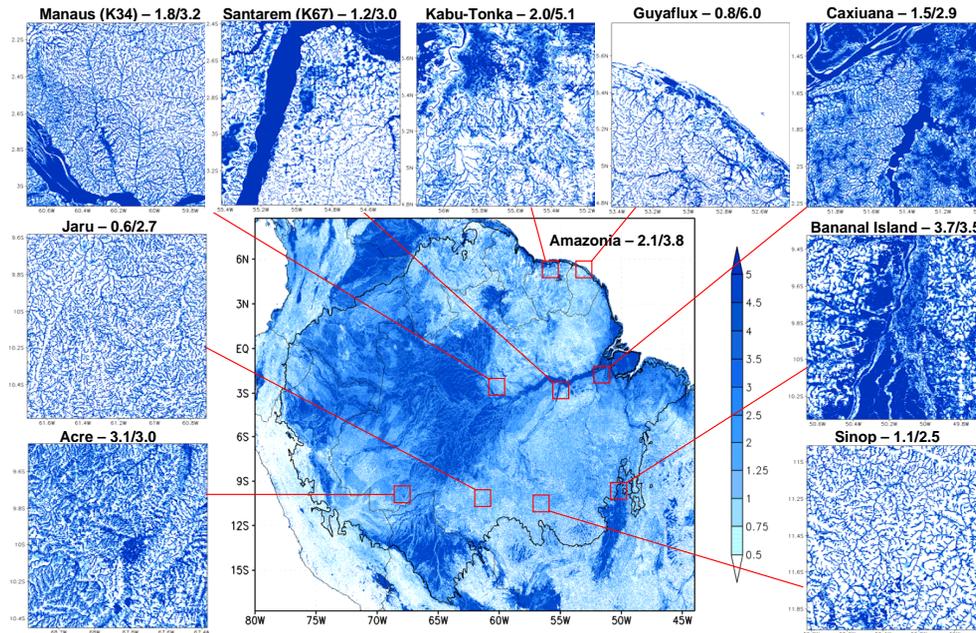


Fig. 12. Calculated capillary flux (mm day^{-1}) from the CLM-forced water table simulation in the Amazonia, with details around nine sites with ET estimates. The numbers given are capillary flux from the water table over total ET, both in mm day^{-1} . ET source: Manaus, Santarem, Caxiuana, Jaru and Sinop (Juarez et al., 2007), Bananal Island (Borma et al., 2009), Guyaflex and Amazonian mean (Fisher et al., 2009), Kabu-Tonka (Poels, 1987), Acre (Duarte et al., 2008).

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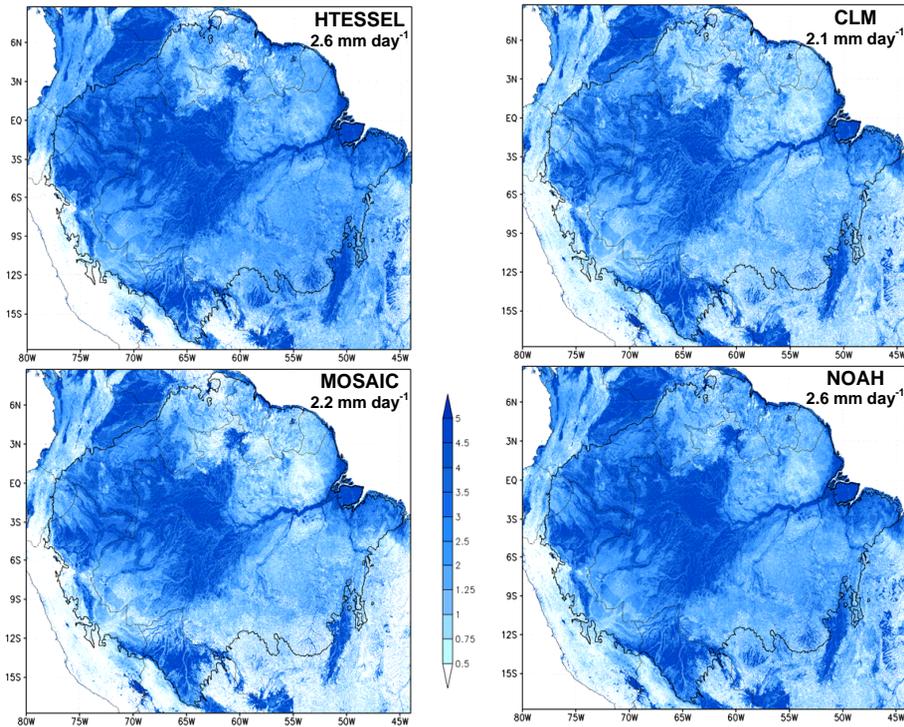


Fig. 13. Maps of capillary flux (mm day^{-1}), showing Amazonian mean, based on the four WTD simulations.

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