The Partitioning of Evapotranspiration into Transpiration, Soil Evaporation, and Canopy Evaporation in a GCM: Impacts on Land–Atmosphere Interaction

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ABSTRACT

Although the global partitioning of evapotranspiration (ET) into transpiration, soil evaporation, and canopy evaporation is not well known, most current land surface schemes and the few available observations indicate that transpiration is the dominant component on the global scale, followed by soil evaporation and canopy evaporation. The Community Land Model version 3 (CLM3), however, does not reflect this global view of ET partitioning, with soil evaporation and canopy evaporation far outweighing transpiration. One consequence of this unrealistic ET partitioning in CLM3 is that photosynthesis, which is linked to transpiration through stomatal conductance, is significantly underestimated on a global basis. A number of modifications to CLM3 vegetation and soil hydrology parameterizations are described that improve ET partitioning and reduce an apparent dry soil bias in CLM3. The modifications reduce canopy interception and evaporation, reduce soil moisture stress on transpiration, increase transpiration through a more realistic canopy integration scheme, reduce within-canopy soil evaporation, eliminate lateral drainage of soil water, increase infiltration of water into the soil, and increase the vertical redistribution of soil water. The partitioning of ET is improved, with notable increases seen in transpiration (13%–41% of global ET) and photosynthesis (65–148 Pg C yr\(^{-1}\)). Soils are wetter and exhibit a far more distinct soil moisture annual cycle and greater interseasonal soil water storage, which permits plants to sustain transpiration through the dry season.

The broader influences of improved ET partitioning on land–atmosphere interaction are diverse. Stronger transpiration and reduced canopy evaporation yield an extended ET response to rain events and a shift in the precipitation distribution toward more frequent small- to medium-size rain events. Soil moisture memory time scales decrease particularly at deeper soil levels. Subsurface soil moisture exerts a slightly greater influence on precipitation. These results indicate that partitioning of ET is an important responsibility for land surface schemes, a responsibility that will gain in relevance as GCMs evolve to incorporate ever more complex treatments of the earth’s carbon and hydrologic cycles.

1. Introduction

One of the primary roles of a land surface scheme in a global climate model (GCM) is to partition surface available energy into latent and sensible heat fluxes. Determination of latent heat flux, which is directly proportional to total evapotranspiration (ET), is more involved than for sensible heat flux since it is the sum of transpiration (\(E_T\)), soil evaporation (\(E_S\)), and canopy evaporation (\(E_C\)). The relative contribution of each component of ET is calculated within the land surface scheme relative to potential evaporation according to independent resistances to evaporation for each component due to turbulent transfer, moisture limitations, and, in the case of transpiration, stomatal physiology.

Global- and regional-scale partitioning of ET is not accurately known since large-scale observations of ET, let alone its partitioning, are simply not available. Perhaps as a consequence, few researchers have diagnosed ET partitioning in land surface schemes, even though it would provide a powerful constraint on model physics. Choudhury and DiGirolamo (1998) and Choudhury et al. (1998) use a biophysical process-based model forced with observed precipitation, air temperature, net radiation, and vapor pressure deficit to derive global climatological quantities for ET and ET partitioning. They
find that $E_T$ accounts globally for about 52% of total ET while $E_S$ contributes 28% and $E_C$ the remaining 20%. A more recent estimate of global ET partitioning is obtained from multimodel output from the Global Soil Wetness Project 2 (GSWP2; Dirmeyer et al. 2005). In GSWP2, the multimodel mean estimate of global ET partitioning is 48% $E_T$, 36% $E_S$, and 16% $E_C$.

Strict validation of a global land surface model’s ET partitioning is not possible because field measurements of ET partitioning are very sparse. Transpiration and soil evaporation can be isolated either through a combination of stable isotope, sap flow, and eddy covariance techniques (Williams et al. 2004) or with porometer, lysimeter, and Bowen ratio techniques, although errors in ET partitioning calculated by any of these methods may be large (Herbst et al. 1996). A literature survey reveals only a limited number of studies that report ET partitioning. It is difficult to extract clear and useful guidance for GCM development from these disparate studies, although it is reasonable to interpret that transpiration is the dominant component of ET across a variety of ecologically based systems (Wallace et al. 1993; Ashktorab et al. 1994; Leuning et al. 1994; Black et al. 1996; Wilson et al. 2001; Ferretti et al. 2003; Williams et al. 2004). However, when the canopy is sparse, the partitioning is more complex (Baldocchi et al. 2004) and soil evaporation can at times dominate (Allen 1990; Yunusa et al. 1997).

By comparison to other models, and the limited observations, the Community Land Model version 3 (CLM3) partitions ET in an unrealistic manner. Averaged over the global land surface, ET partitioning in CLM3, when forced by observed meteorology, is 13% $E_T$, 44% $E_S$, and 43% $E_C$ (Fig. 1 shows the geographic distribution of ET partitioning). The partitioning is just as unrealistic when CLM3 is coupled to the Community Atmosphere Model version 3 (CAM3) (11% $E_T$, 57% $E_S$, and 32% $E_C$). This unrealistic partitioning of ET is likely to affect both mean ET and its temporal evolution. Since the time scales of response differ for each ET component (fast for $E_C$, slower for $E_S$, and slowest for $E_T$), the time scale of ET response and the local climate response, such as to a rainfall event or a seasonal precipitation anomaly, is likely to be affected by how CLM3 executes the partitioning. Weak $E_T$ also may affect the amplitude and regionality of CAM3–CLM3 land–atmosphere coupling. Transpiration by plants is also critically important due to its interactive role in photosynthetic gas exchange and productivity. As an example, when CLM3 is coupled to a dynamic global vegetation model, weak $E_T$ spawns conversion of forests to grassland in areas where forests should dominate, such as in Amazonia and the southeastern United States (Bonan and Levis 2006). Low photosynthesis during the Amazonian dry season prevents establishment of a broadleaf evergreen forest; instead a decidu-
ous forest dominates. For all of these reasons, a substantial improvement to ET partitioning is important within the context of ongoing development goals of the Community Climate System Model version 3 (CCSM3) with respect to modeling the hydrologic cycle (Hack et al. 2006) and the carbon cycle (Thornton and Zimmerman 2007).

In this paper we describe a series of simple modifications to CLM3 hydrology and vegetation parameterizations that together substantially improve the partitioning of ET in both offline and coupled configurations through reduced canopy interception, increased soil moisture availability for transpiration, and increased soil moisture storage. We also evaluate how the partitioning of ET affects climate simulations with an emphasis on how ET partitioning influences land–atmosphere interactions. It should be noted that this research has been conducted within the context of a longer-term CLM community-oriented project that is focused on a thorough reconsideration, evaluation, and development of CLM hydrology. The modifications documented here represent significant practical improvements to the released version of CLM3, thereby providing an interim model version that can be used by researchers and model developers while a new hydrology scheme for CLM is developed.

2. Model description

We use CLM3 in both offline mode and coupled to CAM3 (Collins et al. 2006). CLM3 is described in detail in Oleson et al. (2004) and its performance is documented in Dickinson et al. (2006). Subgrid-scale surface-type heterogeneity is represented in CLM3 through satellite-derived fractional coverage of lakes, wetland, bare soil, glacier, and vegetation consisting of up to four plant functional types in each grid box. Fluxes of energy and moisture are modeled independently for each surface type and aggregated before being passed to the atmosphere model. The four plant functional types share a single soil column, with 10 soil layers extending to 3.43-m depth (layer depths increase exponentially from 0.0175 m at the top to 1.14 m at the bottom). Soil moisture heterogeneity is simply represented through differing runoff formulations for the saturated and unsaturated fractions of the soil column, which are dynamically defined according to water table depth. Further details of CLM3 directly relevant to the ET partitioning are described in the following section.

3. Modifications to model physics

In this section we describe a series of modifications to CLM3 model physics and parameterizations that result in a more reasonable partitioning of ET, as well as wetter soils and greater interseasonal soil water storage. The use of the somewhat ambiguous term “more reasonable” is deliberate. As noted in the introduction, strict validation of a global model’s ET partitioning is not possible because the required observations do not exist. The few local-scale observations of ET partitioning can, at best, provide limited guidance on the relative importance of $E_T$, $E_S$, and $E_C$ for various ecosystems. Hence, we consider the multimodel global offline results provided in GSWP2 (Dirmeyer et al. 2005) to be the scientific communities’ current best estimate of ET partitioning, and we use these values as a soft target to partially direct the model changes. To further constrain the modifications to CLM3, we require that the runoff simulation, for which there is validation data, is not degraded. Note again that although the changes outlined here, to the extent possible, are physically realistic and defensible solutions to correct known deficiencies in model behavior, the changes are also motivated strongly by an urgent need for a more reasonable representation of ET partitioning and soil hydrology that will permit ongoing hydrologic and carbon cycle model development while a more thorough reconsideration of CLM hydrology is undertaken.

The changes are presented as serial modifications to the original model (CONTROL) and are tested and evaluated for their impact on ET partitioning and runoff in offline mode. Each model configuration is run for 15 yr (1987–2001), forced with observed precipitation, temperature, downward solar and longwave radiation, surface wind speed, specific humidity, and air pressure from a recently updated and improved historical forcing dataset described in Qian et al. (2006). The first 10 yr are devoted to spinup of soil moisture and soil temperature and the remaining 5 yr are used to form climatological annual and seasonal averages of key model diagnostics. Annual global mean results from the sequential experiments described below are cataloged in Table 1.

a. Canopy interception

Clearly, the most egregious bias in ET partitioning is the excessively large contribution from $E_C$. A high canopy interception rate and high $E_C$ means that much of the atmospheric demand for evaporation is satisfied by $E_C$, thereby limiting transpiration. Furthermore, in CLM3 plants can only transpire from the dry portion of the canopy, which tends to be a very small part of the canopy during and following a rain event. The canopy interception rate $q_{\text{int}}$ in CLM3 is defined as

$$q_{\text{int}} = \alpha_i(q_{\text{rain}} + q_{\text{smo}})[1 - \exp(-0.5(LAI + SAI))]$$

(1)
Global and annual mean ET and runoff for each of the series of experiments described in section 3. Also shown are two independent estimates of ET partitioning. Global observed runoff estimates are 0.82 mm day\(^{-1}\) (Fekete et al. 2002). \(Q_{\text{DRAIN}}\) is drainage or subsurface runoff, and \(Q_{\text{TOT}}\) includes runoff from glaciers, wetlands, and lakes as well as surface and subsurface runoff.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>(E_T) (%, mm day(^{-1}))</th>
<th>(E_S) (%, mm day(^{-1}))</th>
<th>(E_C) (%, mm day(^{-1}))</th>
<th>ET (mm day(^{-1}))</th>
<th>(Q_{\text{SURF}}) (mm day(^{-1}))</th>
<th>(Q_{\text{DRAIN}}) (mm day(^{-1}))</th>
<th>(Q_{\text{TOT}}) (mm day(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>GSWP2 (Dirmeyer et al. 2005)</td>
<td>48.064</td>
<td>36.048</td>
<td>16.022</td>
<td>1.34</td>
<td>0.32</td>
<td>0.63</td>
<td>0.95</td>
</tr>
<tr>
<td>CONTROL(_{\text{surf}})</td>
<td>13.017</td>
<td>44.057</td>
<td>43.056</td>
<td>1.31</td>
<td>0.34</td>
<td>0.28</td>
<td>0.74</td>
</tr>
<tr>
<td>+ Canopy interception</td>
<td>25.027</td>
<td>56.062</td>
<td>18.020</td>
<td>1.09</td>
<td>0.45</td>
<td>0.39</td>
<td>0.95</td>
</tr>
<tr>
<td>reduced</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>+ Two-leaf model</td>
<td>28.032</td>
<td>55.064</td>
<td>18.020</td>
<td>1.16</td>
<td>0.44</td>
<td>0.34</td>
<td>0.89</td>
</tr>
<tr>
<td>+ Soil water availability</td>
<td>33.038</td>
<td>50.058</td>
<td>17.020</td>
<td>1.16</td>
<td>0.38</td>
<td>0.25</td>
<td>0.88</td>
</tr>
<tr>
<td>to LSM (\beta_s)</td>
<td>38.044</td>
<td>44.051</td>
<td>18.020</td>
<td>1.15</td>
<td>0.43</td>
<td>0.34</td>
<td>0.89</td>
</tr>
<tr>
<td>+ Soil evaporation</td>
<td>41.052</td>
<td>42.053</td>
<td>17.20</td>
<td>1.27</td>
<td>0.30</td>
<td>0.35</td>
<td>0.77</td>
</tr>
<tr>
<td>beneath canopy</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>reduced (\text{VEGHYD}_{\text{surf}})</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>+ Soil hydrology</td>
<td>44.056</td>
<td>39.050</td>
<td>17.022</td>
<td>1.29</td>
<td>0.28</td>
<td>0.35</td>
<td>0.75</td>
</tr>
</tbody>
</table>

where \(q_{\text{rain}}\) and \(q_{\text{snow}}\) are the incident rain- and snow water rates, LAI and SAI are the leaf and stem area indexes, and \(\alpha\) is a scaling factor that reflects the total fractional area of a leaf that collects water. We reduce canopy interception, and therefore canopy evaporation, by reducing the tuning parameter \(\alpha\) from 1.0 to 0.25, a value that more realistically reflects that only one side of a leaf can collect water and that rainwater tends to bead on the leaf and does not typically wet the entire exposed leaf surface. Global mean \(E_C\) drops to 18% of total ET while \(E_T\) increases to 25% and \(E_S\) rises to 56% following this change (Table 1).

### b. Two-leaf model

A new model for canopy integration, developed in Thornton and Zimmerman (2007), is introduced in which canopy leaf area is decomposed into two fractions, sunlit and shaded. While the shaded portion of the canopy is essentially inactive in CLM3, in the new model, shaded leaves receive a fraction of the diffuse radiation and are therefore permitted to photosynthesize and transpire, although at a lower rate than the sunlit portion of the leaves. This modification represents an obvious improvement in the physical realism of the model and significantly increases canopy transpiration rates. The transpiration \(E_T\) rises slightly following this change (Table 1).

### c. Soil water availability

The soil water availability factor (\(\beta_s\)) relates soil moisture conditions to soil moisture stress on plant photosynthesis and transpiration. The factor \(\beta_s\) is a function of soil water matric potential and root distribution and ranges between 0 (at wilting point potential) and 1 (at saturated soil matric potential) (Oleson et al. 2004). All other factors being equal, low values of \(\beta_s\) act to decrease the rate of photosynthesis, stomatal conductance, and \(E_T\). Here, we revert from the CLM3 parameterization of soil water availability to that of the National Center for Atmospheric Research Land Surface Model (NCAR LSM), which ranges linearly with soil water from wilting point \((\beta_s = 0)\) to maximum \((\beta_s = 1\) (Bonan 1996). For particular soil moisture conditions, the LSM representation yields \(\beta_s\) values that are greater than or equal to that using the standard CLM3 representation, a difference that increases with drier soils (Bonan et al. 2002). This change effectively permits plants to more readily access soil water and leads to a global mean increase in \(\beta_s\) of about 0.05, which contributes to a further rise in \(E_T\) to 33% of total ET (Table 1).

### d. Soil evaporation from beneath a canopy

Although the alterations introduced to this point improve ET partitioning, the fraction derived from \(E_S\) remains high. On the global scale, high \(E_S\) relative to \(E_T\) may be related to CLM3’s dry bias in deep soil layers, which results in high soil moisture stress, even in regions where heavy rainfall should provide ample water for transpiration, for example, in the southeastern United States and Amazonia. Improvements to CLM3 soil hydrology are addressed below. The high \(E_S\) contribution relative to \(E_T\), however, cannot fully be attributed to the dry soil bias. Figure 2 shows \(E_T/(E_T + E_S)\)
as a function of LAI for saturated soil conditions for a simulation that includes all the modifications introduced to this point. The relationship between \( E_T/(E_T + E_S) \) and LAI is nearly linear. Here \( E_T \) accounts for 50% of \( E_T + E_S \) at LAI \( \approx 2.5 \) and reaches 80% at LAI \( \approx 5 \). This behavior is in contrast with that found by Schulze et al. (1994), who show that the \( E_T/(E_T + E_S) \) versus LAI relationship is nonlinear with the contribution from \( E_T \) rising sharply at low LAI, hitting 50% at LAI \( \approx 1 \), before reaching around 90% at LAI \( \approx 5 \). The discrepancy between the Schulze et al. analysis and CLM3 behavior suggests that \( E_S \) from beneath a canopy is excessive in CLM3. This interpretation is supported, for example, by observations at an aspen forest site in Saskatchewan, Canada, where only 5% of annual ET comes from soil evaporation (Black et al. 1996).

The relative contribution of \( E_T \) and \( E_S \) in vegetated regions can be altered by adjusting two parameters in the equation for the turbulent transfer coefficient between the soil and the canopy air (\( C_S \)). In CLM3, \( C_S \) is defined as (Zeng et al. 2005)

\[
C_S = C_{S,\text{bare}}W + C_{S,\text{dense}}(1 - W),
\]

where the weight \( W \) is

\[
W = e^{-\alpha(\text{LAI} + \text{SAI})}.
\]

\( C_{S,\text{bare}} \) is the turbulent transfer coefficient for bare soil and is a function of the wind velocity incident on the leaves, \( C_{S,\text{dense}} = 0.004 \) and is the value used in CLM3 for a dense canopy, and \( \alpha \) is set to 1 in CLM3. This formulation for the turbulent transfer coefficient calculation is derived in Zeng et al. (2005) and was motivated by the need to reduce a large CLM2 surface air temperature bias seen in low-LAI, semiarid regions. Here, we adjust \( C_{S,\text{dense}} \) to 0.0025 and \( \alpha \) to 2, parameter values that generate an \( E_T/(E_T + E_S) \) versus LAI relationship that is in better agreement with Schulze et al., further improving global ET partitioning by reducing \( E_S \) (38% \( E_T \), 44% \( E_S \), and 18% \( E_C \); Table 1) while maintaining reasonable surface air temperatures. Note that we tested with a number of parameter values for \( C_{S,\text{dense}} \) and \( \alpha \). Although somewhat larger changes to these parameters result in a better agreement with Schulze et al., they also generate undesirably large changes in surface air temperature.

e. Soil hydrology

The preceding changes to the model significantly improve the partitioning of ET, but global fractional \( E_T \) remains low compared to other model estimates. CLM3 is characterized by dry deep soil layers and a small amplitude soil moisture annual cycle that result in excessive soil moisture stress on transpiration, especially during “dry” seasons. This unrealistic behavior is illustrated in Fig. 3, which shows mean Amazon basin annual cycle time series for a number of hydrologic cycle quantities for the CONTROL \_off run and a simulation that includes all the changes introduced to this point, hereafter referred to as VEG \_off for vegetation scheme-related changes (note that results from the experiment VEGHYD \_off that are also shown in Fig. 3 are described and discussed below). Observational and model esti-
mates indicate that the ET annual cycle in the Amazon basin is relatively flat, with slightly higher values seen during the wet season and slightly lower values seen toward the end of the dry season but not varying substantially across the year (Nobre et al. 1996; Malhi et al. 2002; Werth and Avisar 2004). The ET annual cycle in CONTROL_{off} exhibits what appears to be an unrealistically large annual cycle. High ET during the rainy season is reduced in VEG_{off} due to a reduction in $E_C$ that is not fully compensated for by increased $E_T$. In both simulations, the small soil moisture store is quickly depleted, leading to high soil moisture stress on $E_T$ and low total ET and photosynthesis from June to September. Shuttleworth (1988) finds that $E_T$ actually increases during the dry season, with plants drawing on large soil moisture stores that swell during the rainy season and making use of the high incoming solar radiation associated with low dry-season cloud cover. The biases described here are accentuated in the coupled model because of a deficit in precipitation compared to observations.

After a series of tests, the following modifications were introduced that together combine to moisten deep CLM3 soils and increase interseasonal soil moisture storage, while maintaining a reasonable runoff simulation:

- CLM3 subsurface runoff is the sum of lateral drainage from soil layers 6–9 and drainage out through the bottom of the soil column. Lateral drainage is very efficient in CLM3, leading to a dry and unvarying bottom soil layer and rapid flushing of soil water out of the column, which suppresses interseasonal soil water storage. This flow out of the grid cell is directed into runoff rather than to an adjoining cell so the soil water is effectively lost from the soil. Here, lateral drainage is turned off, which permits greater soil moisture filtration into the bottom layer and enhances interseasonal soil moisture storage.
- VEG_{off} has more total runoff compared to CONTROL_{off} (see Table 1 and Fig. 3), partly because substantially more rain and dripfall reaches the soil surface due to

![Diagram of Amazonia (10S-0,70-50W)](image)

Fig. 3. Mean annual cycle of components of the hydrologic cycle for CONTROL_{off}, VEG_{off}, and VEGHYD_{off} experiments. $Q_{SURF}$ is surface runoff, $Q_{DRAIN}$ is drainage. GPP is gross primary production.
the reduction in canopy interception. Consequently, significantly more incident water is directed into surface runoff in VEG$_{off}$ (21% globally, 22% Amazon) compared to CONTROL$_{off}$ (16.5% globally, 14% Amazon). The increase in surface runoff, particularly in the Amazon region is undesirable as observations suggest that there is very little surface runoff even after heavy precipitation (Hodnett et al. 1996). To encourage infiltration of water into the soil column, we introduce an infiltration enhancement factor, similar to that used in the Met Office Surface Exchange Scheme (Essery et al. 2001) for the unsaturated fraction of the grid box. The enhancement factor is derived from the weighted mean of root density in the top three soil layers and is meant to represent water conduits into the soil provided by roots. Note that an alternative way to increase infiltration is through a two-mode soil pore size distribution that permits the representation of soil macropores (Liu and Dickinson 2003). This formulation also enhances infiltration but requires a new global surface dataset for soil particle size distribution. For the sake of simplicity, the infiltration enhancement factor is used here.

- Saturated hydraulic conductivity ($K_{sat}$) contains a scaling factor in CLM3 that generates an exponential decrease in $K_{sat}$ with soil depth that reflects the decrease in permeability at deeper soil levels (Beven 1984; Elsenbeer et al. 1992). However, this vertical variation in $K_{sat}$ is accounted for explicitly in CLM3 as vertical variations in soil texture incorporated into the soil dataset (increasing clay content with depth). To eliminate this “double counting” of the vertical variation of $K_{sat}$, the exponential scaling factor for $K_{sat}$ is removed.

Offline tests of the modifications to soil hydrology (hereafter VEGHYD$_{off}$ for all vegetation and hydrology modifications) show a significant impact on soil wetness, runoff, and ET partitioning. Globally, soils are considerably wetter, especially at deep soil levels, and the annual cycle of soil moisture is far more pronounced in most areas. Figure 4 shows observed annual cycles of soil moisture for the Anglo-Brazilian Amazonian Climate Observation Study (ABRACOS) site in Rondônia, Brazil (Gash et al. 1996); station data averaged over 19 sites in Illinois (Hollinger and Isard 1994); and station data averaged over 11 sites in central Asia (Vinnikov and Yeserkepova 1991). The latter two datasets were obtained from the Global Soil Moisture Data Bank (Robock et al. 2000). For both ABRACOS and Illinois, the new hydrology parameterizations result in both better mean soil moisture and, more notably, an improved representation of the annual cycle. The impact in the semiarid climate of central Asia is minimal.

Wetter root-zone soil moisture contributes to an increase in global $E_T$ (41% $E_T$, 42% $E_S$, and 17% $E_C$; Table 1). Total ET is only slightly lower in VEGHYD$_{off}$ (1.27 mm day$^{-1}$) compared to CONTROL$_{off}$ (1.31 mm day$^{-1}$). There is a corresponding increase in total runoff; an increase that is in the right direction as CONTROL$_{off}$ has a small negative bias in mean annual total runoff compared to the Fekete et al. (2002) runoff climatology. The global runoff ratios (runoff divided by precipitation) for Fekete, CONTROL$_{off}$, and
VEGHYD_{off} are 0.40, 0.36, and 0.38, respectively. This bias is also apparent in long-term mean global river discharge. Dai and Trenberth (2002) estimate river discharge to be $1.20 \times 10^6$ m$^3$ s$^{-1}$, while CONTROL_{off} and VEGHYD_{off} produce discharges of 1.08 and 1.15 × $10^6$ m$^3$ s$^{-1}$.

The impact of the hydrology modifications varies regionally but can be summarized by their impact in Amazonia (see Fig. 3). The clearest impact is that the deep soil levels now accumulate moisture throughout the wet season, which permits plants to transpire throughout the dry season, drawing on the stores of moisture accumulated during the wet season. Because of the increase in dry-season $E_T$, total ET is much less variable over the course of the year, as desired. Note also that while surface runoff is higher in VEG_{off}, the infiltration enhancement factor brings it back down in VEGHYD_{off}, thus permitting more incident water to enter the soil.

The global ET partitioning in VEGHYD_{off} is shown in Fig. 5. The partitioning (41% $E_T$, 42% $E_S$, and 17% $E_C$) is in much better agreement with prior estimates, although $E_T$ remains somewhat low and $E_S$ a little high compared to other estimates. An improved CLM3 surface dataset based on Moderate Resolution Imaging Spectroradiometer (MODIS) data became available during the latter stages of this study, which significantly alters prescribed LAI (P. Lawrence 2005, personal communication). In the new dataset many regions, such as the tropical rain forests and South Asia, are characterized by notably higher LAI, while in predominantly agricultural regions, such as the eastern United States and Europe, the shift is toward smaller LAI. When VEGHYD_{off} is run with the new CLM3 surface dataset, ET partitioning is further improved (44% $E_T$, 39% $E_S$, and 17% $E_C$).

Overall, ET partitioning in VEGHYD_{off} is more in line with what one would intuitively expect with $E_T$ dominant in tropical rain forest and heavy agricultural regions, while $E_S$ is a more significant component of ET in arid and semiarid regions. The canopy evaporation $E_C$ is much reduced globally but remains a significant component of total ET in areas with dense canopies, such as the Amazon and central Africa. The large shift in ET partitioning is particularly apparent in Amazonia where partitioning shifts from 11% $E_T$, 18% $E_S$, and 71% $E_C$ in CONTROL_{off} to 51% $E_T$, 19% $E_S$, and 30% $E_C$ (58% $E_T$, 9% $E_S$, and 33% $E_C$ with the new LAI dataset) in VEGHYD_{off}.

4. Impact of ET partitioning on climate simulations

a. Mean climate

In this section, the impact of the imposed changes to CLM3 on the CAM3–CLM3 climate is evaluated.
Photosynthesis, contributes to a reduction of a volume. Annual mean properties of CAM3–CLM3 CONTROL and VEGHYD simulations averaged over all land points.

<table>
<thead>
<tr>
<th>Variable</th>
<th>CAM3–CLM3 CONTROL</th>
<th>CAM3–CLM3 VEGHYD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation (mm day⁻¹)</td>
<td>2.31</td>
<td>2.37</td>
</tr>
<tr>
<td>2-m temperature (K)</td>
<td>282.6</td>
<td>282.3</td>
</tr>
<tr>
<td>Latent heat (W m⁻²)</td>
<td>42.0</td>
<td>45.5</td>
</tr>
<tr>
<td>Sensible heat (W m⁻²)</td>
<td>29.8</td>
<td>26.0</td>
</tr>
<tr>
<td>Transpiration (% mm day⁻¹)</td>
<td>11.0, 0.16</td>
<td>33 (42), 0.52</td>
</tr>
<tr>
<td>Soil evaporation (%)</td>
<td>57.0, 0.82</td>
<td>54 (41), 0.84</td>
</tr>
<tr>
<td>Canopy evaporation (%)</td>
<td>32.4, 0.46</td>
<td>13 (17), 0.21</td>
</tr>
<tr>
<td>Total evapotranspiration (mm day⁻¹)</td>
<td>1.44</td>
<td>1.57</td>
</tr>
<tr>
<td>Surface runoff (mm day⁻¹)</td>
<td>0.432</td>
<td>0.340</td>
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<tr>
<td>Subsurface runoff (mm day⁻¹)</td>
<td>0.365</td>
<td>0.396</td>
</tr>
<tr>
<td>Incoming solar (W m⁻²)</td>
<td>179.3</td>
<td>175.9</td>
</tr>
<tr>
<td>Absorbed solar (W m⁻²)</td>
<td>135.0</td>
<td>132.6</td>
</tr>
<tr>
<td>Incoming longwave (W m⁻²)</td>
<td>311.5</td>
<td>312.3</td>
</tr>
<tr>
<td>Emitted longwave (W m⁻²)</td>
<td>374.2</td>
<td>372.7</td>
</tr>
<tr>
<td>Net radiation (W m⁻²)</td>
<td>72.4</td>
<td>72.1</td>
</tr>
<tr>
<td>Photosynthesis (GPP, Pg C yr⁻¹)</td>
<td>65</td>
<td>148</td>
</tr>
<tr>
<td>Soil moisture stress (Bₛ)</td>
<td>0.35</td>
<td>0.51</td>
</tr>
<tr>
<td>Soil moisture (mm)</td>
<td>301</td>
<td>342</td>
</tr>
</tbody>
</table>

The CONTROL and VEGHYD versions of the model are run for 22 yr (1979–2000), forced by observed sea surface temperatures and initialized with a spunup land surface state derived from the respective offline experiments. Only the last 17 yr are analyzed to allow the model to adjust to the initial conditions. Table 2 lists annual mean values, averaged over all land points, for the CONTROL and VEGHYD experiments. ET partitioning is improved from 11% $E_T$, 57% $E_S$, and 32% $E_C$ in CONTROL to 33% $E_T$, 54% $E_S$, and 13% $E_C$ in VEGHYD. The improvement is less than that seen in the offline experiments. To a certain extent, the difference in ET partitioning between the CAM3–CLM3 coupled and CLM3 offline experiments is related to biases in CAM3–CLM3 precipitation, biases that are essentially unaffected by the CLM3 VEGHYD changes. These include a deficit of precipitation in some heavily forested areas, where $E_T$ should dominate but does not, such as the southeastern United States and Amazonia, and significant overestimates of precipitation in semiarid and arid regions, where $E_S$ contributes significantly, particularly in Saudi Arabia and India (Hack et al. 2006). If one considers only locations where CAM3 annual mean precipitation is within 20% of the Global Precipitation Climatology Project (GPCP; Adler et al. 2003) observed precipitation estimates, the ET partitioning is similar to that of the offline model (42% $E_T$, 41% $E_S$, and 17% $E_C$).

There are some additional broad changes to climate seen in VEGHYD, apart from ET partitioning. For example, the global hydrologic cycle is slightly more vigorous with modest increases seen in both global precipitation (2.31 to 2.37 mm day⁻¹; GPCP estimate, 2.15 mm day⁻¹) and ET (1.44 to 1.57 mm day⁻¹) while runoff decreases (0.86 to 0.80 mm day⁻¹). The increase in precipitation may reflect the global mean 3 W m⁻² redistribution of surface available energy into latent heat flux at the expense of sensible heat flux, a redistribution that is likely driven by the overall wetter soils. On average, global soil water storage rises by 13%, an increase that corresponds to a more substantial 46% increase in global mean $\beta$, (0.35 to 0.51). Photosynthesis, or gross primary production (GPP), also increases substantially from 64 to 148 Pg C yr⁻¹ (Schlesinger 1997), especially when one considers that nitrogen limitation on photosynthesis is not yet incorporated into CLM.

The changes to model climate exhibit significant regional differences. Seasonal mean maps of VEGHYD minus CONTROL differences in precipitation, temperature, ET, runoff, $\beta$, and photosynthesis are shown for June–August (JJA; Fig. 6) and December–February (DJF; Fig. 7). Of initial concern is whether or not the modifications improve or degrade the overall climate simulation. The impacts on precipitation cannot be classified clearly as an improvement or degradation of the model precipitation climatology, although the increase in summer precipitation in the southeastern United States is an improvement. The impacts on temperature, however, broadly drive the model toward improved agreement with the observations (Willmott and Matsuura 2000). In particular, higher ET in the Northern Hemisphere midlatitudes (30°–50°N) and Amazonia contributes to a reduction in JJA warm biases. Similarly, higher dry-season $E_T$ contributes to a reduction of a significant DJF warm bias in India.

The reduction seen in JJA runoff over southern Asia and Alaska (see Fig. 6) is related to an approximately 1-month delay in the timing of peak runoff as soil water is forced in VEGHYD to percolate through the entire column before exiting the bottom of the soil column. In India, for example, runoff peaks during JJA in CONTROL but peaks during July–September in VEGHYD, leading to the reduced JJA runoff seen in Fig. 6. This effect occurs in Southeast Asia as well, but with additional reduction in runoff related to enhanced $E_T$ during DJF that is not balanced by increased precipitation.
b. Soil water storage and soil moisture memory

Overall, soils are wetter throughout the year in most locations, except in very high latitude and arid regions. The factor $\beta$, is correspondingly higher across the same regions, as is photosynthesis. Wetter soils correspond to greater soil water storage, which in turn permits plants to continue to transpire during dry periods. This transformation in model behavior is emphasized in Fig. 8, which shows, for CONTROL and VEGHYD, the average number of months during the year when evapotranspiration exceeds precipitation ($P - ET < 0$). In CONTROL, ET rarely exceeds precipitation, suggesting that typically precipitation is rapidly recycled as ET or lost from the soil column through runoff. In contrast, VEGHYD exhibits numerous grid points where $P - ET < 0$ for more than 3 month yr$^{-1}$.

Does this change in $P - ET$ behavior indicate longer soil moisture memory time scales in VEGHYD? Not necessarily: it may simply reflect higher mean soil moisture values. It is interesting, nonetheless to consider how the hydrology modifications affect time scales of soil moisture memory. Figure 9 shows maps and vertical profiles averaged over a number of latitude bands of the lag-one autocorrelation coefficient, a single parameter measure of soil moisture memory (Wu and Dickinson 2004). For each grid point and soil layer, the anomaly time series of volumetric soil water is correlated with itself but lagged by one month. The anomaly time series is generated by subtracting the climatologi-
cal mean annual cycle from the monthly mean time series. Higher autocorrelation values indicate longer anomaly decay time scales; autocorrelations of 0.8, 0.6, 0.4, and 0.2 correspond to e-folding anomaly decay time scales of 4.5, 2.0, 1.1, and 0.6 months, respectively (Wu and Dickinson 2004). For the most part, differences in soil moisture memory between CONTROL and VEGHYD can be attributed to the impacts of the model modifications to ET, the partitioning of incident precipitation into infiltration and runoff, and the rate at which water percolates through (and out of) the soil column. However, soil moisture memory, as discussed in Koster and Suarez (2001), also depends on the seasonality in the statistics of atmospheric forcing and the strength of the land–atmosphere feedback. While the contribution of changes in these influences across the two models cannot fully be discounted, it is reasonable to expect that the direct changes to model parameterizations are more likely responsible for the soil moisture memory differences seen than the more subtle changes to model climate.

CONTROL and VEGHYD exhibit significant differences in the vertical profile of the lag-one autocorrelation with depth. In general, soil moisture autocorrelations increase with soil depth. In the uppermost part of the soil column, the time scale of soil moisture persistence is low due to the influence of strong precipitation and evaporation forcing from above, whereas deeper in the soil column, a soil moisture anomaly can persist for many months and is set more strongly by the rate at which water permeates through the soil column. Below about 0.5-m depth soil moisture persistence is considerably shortened in VEGHYD. This reduction in soil moisture memory deep in the soil is related to the en-
hanced soil moisture annual cycle in VEGHYD. In
CONTROL, the efficient export of soil water out of the
soil column as midlevel drainage prevents the deepest
soil layers from experiencing much soil moisture varia-
tion on annual time scales. The relative constancy cor-
responds to high lag-one autocorrelation values and ex-
tended soil moisture memory time scales. Even though
the diagnosed time scales are lengthy in CONTROL, the actual soil moisture anomalies are very small and
are not felt by the atmosphere despite persisting for
long periods. In contrast, interannual variability at
depth is considerably higher in VEGHYD than in
CONTROL (not shown). Therefore, even though the soil moisture anomaly persistence time scales are
shorter, due to more rapid loss of water through tran-
spiration and faster percolation of water through the
soil column, an anomaly that develops is far more likely
in VEGHYD to exert a detectable influence on subse-
quent seasons’ climate.

c. Response to rain event

The shift in how CLM3 partitions ET is likely to
affect the turbulent flux and temperature responses to a
precipitation event, which can, in turn, influence land–
atmosphere feedbacks. A priori, one would expect that
in CONTROL the high interception and soil evapora-
tion would lead to a sharp increase in ET directly fol-
lowing a rain event, with ET dropping off quickly after
the canopy water store is exhausted. On the other hand,
since $E_T$ is more important in VEGHYD, one would
anticipate the ET response to a rain event in VEGHYD
to be slightly delayed relative to the precipitation event
and to extend for a longer time period. This behavior is
precisely what is seen in composites of the 5-day evolu-
tion of ET and temperature subsequent to a heavy
rain event. We calculate lagged composites of ET and
temperature relative to rain events that meet the fol-
lowing conditions: (i) rain exceeds 5 mm day$^{-1}$, and (ii)
event is followed by five consecutive days with rainfall
rates less than 0.5 mm day$^{-1}$. A total of 16-yr of daily
JJA data from CONTROL and VEGHYD are consid-
ered for the composites. Results are shown for two re-
gions, both of which are characterized by strong shifts
in ET partitioning: the eastern United States (Fig. 10)
and Amazonia (Fig. 11). By coincidence rather than
design, the day $−2$ to day 0 precipitation composites are

Fig. 8. Global maps showing number of months during the year when evapotranspiration exceeds precipitation
($P − ET < 0$). Grid points with lake or wetland fractional coverage exceeding 50% are not considered.
nearly identical in CONTROL and VEGHYD. Considering first the eastern U.S. composites, ET peaks on day 0 in both CONTROL and VEGHYD, but the peak is lower in VEGHYD due to the smaller $E_C$ contribution on the same day as the rainfall event. Both $E_T$ and $E_S$ peak on day +1 for both models, but the $E_T$ peak is stronger and the $E_S$ peak is weaker in VEGHYD. ET follows a similar downward trajectory from day +1 to day +5 in CONTROL and VEGHYD, but ET is about 1 mm day$^{-1}$ higher, due mainly to higher $E_T$, in VEGHYD. The temperature response is consistent with the ET response, with temperatures hitting a minimum on the day of the rainfall event in CONTROL, due to enhanced ET and reduced incident solar radiation due to cloud cover, and recovering quickly in the subsequent days. Persistently high ET after the rain event leads to a delayed temperature recovery in VEGHYD. The composite picture is similar for Amazonia, with the main difference being a more subdued ET response to a rain event in VEGHYD due to the relatively weak mean moisture limitation on ET.

The number of events with rain followed by five consecutive dry days is sharply lower for both focus regions (eastern United States: 369 in CONTROL and 157 in VEGHYD; Amazonia: 412 in CONTROL and 114 in VEGHYD). Mean precipitation, however, is essentially

![Fig. 9. The average of soil moisture 1-month-lag autocorrelation for JJA for (upper maps) top 0.1 m of soil and (lower maps) top 2.0 m of soil for CONTROL and VEGHYD. Latitude belt mean lag-one autocorrelation coefficients vs soil depth for JJA and DJF in CONTROL and VEGHYD. Latitude belts are Tropics ($14^\circ S$–$14^\circ N$), northern subtropics ($14^\circ$–$33^\circ N$), northern midlatitudes ($33^\circ$–$56^\circ N$), and northern high latitudes ($56^\circ$–$73^\circ N$).](http://journals.ametsoc.org/jhm/article-pdf/8/4/862/4160786/jhm596_1.pdf)
unchanged between CONTROL and VEGHYD for both regions. This implies a shift in the precipitation distribution. Figures 12a,b show the JJA precipitation distribution for CONTROL and VEGHYD for the eastern U.S. and Amazonia regions. In both regions, the shift is toward more frequent light rainfall at the expense of heavy rainfall. Intuitively, the shift makes sense, especially when it is considered in the context of the known strong sensitivity of precipitation to surface turbulent heat flux forcing in CAM3–CLM3 (Guo et al. 2006). Due to the steadying influence of transpiration, which responds slowly to longer-term weather anomalies, the day-to-day variability of latent heat flux is lower in VEGHYD compared to CONTROL, in both regions. In the absence of strong forcing associated with large latent heat flux excursions from the norm, precipitation rates remain fairly steady in VEGHYD. A scatter diagram of the change in latent heat flux variability versus the change in the median wet-day rain rate is shown in Fig. 12d [a change in the median wet-day rain rate is a simple way to detect a shift in precipitation distribution; Osborn et al. (2000)]. In general, an increase in latent heat flux variability corresponds to a shift toward more heavy rain events and vice versa, although the correlation coefficient is modest ($r = 0.43$), indicating that other factors are likely involved. Changes in mean latent heat flux do not, however, correlate with shifts in the precipitation distribution ($r = -0.09$, Fig. 12c).

**d. Global Land–Atmosphere Coupling Experiment**

The GLACE model intercomparison (Guo et al. 2006; Koster et al. 2006) demonstrated that the strength of interaction between the land surface and the atmosphere varies widely across current GCMs. CAM3–CLM3 is one of the strongest models in terms of the diagnosed influence of subsurface soil moisture on precipitation even though $E_T$, the portion of ET that is directly influenced by subsurface soil moisture, is only a small fraction of total ET in CONTROL, the model version used for the GLACE study. It is interesting, therefore, to recalculate land–atmosphere coupling strength for VEGHYD to see whether and how cou-
Fig. 11. Same as in Fig. 10, but for Amazonia (10°S-0°, 50°-70°W).

Fig. 12. JJA precipitation distribution, calculated based on daily precipitation from 16 summers, for CONTROL (solid line) and VEGHYD (dashed) for (a) the eastern United States and (b) Amazonia. Scatterplot showing change (VEGHYD − CONTROL) in (c) mean and (d) standard deviation of latent heat flux vs change in median wet-day rain rate. Only nonice, nondesert land grid points are plotted. Correlation coefficients (r) are also shown.
pling strength is affected in a version of the model with more reasonable ET partitioning.

Land–atmosphere coupling strength, or the influence of subsurface soil moisture (SM$_{sub}$) on precipitation ($P$), is determined in GLACE through an experiment consisting of two ensembles of simulations; the first set is run with a freely evolving land surface (the W ensemble in GLACE) and the second set is forced with subsurface soil moisture taken from one of the W ensemble members (denoted the S ensemble in GLACE). Each ensemble consists of 16 simulations of the same boreal summer (JJA, 1994), with each ensemble member forced with different initial conditions. The similarity of time series across the S or the W ensemble ($\Omega[S]$, $\Omega[W]$, respectively) varies between 0 (no correlation) and 1 (perfect correlation). We refer the reader to Koster et al. (2006) for more details on the experimental design and calculation of the $\Omega$ diagnostic. A land–atmosphere feedback is implied when there is a greater similarity across ensemble member time series in the forced soil moisture ensemble (S) relative to the freely evolving ensemble (W), for example, when $\Delta\Omega = \Omega[S] - \Omega[W] > 0$.

The global distribution of land–atmosphere coupling strength for precipitation ($\Delta\Omega_p$) and evapotranspiration ($\Delta\Omega_E$) for CONTROL and VEGHYD is shown in Fig. 13. It is interesting that even though $E_T$ makes up a substantially larger contribution to total ET in VEGHYD, the SM$_{sub}$–ET relationship is not substantially altered. While there are some regional differences in $\Delta\Omega_E$, particularly in Amazonia, where soil moisture stress on $E_T$ is reduced in VEGHYD leading to a weaker SM$_{sub}$–ET relationship there, overall the impact of the changes to ET partitioning on the SM$_{sub}$–ET relationship is modest. Globally averaged values rise only slightly ($\Delta\Omega_E = 0.139$ in CONTROL; $\Delta\Omega_E = 0.147$ in VEGHYD). The SM$_{sub}$–$P$ relationship also strengthens slightly ($\Delta\Omega_p = 0.035$ in CONTROL; $\Delta\Omega_p = 0.045$ in VEGHYD).

Global averages of $\Omega[S] - \Omega[W]$, $\Omega[S]$, and $\Omega[W]$ over nonice land points for precipitation, ET, 2-m air temperature, $E_T$, $E_S$, and $E_C$ are listed in Table 3. The relatively weak impact of the differences in ET partitioning on the SM$_{sub}$–ET relationship may at least partially be explained by the relatively large influence on $\Omega[W]$: high $\Omega[W]$ values are seen when there is a large

![Fig. 13. Global distribution of land–atmosphere coupling strength ($\Omega[S] - \Omega[W]$) for (top) precipitation and (bottom) evapotranspiration for CONTROL and VEGHYD experiments. Global average values of $\Delta\Omega = \Omega[S] - \Omega[W]$ over all nonice land points are shown at bottom of each plot.](http://journals.ametsoc.org/jhm/article-pdf/8/4/862/4160786/jhm596_1.pdf)
degree of similarity across the unforced W ensemble time series due to the annual cycle (Koster et al. 2006). For example, $\Omega_{\text{ET}}[W]$ is 0.302 in CONTROL and is 0.366 in VEGHYD, reflecting the larger amplitude soil moisture annual cycle and its greater control on $E_T$. The 0.064 increase in $\Omega_{\text{ET}}[W]$ accounts for most of the 0.074 increase in $\Omega_{\text{ET}}[S]$ and hence the change in $\Delta \Omega_{\text{ET}}$ is only 0.010. This suggests that even though VEGHYD has higher mean $E_T$, the influence of SM$_{\text{soil}}$ anomalies on $E_T$ is roughly equivalent across the two versions of the model. It is interesting that $\Delta \Omega_{\text{ET}}$ is roughly the same as $\Delta \Omega_{\text{ET}}$ even though soil moisture is specified in the S ensemble only at soil levels deeper than 5 cm. A high $\Delta \Omega_{\text{ES}}$ implies that even though the soil moisture in the top soil layers is permitted to evolve freely in the S ensemble, it is modulated by what is happening at deeper levels, presumably through suction of soil water from below when the upper soil is dry and the deep soil is wet and more efficient transfer of water downward when the upper soil is wet and the deep soil is dry.

5. Summary

The partitioning of ET in the control version of CLM3 is poor with global canopy evaporation and soil evaporation far outweighing transpiration, a partitioning that is inconsistent with other offline model estimates. A series of modifications to CLM3 vegetation and soil hydrology parameterizations significantly improve the partitioning of ET, with notable increases seen in transpiration and photosynthesis. The global partitioning is improved in the offline model from 13%, 0.17 mm day$^{-1}$ $E_T$: 44%, 0.57 mm day$^{-1}$ $E_C$ and 43%, 0.56 mm day$^{-1}$ $E_C$ to 41%, 0.52 mm day$^{-1}$ $E_T$: 42%, 0.53 mm day$^{-1}$ $E_S$: and 17%, 0.21 mm day$^{-1}$ $E_C$. The broader impact on climate simulations is on the balance beneficial, particularly in the Amazon basin where large biases in dry-season evapotranspiration and air temperatures are sharply reduced. Although the impacts of the new model on mean climate are relatively small, the influence on the nature of the climate simulation is diverse. Stronger transpiration and reduced canopy evaporation result in a more muted and extended ET response to a rain event. In addition, the influence of subsurface soil moisture on precipitation is slightly higher due to increased influence of subsurface soil moisture on transpiration, although the results shown here do not alter the broad conclusions reached in Guo et al. (2006) that convection in CAM3 is overly sensitive to surface heat flux forcing. These analyses confirm the importance of ET partitioning on climate simulations and land–atmosphere interaction in climate models.

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