Weak Land–Atmosphere Coupling Strength in HadAM3: The Role of Soil Moisture Variability

DAVID M. LAWRENCE AND JULIA M. SLINGO

NCAS Centre for Global Atmospheric Modelling, University of Reading, Reading, United Kingdom

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ABSTRACT

A recent model intercomparison, the Global Land–Atmosphere Coupling Experiment (GLACE), showed that there is a wide range of land–atmosphere coupling strengths, or the degree that soil moisture affects the generation of precipitation, amongst current atmospheric general circulation models (AGCMs). Coupling strength in the Hadley Centre atmosphere model (HadAM3) is among the weakest of all AGCMs considered in GLACE. Reasons for the weak HadAM3 coupling strength are sought here. In particular, the impact of pervasive saturated soil conditions and low soil moisture variability on coupling strength is assessed. It is found that when the soil model is modified to reduce the occurrence of soil moisture saturation and to encourage soil moisture variability, the soil moisture–precipitation feedback remains weak, even though the relationship between soil moisture and evaporation is strengthened.

Composites of the diurnal cycle, constructed relative to soil moisture, indicate that the model can simulate key differences in boundary layer development over wet versus dry soils. In particular, the influence of wet or dry soil on the diurnal cycles of Bowen ratio, boundary layer height, and total heat flux are largely consistent with the observed influence of soil moisture on these properties. However, despite what appears to be successful simulation of these key aspects of the indirect soil moisture–precipitation feedback, the model does not capture observed differences for wet and dry soils in the daily accumulation of boundary layer moist static energy, a crucial feature of the feedback mechanism.

1. Introduction

In a recent model intercomparison study involving four atmospheric general circulation models (AGCMs), Koster et al. (2002) objectively demonstrate that the degree of land–atmosphere interaction, or the extent that land surface conditions, particularly soil moisture, influence the generation and strength of convective precipitation, differs significantly amongst the four AGCMs considered. The results of the pilot study were confirmed in a more extensive model intercomparison project, the Global Land–Atmosphere Coupling Experiment (GLACE), involving 12 AGCMs (Koster et al. 2004). The reasons behind the notable range in land–atmosphere coupling strengths are not yet fully understood. Progress toward a better understanding is complicated by the complexity of the theoretical coupling mechanism (described in detail below), which prevents a simple classification of coupling strength relative to model formulation.

While Koster et al. stress that real-world coupling strength is not known, there is some evidence, mostly indirect, to suggest that the soil moisture–precipitation feedback is an important part of the climate system, at least regionally (Findell and Eltahir 1997; Taylor and Lebel 1998; Koster et al. 2003). The feedback mechanism can operate on a wide range of time scales from hours to years. For example, Taylor and Lebel (1998) have demonstrated that the transient moistening of the surface soil layers by individual storms over West Africa can predispose the system to further convection in the subsequent hours. On the longer time scales, soil moisture feedbacks have been implicated in seasonal droughts over the United States (e.g., Beljaars et al. 1996; Schubert et al. 2004) and in the decadal drying of the Sahel (e.g., Walker and Rowntree 1977). Koster et al. focused on the feedbacks operating on the shorter time scales of hours to days and that will also be the case for this study.

The significant differences in AGCM land–atmosphere coupling strength are not entirely surprising.
given the complexity of the soil moisture–precipitation feedback processes. A sequence of papers published in the late 1990s (Betts and Ball 1995; Betts et al. 1996; Eltahir 1998; Schär et al. 1999) outline a mechanism for what is known as the indirect soil moisture–precipitation feedback. The feedback is termed “indirect” because the influence of anomalously wet soil on precipitation is not simply via a recycling of the extra water vapor associated with enhanced surface evaporation into rainfall (the “direct” feedback). Instead, the indirect feedback, which some studies suggest is the dominant feedback (e.g., Schär et al. 1999), considers how soil moisture affects precipitation through its influence on boundary layer characteristics and atmospheric stability. The indirect feedback mechanism, illustrated in schematic form in Fig. 1, works in the following way.

Anomalously wet soil generated by a significant rainfall event induces a strong evapotranspiration response through enhanced plant transpiration and soil evaporation. Consequently, more of the available surface energy is devoted to latent heat (LH) rather than sensible heat (SH) flux, and the Bowen ratio (SH/LH) is correspondingly low, resulting in a relatively shallow boundary layer. The enhanced moisture flux into a shallow boundary layer leads to an anomalously moist and therefore less stable boundary layer. In addition, wet soils tend to be dark and relatively cool, which together enhances net surface radiation (greater absorption of solar radiation coupled with reduced longwave emission), resulting in a greater total heat flux into the boundary layer. The above factors combine to generate a larger moist static energy (MSE) per unit mass of boundary layer air over wet soils compared to over dry soils, and therefore may lead to enhanced precipitation under conditions when deep moist convection can be triggered by boundary layer instability. Note that the feedback mechanism described here neglects the influence of cloud cover, which complicates the feedback because of its influence on surface net radiation.

The Hadley Centre AGCM (HadAM3), which incorporates the Met Office Surface Exchange Scheme (MOSES2) land surface scheme, exhibits an extremely weak soil moisture–precipitation feedback such that when averaged globally, HadAM3 ranks at or near the bottom of the 12 AGCMs considered in Koster et al. (2004). In two recent studies, HadAM3–MOSES2 is
used to evaluate the influence of vegetation on climate (Lawrence and Slingo 2004b; Osborne et al. 2004). Those studies conclude that although vegetation characteristics have a demonstrable effect on surface evaporation and other aspects of the surface climate, precipitation is largely insensitive to vegetation-induced variations in surface evaporation, at least in HadAM3. Considered in the context of the GLACE result, the conclusion that vegetation does not affect precipitation may actually be a consequence of HadAM3’s weak coupling, and any more general conclusions about the sensitivity of precipitation to vegetation cover cannot reasonably be made. This uncertainty highlights the need to understand the nature of the wide range of AGCM coupling strengths and invokes an unresolved overarching question for the climate change research community. That is, how sensitive are conclusions about the influence of, for example, land cover change on climate to the inherent land–atmosphere coupling strength of a particular model?

The preceding question is clearly difficult to answer. A first step is to identify what aspects of the climate modeling system regulate land–atmosphere coupling strength. It would appear that to reproduce the indirect soil moisture–precipitation feedback in a climate model requires realistic simulation of, at the least, soil moisture and its variability, partitioning of surface energy into latent and sensible heat fluxes, the boundary layer response to surface fluxes, and the relationship between MSE and convective initiation. In this paper, we begin to investigate the weak HadAM3 soil moisture–precipitation feedback by evaluating whether or not enhancing the soil moisture–evaporation relationship through changes to the soil model affects the diagnosed land–atmosphere coupling strength.

2. Description of model

The AGCM assessed in this study is HadAM3 (Pope et al. 2000). The model is run at standard climate resolution, a 2.5° latitude by 3.75° longitude Arakawa B grid, with 19 vertical levels in the atmosphere. HadAM3 has performed well in the Atmospheric Model Intercomparison Project (AMIP), producing realistic contemporary patterns of temperature and precipitation.

The land surface scheme utilized by HadAM3 is MOSES2 (Cox et al. 1999; Essery et al. 2003). Each land grid box is represented by a mixture of five vegetation or plant-functional types (PFTs; broadleaf trees, needleleaf trees, temperate C₃ grass, tropical C₄ grass, and shrubs) and four nonvegetated surface types (urban, inland water, bare soil, and ice). Surface fluxes and temperatures are calculated separately for each surface type and are aggregated according to each tiles’ fractional coverage before being passed to the atmospheric model. The standard configuration contains four soil levels of depths 0.1, 0.25, 0.65, and 2.0 m. Drainage through the soil is calculated via the Richards’ equation, and hydraulic conductivity and suction are calculated using Clapp–Hornberger characteristic curves (Clapp and Hornberger 1978). Precipitation is assumed to fall on a fraction of each grid box (0.1 for convective rain and 0.5 for large-scale rain). A portion of the rainfall is intercepted by the canopy with the remainder falling to the surface as throughfall. The throughfall infiltrates the soil at a rate equal to the saturated hydraulic conductivity multiplied by an infiltration enhancement factor, which is dependent on the presence and type of vegetation. Variations to the documented version of MOSES2 include the addition of an imposed vegetation annual cycle based on satellite information and the introduction of a modest soil moisture dependence on soil albedo (Lawrence and Slingo 2004a). The standard MOSES soil parameter dataset described in Cox et al. (1999) is used here although it is worth noting that Osborne et al. (2004) found that the model’s climate, and its response to tropical deforestation, is sensitive to the soil parameter dataset used.

3. Land–atmosphere coupling strength in HadAM3

A method for objective assessment of the degree of interaction between the land surface and the atmosphere in a climate model is described in Koster et al. (2002). The experiment consists of two sets of 16-member 92-day simulations spanning June–August, forced only by a particular year’s SSTs, 1994 in this case. Each ensemble member is initialized with 1 June conditions from each year of a 16-yr control run. The first ensemble, denoted W for write, is run with a fully interactive land surface. For one W ensemble member, for example, W1, soil moisture values for each of the four soil levels are recorded at each time step and saved to a file. In the second set of simulations, denoted S for soil moisture, the simulated soil moisture is replaced at each time step with values from the W1 simulation. Thus, the atmosphere in all S member simulations sees the same time-varying, spatially varying anomalies of soil moisture. The degree of similarity amongst the members of a simulation is measured as Ω, a parameter derived from 5-day mean time series that varies from 0 (implying a complete lack of coherence) to 1 (implying maximal coherence, such that all simulations produce precisely the same time series). The use of 5 days for
the time averaging is arbitrary; other averaging periods produce qualitatively similar results. The $\Omega$ parameter is defined in Koster et al. (2002) as

$$\Omega = \frac{16\sigma^2 - \bar{\sigma}^2}{15\sigma^2},$$  

where $\sigma$ is the standard deviation computed across the 288 five-day totals included in the 16 simulations for a given ensemble, and $\bar{\sigma}$ is the standard deviation of the ensemble mean time series computed from the 18 five-day totals in the 90-day ensemble mean time series. By comparing $\Omega[S]$, calculated from the W ensemble, to $\Omega[W]$, calculated from the S ensemble, the influence of soil moisture on the atmospheric state can be quantified. The standard setup for the GLACE experiments calls for soil moisture to be updated only in the subsurface root zone, which is defined to be all soil levels below 5-cm depth. Since the uppermost MOSES2 soil layer depth is deeper than 5 cm, soil moisture is prescribed for all four model soil layers for the S ensemble.

Global maps of the $\Omega$ coupling strength diagnostic for precipitation, 2-m air temperature, evaporation, and Bowen ratio are shown in Fig. 2. As noted in the introduction, HadAM3 exhibits very weak coupling strength for precipitation ($\Omega[S] - \Omega[W] \approx 0$), which effectively says that soil moisture conditions, either local or remote, do not have a detectable influence on rainfall, at least on subweekly time scales. This result does not necessarily mean, however, that soil moisture conditions have no influence on the atmosphere. Evaporation and 2-m air temperature are both considerably impacted by soil moisture, at least regionally. The soil moisture–evaporation relationship is strongest in arid and semiarid regions where evaporation is typically moisture limited rather than energy limited. In regions where evaporation is typically energy limited, such as in tropical rainforest and high-latitude regions, the soil moisture–evaporation relationship is weak. Areas of high $\Omega_E$ are coincident with areas of high $\Omega_{BR}$(Bowen ratio), indicating, not surprisingly, that where soil moisture controls evaporation it also regulates the partitioning of surface available energy into latent and sensible heat fluxes.

At first glance, the fact that $\Omega_E$ is relatively large in some areas might lead one to conclude that the model’s soil moisture–evaporation relationship is not the source of the model’s weak soil moisture–precipitation feedback. However, since the soil moisture–evaporation re-
relationship is crucially required in the indirect soil moisture–precipitation feedback, it is worth considering whether or not this initial conclusion is correct. There are at least two possible ways in which a model’s soil moisture–evaporation relationship, while present, may be inadequate to force a feedback on precipitation. The first possibility is that regions of high $\Omega_E$ do not necessarily have large enough evaporation variability to strongly influence the boundary layer. A second possibility is that some locations may exhibit a high calculated $\Omega_E$ even though evaporation variability is low. In such a situation, the influence of soil moisture on the atmosphere will be minimal. Only when a high $\Omega_E$ value is accompanied by significant evaporation variability can soil moisture anomalies have a strong influence on the atmosphere.

Another possibility is that the regional distribution of high $\Omega_E$ may not correspond to regions where surface forcing is a standard trigger of moist convection. Findell and Eltahir (2003a,b) showed that the atmospheric regime can strongly regulate the local feedback between the land surface and the atmosphere and that certain atmospheric profiles favor a positive feedback between soil moisture and convection whereas in other regions the likelihood of moist convection is determined more by oceanic influences rather than land surface conditions. Findell and Eltahir find that in the United States, moist convection is primarily atmospherically controlled in the west whereas in other regions, such as in the southeastern United States, moist convection is more susceptible to surface forcing. In the southeastern United States region in HadAM3, soil moisture stress on vegetation is negligible and surface soil moisture is relatively constant; therefore, soil moisture simply has little influence on the atmosphere there.

**Modified soil model experiment**

It is possible, therefore, that the soil moisture–precipitation feedback in HadAM3 is thwarted because of either a weaker soil moisture–evaporation relationship, due to low evaporation variability or incorrect regionality, than is required to simulate a detectable influence of soil moisture on the triggering of moist convection. To test this hypothesis, we repeat the land–atmosphere coupling strength experiments, but instead of using the standard soil model, we introduce two alterations designed to increase soil moisture variability and, consequently, evaporation variability globally. The modifications directly affect the simulation of soil moisture availability (SMA) and the surface layer soil moisture fraction (SMF). SMA is a parameter that describes the regulation of transpiration due to soil moisture stress (SMA = 1 means no soil moisture stress, SMA = 0 means soil moisture is below wilting point). In HadAM3, a large percentage of the global nonice land surface exhibits summertime mean SMA values near 0 (21%) or 1 (26%), which results in low evaporation variability in these regions. Additionally, the standard deviation of SMF tends to be fairly low in HadAM3, less than 0.05 over 41% of global nonice land surface.

Specifically, the following changes to the soil model are made:

- Adjust $\Theta_{\text{WILT}}$ (volumetric soil moisture concentration at wilting point) and $\Theta_{\text{CRIT}}$ (volumetric soil moisture concentration at critical point), two soil parameters that are required in SMA calculations, so that the seasonal mean soil moisture content (averaged over the 16 W ensemble members) gives an SMA value of 0.5. SMA for a layer is defined as $0 < \Theta_k/(\Theta_{\text{CRIT}} - \Theta_{\text{WILT}}) < 1$; $\Theta_k$ is the simulated volumetric soil moisture concentration of a particular soil layer. See Fig. 3 for schematic illustration of adjustment. The adjustment is conducted separately at each grid point based on that grid box’s simulated mean soil moisture. At the same time, the SMA curve is steepened by shifting $\Theta_{\text{CRIT}}$ and $\Theta_{\text{WILT}}$ closer together by a factor of one-third. These changes are designed to strengthen the dependence of transpiration on soil moisture and should increase the response of transpiration to large precipitation events. The impact of this change should be seen most distinctly for locations where June–August mean SMA is at or near 1.0 in the standard soil experiments.
Decrease top-layer soil depth from 10 to 4 cm. A smaller water holding capacity of the top soil layer should increase SMF variability and therefore soil evaporation variability. A number of surface soil layer depths were tested. The chosen depth of 4 cm represents the shallowest surface soil layer depth that the model could accommodate without generating numerical instability.

It is important to note that the changes to the soil hydrology model described here are motivated solely by the short-term goal of increasing coupling between the land and the atmosphere and are based on prior experience with the land surface model. The changes do not necessarily signify improvements to the treatment of soils in the model.

Overall, the soil model modifications do not significantly alter the mean climate. Near-surface air temperatures are slightly (~1°C–2°C) warmer (not shown) in the modified soil experiments over areas where SMA is at or near 1 in the standard soil experiments. This change in temperature reflects weaker evaporation, and hence lower latent heat flux, due to the forced soil moisture stress on vegetation. Arid and semiarid regions, where SMA in the standard soil experiment is near 0, exhibit a small cooling of the near-surface air temperature associated with greater than normal transpiration. Seasonal mean precipitation is essentially unchanged in the modified soil experiments with only a few grid boxes showing a statistically significant change.

Figure 4 shows how the modifications to the soil model affect the mean and variability of soil moisture. By design, vegetation in the modified soil model experiment is neither completely stressed nor completely unstressed (seasonal mean SMA is neither 0 or 1). Note that \( \Theta_{\text{CRIT}} \) and \( \Theta_{\text{WILT}} \) were adjusted for every grid box so that the control experiment seasonal mean soil moisture gives an SMA value of 0.5. In the modified soil experiments, SMA is not restricted to stay near 0.5 and in fact drifts away from that value as soil moisture accumulates or evaporates over the course of the season. The difference maps (modified soil minus standard soil) of \( \sigma_{\text{SMF}_1} \), \( \sigma_{\text{SMA}} \), and \( \sigma_{\text{SMF}_1} \) show that in most locations a shallower top soil layer leads to a drier and more variable surface soil layer. The primary exception is in the Sahel, where \( \sigma_{\text{SMF}_1} \) is reduced due to a smaller difference between early and late season surface soil moisture in the modified soil experiment. SMA variability is also enhanced globally since an addition or subtraction of soil moisture is now more likely to result in a change in SMA.

The increases in SMA and SMF variability result in a strengthening of the soil moisture–evaporation relationship. The global average \( \Omega_E \) value for nonice land
points rises 15% from 0.32 in the standard experiments to 0.37 in the modified soil experiments. The percentage of the global nonice land surface where $\Omega_E > 0.4$ expands from 35% to 42%. The regional impacts of the soil model changes are especially apparent in plots of $\Omega_E$ (Fig. 5). Global mean $\Omega_E$ rises by 52% from 0.14 to 0.22. The expansion of a notable soil moisture–evaporation relationship into the southeast United States is clearly apparent as is a strengthening of the relationship through enhanced soil moisture variability in the midwest United States. The only notable area where the soil moisture–evaporation relationship weakens is in the Sahel where the soil model changes actually generate lower soil moisture variability than in the standard runs. Despite an overall strengthened soil moisture–evaporation relationship, the soil moisture–precipitation feedback is essentially unaffected and remains small globally; $\Omega_P$ rises from a global mean value of 0.002 to 0.004. The areas of increased $\Omega_P$ that contribute to the global mean increase in $\Omega_P$ do not correspond to the regions of increased $\Omega_E$, indicating that even the small rise in $\Omega_P$ is more likely a statistical artifact than a real change.

4. Discussion

The modified soil model experiment represents, at least to a certain degree, a model setup wherein soil moisture exerts as much control on evaporation as the simulated climate will permit. The fact that a soil moisture–precipitation feedback is still not supported in the model suggests that, at least to first order, it is not the treatment of soil moisture that is restricting a potential soil moisture–precipitation feedback. Therefore, the cause of the weak soil moisture–precipitation feedback must be related either to how the boundary layer adjusts to the surface forcing or how moist convection responds to different boundary layer conditions. In this section, we review a number of known systematic errors in HadAM3—namely an overly frequent rainfall frequency and poor simulation of the timing of the precipitation diurnal cycle over land—that may be related to these aspects of the model’s indirect soil moisture–precipitation feedback.

First, we consider the overly frequent rainfall frequency bias in HadAM3. Maps of June–August wet-day frequency are shown in Fig. 6 for two daily total rainfall thresholds, 0.5 and 2.0 mm day$^{-1}$ (the 16 W simulations are used to calculate wet-day frequency). Also shown in Fig. 6 is the difference between HadAM3 and observed wet-day frequency. The observed estimate is derived from 5 yr [June–July–August (JJA) 1997–2002] of daily satellite/gauge precipitation estimates provided by the Global Precipitation Climatology Project (GPCP; Huffman et al. 2001). Prior to calculation of the GPCP wet-day frequency, the data are regridded from their native $1^\circ \times 1^\circ$ resolution to the HadAM3 grid. Wet-day frequency maps from the Climate Research Unit (not shown; New et al. 1999) exhibit a high degree of both spatial and quantitative similarity to the $P > 0.5$ mm day$^{-1}$ GPCP map. The rainyday frequency bias in HadAM3 is considerable. The bias is especially large over the tropical rainforests, the south and Southeast Asian monsoon regions, and much of North America, precisely those regions where seasonal mean SMA is at or near 1.

The high frequency of rainfall events leads to relatively constant soil moisture levels, which in turn corresponds to low evaporation variability. Time series of daily mean precipitation, SMF$_1$, SMA, and evaporation from the W1 experiment are shown in Fig. 7 for a grid box in the southeastern United States. The absence

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**Fig. 5.** Maps of $(\Omega_E[S] - \Omega_E[W])/\sigma_E[W]$ for (top) standard soils, (middle) modified soils, and (bottom) modified–standard soils.
of any prolonged dry periods prevents the soil from experiencing significant drying and therefore keeps daily evaporation nearly constant throughout the summer.

The rain-day frequency bias itself may be related to well-documented errors in the timing of convective precipitation in HadAM3 (Yang and Slingo 2001; Slingo et al. 2004). Convective precipitation tends to peak at least a few hours too early over most land areas, with the peak occurring almost uniformly near noon compared to the near 1500-h LT peak seen in observations (Nesbitt and Zipser 2003; for non-MCS rainfall). While the reason for the systematically early peak in convective precipitation is not fully understood and is likely complex, it is interesting to consider how this bias might be related to the weak soil moisture–precipitation feedback.

The coupling between the land and the atmosphere should be considered in the context of the diurnal cycle because the development of the daytime boundary layer is key to generating a differential buildup of boundary layer MSE over wet and dry soils (Betts and Ball 1995, hereafter BB). Using data taken during the First International Satellite Land Surface Climatology Project Field Experiment (FIFE), BB construct composites of the diurnal cycle for wet and dry near-surface (top 10 cm) soil conditions. They find that the diurnal evolution of evaporative fraction \[ \frac{LH}{LH + SH} \], sensible and latent heat fluxes, net radiation (total heat flux), and equivalent potential temperature \[ \frac{E}{H} \] all exhibit a significant dependence on soil moisture (see Figs. 22–25 of BB).

Here, we compare their composite results to similarly constructed composites of the HadAM3 diurnal cycle, constructed using hourly HadAM3 model output from the W1 simulation for the same southern Great Plains grid box used above, a grid box which is roughly coincident with the FIFE location. Here, wet soil days are identified objectively as days in July and August where \[ SMF_1 > 0.5 \] and dry soil days are defined as days where \[ SMF_1 < 0.5 \]. Betts and Ball (1995) only consider days that are identified as mostly cloud free to eliminate the influence of cloud radiative feedbacks on the signal. Similarly, we include only days where morning (0500–1000 LT) cloud fraction is < 0.25. Morning cloud frac-
tion is used because the high frequency of daily rainfall is reflected by an afternoon cloud fraction that is almost always greater than 0.25. Based on these criteria, 12 wet soil days and 9 dry soil days are identified during the 60-day July–August period with mean SMF$^1$ values of 0.51 (23% by volume) and 0.16 (7% by volume), respectively.

For the most part, the HadAM3 diurnal cycle composites (Fig. 8) are qualitatively similar to those of BB, indicating that the model can, at least to a certain degree, reasonably simulate the influence of soil moisture on surface and boundary layer conditions. As in BB on wet soil days compared to dry soil days, the darker and cooler soil appears to enhance net radiation. Evaporative fraction is similarly high over wet soils during the middle of the day and the correspondingly low Bowen ratio limits the depth of the well-mixed midday boundary layer although the influence of soil moisture on boundary layer depth appears to be somewhat weaker than in BB.

The $\Theta_E$ diurnal cycle composite, however, shows significant and critical differences to the FIFE $\Theta_E$ composite (Fig. 9). For wet soils, the FIFE composite $\Theta_E$ peaks near 1500 LT, having risen steadily throughout the morning and afternoon. In contrast, over dry soils $\Theta_E$ only rises slightly over the course of the day. In HadAM3, although the wet soil composite exhibits a slightly higher midday $\Theta_E$ than the dry soil composite, due largely to differences in mean specific humidity, the diurnal evolution of $\Theta_E$ is essentially the same for both wet and dry soils. An early morning abrupt increase in $\Theta_E$ is associated with a sharp increase in low-level specific humidity related to strong evaporation into a shallow early morning boundary layer. After the swift initial increase, the $\Theta_E$ profile levels out and remains relatively constant throughout the remainder of the morning and early afternoon as increasing surface temperatures are balanced by a drop in specific humidity associated with turbulent mixing of the moist near-surface air into the deepening boundary layer. The absence of a differential buildup of moist static energy over wet and dry soils, as seen in BB, may partially explain the lack of a soil moisture–precipitation feedback in the model. The impact of the poor representation of the precipitation and $\Theta_E$ diurnal cycles on the soil moisture–precipitation feedback is currently being more thoroughly investigated and will be the subject of a future publication.

5. Summary

A soil moisture–precipitation feedback is virtually nonexistent in HadAM3, in contrast to other comparable AGCMs, many of which exhibit a significant feedback in many regions around the world (Koster et al. 2002, 2004). Although the relationship between soil moisture and evaporation is relatively strong in HadAM3 in arid and semiarid regions, it is near zero over much of the rest of the global land surface, including areas such as the southeastern United States, a region that is identified by Findell and Eltahir (2003b) as an area where typical atmospheric profiles enable convection to be responsive to soil-moisture-regulated surface heat flux forcing. When the soil model is modified to globally increase soil moisture stress on transpiration and encourage soil moisture variability, the soil mois-
ture–evaporation relationship is strengthened, but the soil moisture–precipitation feedback remains weak.

In a semiarid region, the model appears able to represent key boundary layer processes involved in the theoretical indirect soil moisture–precipitation feedback. Relationships between soil moisture and the Bowen ratio, total heat flux into the atmosphere, and, to a certain degree, boundary layer depth are consistent

**Fig. 8.** Composite diurnal cycles of latent (thick lines) and sensible (thin lines) heat flux, evaporative fraction \([LH/(LH + SH)]\), boundary layer depth, and net radiation are shown. Composites are compiled relative to daily mean SMF. Wet soil moisture days (dashed lines) are defined as days when \(SMF_1 \geq SMF_1 - 0.5\sigma_{SMF_1}\). Dry soil moisture days (solid lines) are defined as days when \(SMF_1 \leq SMF_1 - 0.5\sigma_{SMF_1}\). Only days where morning cloud fraction < 0.25 are considered. The mean SMF for wet and dry soil days is 0.51 and 0.16, respectively.

**Fig. 9.** Daytime \((\Theta, q)\) plots, composited by soil moisture as in Fig. 8. Points are plotted every hour between 0600 and 1800 LT; solid lines are dry soil composite, and dashed lines are wet soil composite. Sloping lines indicate lines of constant \(\Theta_c\). Composites are shown for (left) HadAM3 and (right) FIFE (reproduction of Fig. 22 of Betts and Ball 1995).
with observed relationships seen in diurnal cycle composites for wet and dry soils. However, the diurnal evolution of boundary layer moist static energy appears largely independent of soil moisture, which, together with an overly frequent rainfall bias, may be a potential source of the weak soil moisture–precipitation feedback in HadAM3.

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