Land–Atmosphere Coupling Strength in the Global Forecast System

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ABSTRACT

The operational coupled land–atmosphere forecast model from the National Centers for Environmental Prediction (NCEP) is evaluated for the strength and characteristics of its coupling in the water cycle between land and atmosphere. Following the protocols of the Global Land–Atmosphere Coupling Experiment (GLACE) it is found that the Global Forecast System (GFS) atmospheric model coupled to the Noah land surface model exhibits extraordinarily weak land–atmosphere coupling, much as its predecessor, the GFS–Oregon State University (OSU) coupled system. The coupling strength is evaluated by the ability of subsurface soil wetness to affect locally the time series of precipitation. The surface fluxes in Noah are also found to be rather insensitive to subsurface soil wetness. Comparison to another atmospheric model coupled to Noah as well as a different land surface model show that Noah is responsible for some of the lack of sensitivity, primarily because its thick (10 cm) surface layer dominates the variability in surface latent heat fluxes. Noah is found to be as responsive as other land surface models to surface soil wetness and temperature variations, suggesting the design of the GLACE sensitivity experiment (based only on subsurface soil wetness) handicapped the Noah model. Additional experiments, in which the parameterization of evapotranspiration is altered, as well as experiments where surface soil wetness is also constrained, isolate the GFS atmospheric model as the principal source of the weak sensitivity of precipitation to land surface states.

1. Introduction

Land–atmosphere coupling strength, a fundamental element of the model-generated climate system, has been examined closely in the Global Land–Atmosphere Coupling Experiment (GLACE; Koster et al. 2004, 2006; Guo et al. 2006). GLACE measured the impact of soil moisture on precipitation and other atmospheric processes for boreal summer, and quantified coupling strength among 12 participating models. One participant in that study was the operational weather and climate forecast model for the National Centers for Environmental Prediction (NCEP). The Global Forecast System Model (GFS) coupled to the Oregon State University (OSU) land surface model (LSM) showed almost no impact of variations of subsurface soil wetness on precipitation and was the weakest of the participating models in most metrics of coupling strength. The current version of the GFS used for operational weather prediction has been coupled to the newer Noah LSM since 2005. The operational NCEP Climate Forecast System (CFS) was still coupled to the OSU LSM into 2010 (Saha et al. 2006), but in its next generation is adopting the configuration with Noah, consistent with the CFS reanalysis and reforecast (CFSRR) model configuration.

Extensive studies of the Noah LSM over the continental United States in uncoupled land data assimilation (Mitchell et al. 2004), over the globe in the Second Global Soil Wetness Project (GSWP-2; Dirmeyer et al. 2006) and Global Land Data Assimilation System (GLDAS; Rodell et al. 2004), and in coupled numerical model...
assimilations (Mesinger et al. 2006; De Haan et al. 2007) reveal that Noah performs well both regionally and globally. Noah also has shown substantial evaporation sensitivity to soil moisture variations when coupled to NCEP’s regionalEta Model (Berbery et al. 2003) and improved performance in forecasting temperature and humidity in the NCEP mesoscale Eta Model (Ek et al. 2003). Noah is also employed as the LSM for the Weather Research and Forecast Model (WRF; Skamarock et al. 2005) used for regional operational weather forecasts over North America.

Given the weak land–atmosphere coupling seen in GLACE, it is natural to ask what the cause was and whether the most recent operational NCEP GFS–Noah model shows a stronger land–atmosphere coupling, which in turn may enhance precipitation prediction skill. After all, a great deal of effort has gone into the development and implementation of Noah for operational forecast models (e.g., Koster et al. 2009). In this study, we compare simulations with the current GFS–Noah model to previous simulations from GLACE, as well as with simulations where Noah has been coupled into a different atmospheric general circulation model (AGCM) where it has shown appreciably stronger land–atmosphere coupling strength than with its native AGCM. These experiments allow us to isolate the impact of land and atmosphere model parameterizations on coupling strength.

The rest of the paper is structured as follows: section 2 describes the coupled model and experiments, section 3 presents results, and the conclusions are provided in section 4.

2. GFS–Noah model and GLACE experiments

The AGCM used is a recent operational version of GFS used at NCEP (Moorthi et al. 2001; Saha et al. 2006) from late 2008. Except for having a coarser horizontal resolution, it is the same as that used for operational weather forecasting. GFS has 64 vertical sigma levels with a spectral triangular truncation of 62 waves (T62) in the horizontal resolution (approximately 1.9° grid). The key model physical parameterizations include the simplified Arakawa–Schubert convection scheme, longwave and shortwave radiation, explicit cloud microphysics, nonlocal vertical diffusion, and gravity wave drag (Pan and Wu 1995; Hong and Pan 1996). The Noah LSM (version 2.7) is the land model (Chen et al. 1996; Koren et al. 1999; Ek et al. 2003), which has four soil layers to a depth of 2 m (0.1, 0.3, 0.6, and 1.0 m from the surface downward). Compared to the OSU scheme, it contains improved treatment of frozen soil, ground heat flux, and energy–water balance at the surface, along with reformulated infiltration and runoff functions and an upgraded vegetation fraction. The initial atmospheric conditions and initial land states for Noah are taken from the NCEP–Department of Energy (DOE) Second Atmospheric Model Intercomparison Project (AMIP-II) reanalysis (Kanamitsu et al. 2002).

A set of complimentary experiments has been conducted. The first experiment serves as a control run, in which an 18-yr GFS–Noah simulation is forced by prescribed climatological sea surface temperature (SST) that minimizes the impact of the SST on land–atmosphere coupling strength. In the second set, GLACE type experiments are conducted. There are three separate 16-member ensembles of GFS–Noah simulations, each simulation covering the period of 1 June–31 August (see details in Koster et al. 2006). The 16 ensemble members vary only in atmospheric and land initial conditions, taken from 1 June in the last 16 yr of the simulation generated from the first experiment (the GFS–Noah control run). In the base ensemble (case W), one member is chosen arbitrarily to be the test basis and all land surface state variables at each model time step from that member are saved. The other two ensembles use the saved states of the test basis from case W as prescribed boundary conditions. In one ensemble, all land surface state variables (soil temperature and moisture at four soil layers, canopy interception, and snow cover) are specified at each time step (case R). In the other, only soil wetness for the three lowest soil layers is specified (case S).

Comparison of cases R and S to case W gives an indication of the impact of locking identical land states into every member of an ensemble. If the land surface were exerting systematic influence on the atmosphere, the ensemble spread in cases R or S should be reduced compared to case W. This can be quantified for any flux or state variable $x$ by the quantity $\Omega$ (Koster et al. 2006):

$$\Omega = \frac{n\sigma_r^2 - \sigma_x^2}{(n-1)\sigma_x^2},$$

where, for ensemble size $n$, 6-day means are calculated throughout the season [14 for June–August (JJA) with the first week excluded, per GLACE standards], the caret indicates variance calculated across the ensemble-mean 6-day averages, while the other variances are across all time periods and ensemble members. If the time series of each ensemble member is identical, $\Omega$ will be equal to unity. An uncorrelated ensemble of time series will yield $\Omega \approx 0$. The $\Omega$ is calculated as a function of space for each case, and an increase from case W to the other cases indicates the quantity is showing a more correlated evolution in time among ensemble members, and thus the constrained land surface states must likewise be constraining variable $x$. 
3. Results

The GFS–Noah control run reveals reasonable mean states of surface air temperature, net radiation, and surface fluxes (latent and sensible). Here, we present a brief comparison of evaporation, precipitation and soil wetness between the current GFS–Noah model and the GFS–OSU model that was used in the original GLACE study. The soil wetness (SW) was calculated as defined by GLACE, that is, “vertically integrated soil moisture above the wilting point divided by the maximum allowable soil moisture above the wilting point” (Seneviratne et al. 2006). Figure 1 shows terrestrial evaporation, precipitation and soil wetness for GFS–Noah (left column) and GFS–OSU (center column) averaged over JJA. Overall, both models have the same general precipitation patterns, with high values over Southeast Asia, equatorial Africa, and the northern part of South America, and low precipitation in desert areas. GFS–Noah has higher mean rainfall over parts of central North America, the Tibetan foothills, and much of South America. The GFS–OSU configuration has higher precipitation rates over much of the tropics, as well as northern Europe, eastern Asia including Siberia, and northwestern North America. However, both models overestimate the most intense precipitation compared to observationally calibrated GSWP-2 forcing data (right column; Zhao and Dirmeyer 2003). As expected, the mean soil wetness in the mid-latitudes and the tropics is generally high in regions of high precipitation. The relatively high values found in the northern high latitudes are likely linked to snow-melt. The soil moisture simulations of GFS–Noah and GFS–OSU bracket that of the GSWP-2 multimodel analysis (Dirmeyer et al. 2006) with GFS–Noah having overall higher soil wetness. Evaporation is generally highest in regions of high soil moisture, and tends to be higher than GSWP-2 in humid regions for both model configurations. GFS–Noah is especially high over the Great Plains, while GFS/OSU places strong evaporation along the East Coast.

Using the $\Omega$ diagnostic defined by Koster et al. (2006), we examine the land surface’s control on subseasonal precipitation variability. The top panels in Fig. 2 show the global fields of $\Omega_p(W)$ (i.e., $\Omega$ for precipitation from the W ensemble), and $\Omega_p(S) - \Omega_p(W)$ for GFS–Noah. As explained in Koster et al. (2006), $\Omega$ measures the coherence within the ensemble members, while $\Omega_p(S) - \Omega_p(W)$ indicates to what degree land subsurface conditions exert some control on the variable in question. In other words, $\Omega_p(W)$ reflects the extent to which low-frequency seasonal variations lead to intraensemble similarity in the precipitation rates, while $\Omega_p(S) - \Omega_p(W)$ isolates the contribution of subsurface soil moisture to precipitation variability on synoptic to subseasonal time scales.
The high values of $\Omega_p(W)$ in Fig. 2 tend to be clustered in the tropics and a few extratropical regions, especially over sub-Saharan Africa, Tibet, the Caucasus, and northwestern North America. This pattern is somewhat more consistent with most of the other 11 GLACE models than was GFS–OSU (see Fig. 2 in Koster et al. 2006). The regions of high $\Omega_p(W)$ tend to be where the seasonal evolution of precipitation across JJA is robust and controlled by regional SSTs—predominantly monsoon regions and other areas undergoing a strong transition in rainfall.

In Fig. 2 $\Omega_p(S) - \Omega_p(W)$ does not show cohesive “hot spots” found over the large regions of Africa, central North America, parts of China, and India as in the multimodel average of Koster et al. (2006). Instead, the values are small and scattered across the globe, with nearly as much area covered by negative as positive values. In other words, GFS–Noah shows a weak coupling strength where many independent models agree that the land–atmosphere coupling is important. As in Koster et al. (2006) and Guo et al. (2006), we focus on case S, atmospheric response to the deeper soil moisture states (a slowly varying field), and return to case R at the end of the section.

Given that evapotranspiration (ET) is the key link between soil moisture anomalies and precipitation, the bottom panels of Fig. 2 show the global distribution of $\Omega_e(W)$ and $\Omega_p(S) - \Omega_p(W)$. The $\Omega_e(W)$ shares some of the same regions of the tropics where the onset of the summer monsoon is causing a strong trend of terrestrial moisture across the season. Across much of the Northern Hemisphere, the summer drying trend is reflected as strong coherence in evaporation. The difference $\Omega_p(S) - \Omega_p(W)$ highlights a few areas in the tropics, monsoon areas, the Great Plains of North America, and many agricultural regions of Eurasia, but most areas do not show a robust response of evaporation to soil wetness. This is similar to the GFS–OSU model used in GLACE, but unlike most of the other models in that experiment (Guo et al. 2006).

Additional model configurations are used to understand what may contribute to the weak land–atmospheric coupling strength in GFS–Noah. One has the same Noah land model coupled to a recent version (version 3.2) of the Center for Ocean–Land–Atmosphere Studies (COLA) atmospheric model (Misra et al. 2007); another is the COLA atmospheric model coupled with a new version of its native Simplified Simple Biosphere (SSiB) land model (for details see Wei et al. 2010a). They are labeled as COLA–Noah and COLA–SSiB, respectively, in Fig. 3. GLACE-type experiments are also performed with these two couple models. It is striking to see that COLA–SSiB places high values of $\Omega_p(S) - \Omega_p(W)$ in India and central North America, as GLACE identified, while no large...
patches of high $\Omega_p(S) - \Omega_p(W)$ can be seen in COLA–Noah over those areas. Both configurations show coupling from soil wetness to precipitation over the Sahel and northern South America. Overall COLA–SSiB is somewhat stronger than COLA–Noah in $\Omega_p(S) - \Omega_p(W)$, yet both are much stronger than seen for GFS–Noah in Fig. 2. This also applies to the global fields of $\Omega_p(S) - \Omega_p(W)$. This indicates that the land–atmosphere coupling strength is somewhat stronger in COLA–SSiB than in COLA–Noah. The only difference between COLA–SSiB and COLA–Noah is the land model. Therefore, the weak land–atmosphere coupling strength in GFS–Noah may relate to the details of the land surface parameterization. However, we also see that COLA–Noah coupling strength is stronger than GFS–Noah. So there may also be an atmospheric component to the weak results for the NCEP model as well. Specifically, we see that in $\Omega_p(S) - \Omega_p(W)$, the connection from soil wetness to evaporation is quite similar in strength and pattern for COLA–Noah (Fig. 3) and GFS–Noah (Fig. 2). The global mean for COLA–SSiB is 3½ times greater than for COLA–Noah and nearly 5 times greater than GFS–Noah, suggesting...
the strength of this terrestrial leg of the land–atmosphere coupling is not very sensitive to the overlying AGCM. The spatial pattern of $\Omega_P(S) - \Omega_P(W)$ for COLA–Noah is similar to both GFS–Noah and COLA–SSiB (spatial correlations significant at $p = 0.05$ and $p < 0.01$, respectively) whereas GFS–Noah and COLA–SSiB are negatively correlated with each other ($p = 0.03$).

The remainder of this paper focuses on a diagnosis of the Noah land surface model and GFS AGCM as potential sources for this weak coupling. It has been suggested that details of the land model parameterization, particularly those associated with soil water–limited transpiration and how it relates in magnitude to bare soil evaporation and canopy interception loss, may explain the low $\Omega_E(S) - \Omega_E(W)$ in GFS–Noah (Guo et al. 2006). The details of the global partitioning of ET into transpiration from the plant canopy ($E_T$), evaporation of soil water from bare soil ($E_S$) and direct evaporation of canopy intercepted water ($E_C$) for boreal summer are presented in Fig. 4. Although the actual global partitioning of ET is not well known (Lawrence et al. 2007), a recent estimate is available from the multimodel analysis of the GSWP2 (Dirmeyer et al. 2006). In GSWP2, the estimate of global ET partitioning is 49% $E_T$, 33% $E_S$, and 17% $E_C$ (Fig. 4, left column). Averaged over the land surface, the global ET partitioning in GFS–Noah is 40% $E_T$, 36% $E_S$, and 24% $E_C$ (Fig. 4, right column). This ranking of ET components is expected since during the warm season key components in the surface moisture flux are bare soil evaporation and plant transpiration. Transpiration is a dominant source of surface moisture flux especially in regions with large vegetation coverage (Ek et al. 2003). The global distribution of transpiration in GFS–Noah is high in southern Europe and the

![Fig. 4. The fraction of total ET from transpiration, canopy evaporation, and soil evaporation for (left) GSWP-2 multimodel analysis (MMA) and (right) GFS–Noah averaged over JJA. The global-averaged fractions are shown in the corners. The values for GSWP-2 add to <1 because additional processes besides the three listed here contribute to total ET.](image-url)
savannas of the Southern Hemisphere, but relatively low in the mid- and high-latitude forests of the Northern Hemisphere. When compared to the GSWP2 multimodel ensemble, the global distribution of $E_C$ is relatively high in vegetated areas.

Evaluations performed at NCEP indicated the Noah LSM had a large positive bias in summer transpiration over regions of nonsparse vegetation cover, such as the eastern United States, that was related to low canopy resistance in the Eta Model implementation (Ek et al. 2003; Campana and Caplan 2005). Changes to the current Noah model, which include attempts to correct the evaporation bias by increasing canopy resistance, were implemented in mid-2005. This modification reduced the surface evaporation over nonarid land regions during seasons of nonsparse green vegetation, in turn, increasing surface sensible heat flux. However, Fig. 4 suggests that perhaps net evaporation was corrected by introducing a compensating error to balance what may be excessive evaporation of intercepted water on the canopy.

Based on this evaluation, two additional experiments have been performed to see the effect of canopy resistance changes on the land–atmosphere coupling strength: 1) resetting the table of canopy resistance values of the Noah LSM to their original settings (labeled “RSMTBL”); and 2) setting canopy resistance factors that relate to atmospheric conditions at their maximums so that only the resistance factor computed from the soil moisture is constrained (labeled “CANRES”). The second experiment is done because canopy resistance in the Noah LSM formulation is modeled as a function of soil moisture availability and atmospheric conditions (i.e., solar insolation, temperature, and humidity). Both RSMTBL and CANRES experiments display a clear increase in transpiration in the mid- and high-latitude forests of the North Hemisphere, southern Europe, and the savannas of the Southern Hemisphere (figure not shown). The global mean partitioning of plant transpiration increases to 43% and 46%, respectively, for RSMTBL and CANRES.

Do these changes of land surface scheme significantly alter the behavior of the land–atmosphere coupling in GFS–Noah? Figure 5 presents the GLACE metric $\Omega_p(S) - \Omega_p(W)$ (left column), and $\Omega_e(S) - \Omega_e(W)$ (right column) for RSMTBL experiment (top row) and CANRES (bottom row). RSMTBL has the greatest impact over sub-Saharan Africa north of the equator while the CANRES experiment affects the model performance mainly over South America. But overall, these changes to the Noah LSM do not significantly increase land–atmosphere coupling strength when sub-surface soil wetness is specified. The atmosphere simply does not perceive the changes. The lack of atmospheric response to the changes in the LSM may also contribute to the weak land–atmosphere coupling.

To further explore this possibility, the indices defined in Guo et al. (2006) are used to measure the strength of coupling along the branches of the water cycle from soil wetness to precipitation. Guo et al. (2006) break down
Table 1. Globally averaged (over nonice land points) land–atmosphere coupling strength, computed between S and W cases, for GFS–OSU from the original GLACE experiment, GFS–Noah, COLA–Noah, and COLA–SSiB. As in Table 1 of Guo et al. (2006), shown are metrics for the path from soil wetness to precipitation (SW–P), soil wetness to ET (SW–ET), and the metric for ET to precipitation (ET–P).

<table>
<thead>
<tr>
<th>Model</th>
<th>SW–P</th>
<th>SW–ET</th>
<th>ET–P</th>
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<tbody>
<tr>
<td>GFS–OSU (GLACE)</td>
<td>−0.004</td>
<td>0.024</td>
<td>−0.167</td>
</tr>
<tr>
<td>GFS–Noah</td>
<td>−0.007</td>
<td>0.067</td>
<td>−0.109</td>
</tr>
<tr>
<td>COLA–Noah</td>
<td>0.016</td>
<td>0.145</td>
<td>0.110</td>
</tr>
<tr>
<td>COLA–SSiB</td>
<td>0.035</td>
<td>0.280</td>
<td>0.125</td>
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</table>

The results from Guo et al. (2006) provide a clue to the weakness of the coupling strength when the Noah LSM is used. They found that the R case for GFS–OSU had a much stronger SW–ET connection than for the S case—the largest difference for any model in GLACE. Both the former OSU LSM and Noah share an important characteristic—they both have a thick (10 cm) surface layer. Furthermore, both include parameterizations for the extraction of soil moisture by vegetation via transpiration (a simple canopy resistance term for OSU, a more physical scheme based on root distributions for Noah) that taps this thick surface layer. The Noah LSM has roots in the top three soil layers (1 m) for forests, and only the top two layers for grasses, savannah, crops, etc. (40 cm). The top layer accounts for 25% of the reservoir for transpiration moisture in grassland areas, and 10% for forests.

Two other GLACE models have LSMs with 10-cm-thick surface layers—the Canadian Climate Centre (CCCma) model and the Third Hadley Centre Atmospheric Model (HadAM3). These schemes also draw moisture for transpiration from this top layer. HadAM3 was the second-weakest GCM in GLACE in terms of coupling strength, but specifically in the category of SW–ET, HaDAM3 was in the middle of the pack, and CCCma was the second strongest. However, the Canadian land surface scheme (CLASS; Verseghy 2000) has a 4.1-m-deep soil column from which roots draw moisture (surface is only 2.4% of the transpiration reservoir), and the Met Office Surface Exchange Scheme (MOSES; Cox et al. 1999) draws transpiration water from the top 3 m of soil for forests and 1 m for grasslands (3.3% and 10%, respectively). Therefore, a much greater proportion of total ET in Noah and OSU come from the uppermost soil layer than is typical for other LSMs, including SSiB or even other LSMs with equally thick top soil layers. The consequence of this characteristic of the Noah and OSU models is that the S experiment of GLACE does little to constrain the soil moisture that mainly affects surface fluxes. In the R case, soil wetness and temperature in all soil layers are constrained, and the Noah LSM behaves on par with other schemes (cf. Fig. 9 of Guo et al. 2006). Table 2 shows the same metrics as Table 1, but for coupling strength calculated when soil moisture and temperature in all soil layers is constrained in the test (R) case. No R case was run for the current version of COLA–SSiB, so that model is not present in Table 2. Compared to Table 1, the SW–ET branch is now much stronger. This translates to an increase in the full-path coupling SW–P, here measured as the global mean of \(\Omega_p(R) \) – \(\Omega_p(W)\). However, comparison to Fig. 7 of Guo et al. (2006) shows this still to be on the low end for global weather and climate models, most of which have values ranging from 0.05 to 0.16.

Table 2. As in Table 1, but for the R case vs the W case.

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<tr>
<th>Model</th>
<th>SW–P</th>
<th>SW–ET</th>
<th>ET–P</th>
</tr>
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<tbody>
<tr>
<td>GFS–OSU (GLACE)</td>
<td>0.036</td>
<td>0.410</td>
<td>0.088</td>
</tr>
<tr>
<td>GFS–Noah</td>
<td>0.013</td>
<td>0.267</td>
<td>0.049</td>
</tr>
<tr>
<td>COLA–Noah</td>
<td>0.036</td>
<td>0.202</td>
<td>0.178</td>
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The ET–P branch is now seen to remain much weaker for GFS than for the COLA AGCM. When each are coupled to Noah, GFS has about one-third the coupling strength in the atmospheric branch of the hydrologic cycle compared to COLA. This suggests that one or more of the parameterizations in the GFS atmospheric model contribute to the weak land–atmospheric coupling, as the signal innate in the surface fluxes is not manifested in precipitation.

4. Conclusions

In this study, GLACE-type experiments were used to investigate the land–atmosphere coupling strength in the most recent GFS–Noah configuration. It turns out that the GFS–Noah model, when compared to the models in the basic GLACE experiment, still shows a weak response of precipitation to the prescribed subsurface soil moisture, much like its predecessor GFS–OSU. Noah model fluxes respond rather weakly to the state of subsurface soil wetness. We find that the thick 10-cm surface layer, which is also involved in supplying moisture for transpiration via root water uptake, dominates the variability in surface latent heat fluxes. Surface fluxes in Noah do respond robustly to surface wetness and temperature variations, but these quantities carry little impetus to enhance predictability of precipitation in GFS. In addition, experiments to isolate the components of ET suggest that changes in the ET parameterization of Noah land model scheme do not alter GFS–Noah coupled behavior in terms of the sensitivity of precipitation to subsurface soil moisture.

We find a relatively weak atmospheric response to the changes in land state in the GFS. Such decoupling is likely to pose great limitations on the role of soil moisture in seasonal precipitation forecasts in the GFS modeling system. Since the linkages among land surface conditions, atmospheric dynamics, and precipitation are constrained by the moisture continuity equation, the lack of atmospheric response to the land states seen in GFS may be attributed to how moisture processes are formulated in the model, particularly moist convection or boundary layer growth. After all, both atmosphere and land contribute to the behavior of the coupled system (Wei and Dirmeyer 2010). Additional analysis and modeling studies are needed to understand the specific cause of weak land–atmosphere interaction in GFS. For instance, a moisture budget analysis can be conducted to investigate the relative contributions of externally advected water vapor versus locally recycled (i.e., evapotranspired) moisture for the summer-season precipitation.

Of course, determination of the actual strength and distribution of land–atmosphere coupling in the real world is still an open question. Wei et al. (2010b) suggest estimates from GLACE may be too strong, based on systematic errors in the spectrum of precipitation variability in current weather and climate models. Nevertheless, the low sensitivity of GFS to land surface variations should be fully diagnosed to determine whether it hampers model forecast ability.

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