Interannual Variability of Land–Atmosphere Coupling Strength

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ABSTRACT

Recent studies in the Global Land–Atmosphere Coupling Experiment (GLACE) established a framework to estimate the extent to which anomalies in the land surface state (e.g., soil moisture) can affect rainfall generation and other atmospheric processes. Within this framework, a multiyear GLACE-type experiment is carried out with a coupled land–atmosphere general circulation model to examine the interannual variability of land–atmosphere coupling strength. Soil wetness with intermediate values are in the range at which rainfall generation, near-surface air temperature, and surface turbulent fluxes are most sensitive to soil moisture anomalies, and thus, land–atmosphere coupling strength peaks in this range. As a result, the “hot spots” with strong land–atmosphere coupling strength appear in regions with intermediate climatological soil wetness (e.g., transition zones between dry and wet climates), consistent with previous studies. Land–atmosphere coupling strength experiences significant year-to-year variation because of interannual variability of soil moisture and the local spatiotemporal evolution of hydrologic regime. Coupling strength over areas with dry (wet) climate is enhanced during wet (dry) years since the resultant soil wetness enters into the sensitive range from a relatively insensitive range, and soil moisture can have stronger potential impact on surface turbulent fluxes and convection. On the other hand, land–atmosphere coupling strength over areas with wet (dry) climate is weakened during wet (dry) years since the soil wetness moves further away from the sensitive range. This results in a positive correlation between the land–atmosphere coupling strength and soil moisture anomalies over areas with dry climate and a negative correlation over areas with wet climate.

1. Introduction

Soil moisture, a slowly varying state variable of the land surface, is recognized as one of the largest sources to atmospheric predictability, particularly over midlatitude continents during summer on subseasonal to seasonal time scales (e.g., Shukla and Mintz 1982; Dirmeyer 2000; Koster et al. 2000). A sufficient condition for a significant contribution to atmospheric predictability from soil moisture is that the atmospheric variables and turbulent fluxes must be sensitive to soil moisture anomalies (Koster and Suarez 2003). Thus, land–atmosphere coupling strength, intuitively defined as the extent to which rainfall generation and other atmospheric processes could be affected by anomalies in land surface state (e.g., soil moisture), has received increased attention in the last decade (e.g., Dirmeyer 2001; Koster et al. 2002, 2004; Lawrence and Slingo 2005; Koster et al. 2006; Guo et al. 2006; Seneviratne et al. 2006; Wang et al. 2007; Wei and Dirmeyer 2010; Zeng et al. 2010; Zhang et al. 2011).

Previous studies mainly focused on the spatial distribution of land–atmosphere coupling and identified the “hot spot” regions with strong soil moisture–precipitation interaction. They found that these regions are mainly located in the transition zones with intermediate soil wetness between dry and wet climate regimes. However, less attention has been paid to the interannual variability of land–atmosphere coupling. In fact some studies do imply the existence of such temporal variability. Koster et al. (2011) found that wet and dry soil moisture initialization has asymmetric contributions to precipitation and air temperature forecast skill over North America as evaporation is sensitive to soil moisture variations near the dry
climate regime, whereas it becomes inactive toward a wet climate. Results from Seneviratne et al. (2006) show that land–atmosphere coupling could influence summer climate variability and the potential migration of transitional climate zones in Europe as a consequence of global warming. The geographic locations of hot spot regions in turn might be changed since they follow the migrated transition zone. These studies have suggested that interannual variability of soil moisture could result in year-to-year variability in land–atmosphere coupling strength. The current study examines the interannual variation in land–atmosphere coupling strength and how it is related to soil moisture anomalies. This is of interest for two main reasons. First, it is an intrinsic characteristic of the temporal and spatial variation of land–atmosphere interaction and could be used to facilitate understanding the land surface’s role in the temporal variability of atmospheric predictability and forecast skill as well as its varying role within a changing climate. Second, it is useful for strategic climate monitoring since observations of soil moisture in regions with strong soil moisture–precipitation coupling may be more useful than measurements elsewhere. Because of both the lack of observational data for soil moisture and surface turbulent fluxes at regional to continental scales and the complex intertwined connection between soil moisture and rainfall generation, coupling

Fig. 1. Multiyear average of (top) land–atmosphere coupling strength $[\Omega_S - \Omega_W]$, (middle) the proxy of the coupling strength $[r(SM, E)E]$, and (bottom) geographic distribution of climatologic soil wetness.
strength is tremendously difficult to determine from observations, and its evaluation must rely on the use of numerical climate models. In this regard, a modeling framework has been established in the pilot and first phases of the Global Land–Atmosphere Coupling Experiment (GLACE; Koster et al. 2002, 2006) to objectively measure the land–atmosphere coupling strength in atmospheric general circulation models (AGCMs). This framework has been extensively used in the aforementioned studies. Section 2 describes the model and GLACE-type experiment design used in this study. Section 3 provides an analysis of the interannual variability in coupling strength. Further discussion and a summary of our findings are presented in section 4.

2. Model and experimental design

This section describes the Center for Ocean–Land–Atmosphere Studies (COLA) AGCM and the specific details of the GLACE experiments performed. Koster et al. (2006) provides a detailed description of GLACE-type experiments and their basic design.

a. COLA AGCM

A recent version (v3.2) of the COLA AGCM (Misra et al. 2007) is used in this study. The horizontal resolution is T62 (about 1.9° × 1.9°), and the model has 28 vertical levels. The model uses the relaxed Arakawa–Schubert deep convection scheme, a nonlocal boundary layer vertical diffusion scheme, the longwave radiation scheme, and the cloud radiation scheme of the National Center for Atmospheric Research (NCAR) Community Climate Model, version 3 (CCM3). The land surface scheme is an updated version of the Simplified Simple Biosphere (SSiB) model, which has some improvements over earlier versions of SSiB (Dirmeyer and Zeng 1999): the number of soil layers has been increased from three to six, a three-layer snow model (Stieglitz et al. 2001) has been coupled to SSiB to replace the original simple snow parameterization, and the Community Land Model (CLM) scheme (Oleson et al. 2004) is used for calculating soil thermal conductivity and soil temperature. Preliminary results show that the new version has corrected a dry bias over much of the globe when coupled with COLA AGCM, and the snow cover is better simulated (Wei et al. 2010).
b. Experiment design

GLACE-type experiments consist of two separate 16-member ensembles of AGCM integrations, each simulation covering the period of 1 June to 31 August. The first ensemble, denoted $W$ for “write,” is run with the land states of each ensemble member varying and interacting with the atmosphere in a fully coupled mode; soil moisture values for each of the six soil levels from one ensemble member are recorded at each time step. In the second set of simulations, denoted $S$ for “subsurface,” the simulated soil moistures in layers 2–6 for each member are replaced at each time step with the single geographically and temporally varying soil moisture fields recorded from the $W$ experiment. By measuring the degree of similarity among the members of an ensemble for a variable and then comparing this similarity between ensembles, the impact of soil moisture on this variable is isolated, and the degree to which the atmosphere responds consistently to specific anomalies in soil moisture (i.e., land–atmosphere coupling strength) is quantified. Since the variability of root-zone soil moisture is much slower than that of other land variables, subsurface soil moisture is the major potential predictor for subseasonal predictability. Here only the subsurface soil moisture is prescribed in the $S$ experiment, and the impact of subsurface soil moisture on precipitation is considered.

In the standard GLACE experiments, the AGCM simulations are forced only by a particular year’s sea surface temperatures (SSTs), and the subsurface soil moisture values for each ensemble member in the $S$ experiment are only prescribed to a single soil moisture field from the $W$ experiment. The results obtained may only reflect the land–atmosphere coupling strength under one specific climate situation and are subject to undersampling in the choice of ensemble member for the soil moisture specification. In the current study, GLACE-type experiments are performed for each of the 25 summers spanning 1982–2006; four different ensemble members are randomly chosen each year from the total 16 members to write out the soil moisture states.
in $W$ experiments, and they are used to prescribe the subsurface soil moisture in $S$ experiments. This is equivalent to 100 independent GLACE-type experiments. For each year, the simulations are forced by the SSTs from a weekly observational dataset (Reynolds et al. 2002), atmospheric initial conditions are derived from the National Centers for Environmental Prediction (NCEP)–NCAR Global Reanalysis 1, and the initial land prognostic variables are obtained from offline simulations with ensemble members perturbed from each other by being chosen from different days within 8 days of the start date in that year.

3. Results

The $\Omega$ diagnostic defined by Koster et al. (2002) is used to measure the land–atmosphere coupling strength. Since the only difference between $W$ and $S$ ensembles is that the subsurface soil moisture is prescribed to be the same among the ensemble members in the $S$ ensembles while moisture and precipitation interact normally in the $W$ ensembles, the changes in the degree of similarity in precipitation time series among ensemble members, measured by the change of $\Omega$ diagnostic between the $S$ and $W$ ensembles, are attributable to the specification of subsurface soil moisture anomalies and the strength of soil moisture–precipitation interaction. Given that the simulations from 1 June to 31 August are 92 days long, and that the first 8 days are ignored to reduce the impact of initial “shocks” in the $S$ ensembles, aggregation of precipitation output from each simulation into time series of 6-day totals provides a dataset consisting of fourteen 6-day totals. The $\Omega$ diagnostic is computed across 224 aggregated 6-day totals (16 ensemble members times 14 intervals in each simulation time series) for each year and each of four choices of ensemble member for specifying the soil moisture (a sample size of 22,400). The average of the $\Omega$ diagnostic among the four cases is computed to reduce the dependence of land–atmosphere coupling index on the choice of ensemble member and is used to represent the coupling strength in that year.

a. Multiyear average of land–atmosphere coupling strength

The multiyear average of the change between experiments in $\Omega [\Omega_p(S) - \Omega_p(W)]$ is computed to identify robust and climatologic regions of significant subsurface soil moisture impact on precipitation (top panel in FIG. 4. (top) Standard deviation of land–atmosphere coupling strength $\Omega_p(S) - \Omega_p(W)$ among 25 yr and (bottom) correlation between $\Omega_p(S) - \Omega_p(W)$ and the simulated average soil wetness in JJA.
The hot spots for precipitation obtained from a multimodel average (Koster et al. 2004, 2006; Guo et al. 2006) are well reproduced by the multiyear average with the COLA AGCM: large values in the Sahel in Africa, the central Great Plains of North America, parts of India, and a few additional regions in Eurasia and the tropics. To facilitate visual comparison, the hot spot regions used for regional calculations by Koster et al. (2004) and Guo et al. (2006) are indicated by the boxes in Fig. 1. Guo et al. (2006) separated the overall impact of soil moisture on precipitation into the ability of soil moisture to affect evaporation (terrestrial path) and the ability of evaporation to control precipitation. A number of studies (Dirmeyer et al. 2009; Guo et al. 2011) have shown that the terrestrial path of land–atmosphere coupling strength, characterized by the product of standard deviation of evaporation \(\sigma_E(W)\) with the temporal correlation between evaporation and soil moisture, serves as a good proxy of land–atmosphere coupling strength. The middle panel in Fig. 1 shows the multiyear average of this proxy. While the evaporation signal in the proxy could explain much of the geographical variation in \(\Omega_p(S) - \Omega_p(W)\) in the midlatitudes, there is an inconsistency between these two metrics over tropical regions. This is not entirely surprising given the complexity of the soil moisture–precipitation feedback processes (Lawrence and Slingo 2005). The coupling strength proxy is based on an assumption that the influence of anomalously soil on precipitation is via a direct hydrological cycle of water vapor. The disparities between the proxy and \(\Omega_p(S) - \Omega_p(W)\) reveals that the indirect feedback may dominate the feedback mechanism in these regions, considering how soil moisture affects precipitation through its influence on boundary layer characteristics and atmospheric stability. Notice that the large soil moisture impacts on precipitation generally occur in the transition zones between humid and arid climate, consistent with previous findings (Koster et al. 2004, 2006; Guo et al. 2006). The bottom panel in Fig. 1 shows the multiyear average of June–August (JJA) soil wetness. The areas with high soil wetness (green) are mainly in an energy-controlled evaporation regime, while the areas with low soil wetness (white) have weak evaporation variability. Neither of these areas have a strong land surface impact on precipitation, corresponding to small values of land–atmosphere coupling strength in the top two panels (Guo et al. 2006; Koster et al. 2011). Only the areas with intermediate soil wetness (red) belong to the soil moisture–controlled evaporation regime and also have large evaporation variability, and they are the climatological regions where soil moisture has potentially significant impact on precipitation.

Figure 3 illustrates the land–atmosphere coupling strength represented by \(\Omega_p(S) - \Omega_p(W)\) over North America for each of the 25 yr from 1982 to 2006. It is obvious that the coupling strength has experienced significant year-to-year variability. For example, the hot spot regions in 1987 and 1990 are quite different from those in 1988 or 1991. This variability is larger than what
would be expected from sampling error, even with four sets of ensembles generated for each year. Multiyear running means of coupling strength reveal low-frequency variations over the course of decades (not shown).

The top panel in Fig. 4 shows the global temporal standard deviation of $V_p(S) - V_p(W)$ among 25 yr. Over much of the midlatitudes, the standard deviation is as large as the magnitude of coupling strength itself, exhibiting significant interannual variability of coupling strength over the globe. Recall that the geographic distribution of coupling strength is highly dependent on the spatial pattern of the hydrological regimes based on a long-term average.

The interannual variability of coupling strength is naturally attributable to the evolution of hydrological conditions. The bottom panel in Fig. 4 displays the temporal correlation between $V_p(S) - V_p(W)$ and the simulated JJA average of soil wetness. The dots indicate grid cells where the correlation is significant at the 95% confidence level. It is found that over certain areas the coupling strength and soil wetness anomalies are significantly positive correlated while over other areas, often nearby, they have significant negative correlation.

The time series in Fig. 5 represent the yearly land–atmosphere coupling strength $V_p(S) - V_p(W)$ (solid

![Multi-year average of $\Omega(S) - \Omega(W)$](image1)

![Standard deviation of $\Omega(S) - \Omega(W)$](image2)

![Correlation between $\Omega(S) - \Omega(W)$ and soil moisture anomalies](image3)
lines) and soil wetness (dotted lines) averaged over the three major hot spots indicated by the boxes in Fig. 1. A significant interannual variability is found in the time series of land–atmosphere coupling strength, and there exists a clear negative correlation between the two variables for both the Great Plains and the Sahel while a positive correlation is present in northern India.

Figure 6 displays the multiyear average of the Ω diagnostic difference for the evaporation \( V_E(S) - V_E(W) \), the standard deviation of \( \Omega_E(S) - \Omega_E(W) \), and the correlation between \( \Omega_E(S) - \Omega_E(W) \) and soil moisture. Similar features are seen as in the maps relating precipitation and soil moisture, although the connection between evaporation and soil moisture is stronger. Positive correlations again are prevalent over arid regions and negative correlations dominate over areas of higher soil moisture.

The geographic distribution of the positive and negative correlations between land–atmosphere coupling strength and soil wetness anomalies is closely related to the spatial pattern of hydrological regimes. This is clearly seen in Fig. 7, illustrating the control of the hydrological regime on correlation between the anomalies of soil wetness and the Ω diagnostic difference for precipitation, near-surface air temperature, and latent and sensible heat fluxes. It shows that the correlations tend to be positive in the dry climate regime and negative in the wet climate regime. The transition in this climate model appears to be around a soil moisture value of 0.4–0.45. More than 40% of the variance in the correlations for the latent and sensible heat fluxes can be explained by the hydrologic climate regime, as can about a quarter of the variance for precipitation and near-surface air temperature.

The dependency of the land–atmosphere coupling strength variability on the hydrologic climate regime is not unexpected. As discussed in the last subsection, intermediate soil wetness is found to be the preferred range stimulating strong land–atmosphere interactions, and the closer to the sensitive range the soil wetness is, the stronger the coupling strength tends to be. Variability of coupling strength differs between dry and wet regimes depending on whether the soil wetness anomaly brings it closer to or further from the most sensitive range. Figure 8 presents the synthesis analysis of the Ω diagnostic difference for precipitation and evaporation.
partitioned according to the hydrological conditions. First, based on ranked JJA soil wetness anomalies at each land grid box, the 25 years were grouped into three categories: 8 dry years, 8 wet years, and 9 neutral years. Then the \( \Omega \) diagnostic difference \( \Omega(S) - \Omega(W) \) was averaged within each category for precipitation and evaporation. Scatterplots for the \( \Omega \) diagnostic difference for precipitation \( \Omega_{Pr}(S) - \Omega_{Pr}(W) \), left] and evaporation \( \Omega_{Ev}(S) - \Omega_{Ev}(W) \), right] versus the 25-yr climatological soil wetness are shown for the dry (top row), neutral

![Graphs showing scatterplots for \( \Omega \) diagnostic difference for precipitation and evaporation for dry, neutral, and wet years.](image)

**Fig. 8.** Scatterplots for \( \Omega \) diagnostic difference for (left) precipitation \( \Omega_{Pr}(S) - \Omega_{Pr}(W) \) and (right) evaporation \( \Omega_{Ev}(S) - \Omega_{Ev}(W) \) vs the climatologic soil wetness for (top to bottom) the dry, neutral, and wet years and the best-fit curves plotted together.
(second row), and wet years (third row). The best-fit curves for the dry, neutral, and wet years are shown together in the bottom row of Fig. 8. The contrast of the opposing behaviors between the dry and wet climate regime is clearly observed: the impacts of soil moisture on atmospheric variability are enhanced (weakened) during wet (dry) years in the dry climate regime, while the opposite holds true in the wet climate regime.

4. Summary and discussion

We have used a coupled land–atmosphere climate model to investigate the interannual variability of the strength of coupling between soil moisture, surface fluxes, and the atmosphere within the framework of past GLACE experiments. In those experiments, a single global soil moisture time series is used as a boundary condition for a single season’s ensemble simulation using observed SSTs. We have expanded the experiment to simulate 25 yr of boreal summer season climate and to perform multiple realizations of each year. This allows us to quantify the classical metrics of land–atmosphere coupling strength for individual years and to examine how they vary from year to year. As initial soil moistures in each year are derived from 16 consecutive days, their relatively small differences among ensemble members may lead to underestimation of intraensemble variability. However, this does not have large impacts on the temporal variability and geographic distribution of the land–atmosphere coupling strength, which is the main focus of this paper. Also, the results most closely represent the land–atmosphere coupling strength under the realistic hydrological conditions in each of the 25 yr.

Intermediate values of soil wetness are found to be optimal for surface turbulent fluxes, near-surface air temperature, and rainfall to show sensitivity to soil moisture anomalies. This property, inherent in the climate model, not only determines the hot spot regions of land–atmosphere coupling strength, but also the interannual variability of the coupling strength. When the soil wetness enters into the sensitive range from its climatologically insensitive position, the land–atmosphere coupling strength will be enhanced. This can occur when a normally arid region experiences wetter than normal conditions or when a humid region is in drought. On the contrary, the coupling strength is weakened when the soil wetness evolves farther from the intermediate sensitive range. There is a positive correlation between the land–atmosphere coupling strength and soil moisture anomalies over areas with a normally dry climate and a negative correlation over areas with a wet climate.

A key result of this study is the finding that hot spots of land–atmosphere interaction are not stationary but can vary from year to year. The fluctuations, however, are not random—much of the interannual variability is linked to the distribution of soil moisture anomalies. Therefore, the predictability (Guo et al. 2012) and prediction skill (Koster et al. 2011) that soil moisture provides to near-surface air temperature and precipitation on subseasonal to seasonal time scales may be predictable as well. Critical regions for soil moisture monitoring appear to extend somewhat outside the classical hot spot regions into adjacent semiarid and semihumid zones, varying as anomaly patterns change from year to year.


