Quantifying Spatiotemporal Variations of Soil Moisture Control on Surface Energy Balance and Near-Surface Air Temperature

CLEMENS SCHWINGSACKL, MARTIN HIRSCHI, AND SONIA I. SENEVIRATNE
Institute for Atmospheric and Climate Science, ETH Zürich, Zurich, Switzerland

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ABSTRACT
Soil moisture plays a crucial role for the energy partitioning at Earth’s surface. Changing fractions of latent and sensible heat fluxes caused by soil moisture variations can affect both near-surface air temperature and precipitation. In this study, a simple framework for the dependence of evaporative fraction (the ratio of latent heat flux over net radiation) on soil moisture is used to analyze spatial and temporal variations of land–atmosphere coupling and its effect on near-surface air temperature. Using three different data sources (two reanalysis datasets and one combination of different datasets), three key parameters for the relation between soil moisture and evaporative fraction are estimated: 1) the frequency of occurrence of different soil moisture regimes, 2) the sensitivity of evaporative fraction to soil moisture in the transitional soil moisture regime, and 3) the critical soil moisture value that separates soil moisture- and energy-limited evapotranspiration regimes. The results show that about 30%–60% (depending on the dataset) of the global land area is in the transitional regime during at least half of the year. Based on the identification of transitional regimes, the effect of changes in soil moisture on near-surface air temperature is analyzed. Typical soil moisture variations (standard deviation) can impact air temperature by up to 1.1–1.3 K, while changing soil moisture over its full range in the transitional regime can alter air temperature by up to 6–7 K. The results emphasize the role of soil moisture for atmosphere and climate and constitute a useful benchmark for the evaluation of the respective relationships in Earth system models.

1. Introduction
Earth’s climate is influenced by a large variety of complex processes and feedbacks. The relative importance of the single contributing factors depends on geographical location, orography, and land cover type. Over continents, the state of the land surface plays a crucial role for local climate (e.g., Koster et al. 2004; Seneviratne et al. 2006; Orth and Seneviratne 2017). It influences the atmosphere through various couplings and feedbacks. Of special importance is the role of soil moisture, since it affects various exchange processes at the surface. Seneviratne et al. (2010) provide an overview on the role of soil moisture for climate variability and land–atmosphere exchange. Based on both observational and modeling studies, various authors have shown that soil moisture has an impact on the evolution of near-surface air temperature (Seneviratne et al. 2006; Koster et al. 2009b; Jaeger and Seneviratne 2011; Seneviratne et al. 2013; Hirschi et al. 2014; Whan et al. 2015), on the formation of precipitation (Koster et al. 2004; Taylor et al. 2012; Guillod et al. 2015), and on the carbon cycle (Ahlström et al. 2015). In particular, soil moisture can have a substantial impact on the formation and severity of droughts and heat waves (Fischer et al. 2007; Zampieri et al. 2009; Hirschi et al. 2011; Mueller and Seneviratne 2012; Miralles et al. 2014; Hauser et al. 2016) and is also an important contribution to projected changes in temperature variability and extremes in the twenty-first century (Seneviratne et al. 2006, 2013, 2016; Douville et al. 2016; Lorenz et al. 2016).

The importance of soil moisture for atmospheric conditions arises from the control it can exert on water and energy fluxes at the land surface, which alters the
atmosphere’s water and energy content. Changes in near-surface air temperature can be a consequence. A conceptual overview of the coupling between soil moisture and near-surface air temperature is depicted in Fig. 1.

Soil moisture does not affect near-surface air temperature directly but rather through its control over latent heat flux. If, for example, soil moisture decreases, less water is available for evapotranspiration/latent heat flux, which results in a positive coupling of latent heat flux with soil moisture. This decline in latent heat flux causes an enhancement of the fraction of net radiation that goes into sensible heat flux, which means that latent and sensible heat fluxes are negatively coupled. Consequently, more heat is transported to the atmosphere, eventually increasing near-surface air temperature (positive coupling between sensible heat flux and temperature). Combining the single relations, this results in an overall negative coupling between soil moisture and near-surface air temperature: a reduction in soil moisture induces an increase in near-surface air temperature.

The strength of soil moisture control on the energy partitioning at the land surface depends on geographical location and can vary over the course of the year. Only in regions where soil moisture is the limiting factor for this partitioning can a dependence of water and energy fluxes on soil moisture be expected. If there is plenty of soil moisture available or if net radiation is the restricting factor, atmospheric conditions do not depend on variations in soil moisture. This fact leads to several questions: Which are the regions where soil moisture plays a major role in land–atmosphere coupling? Is the coupling strength changing with time, for example due to seasonal variations? Is it possible to quantify the coupling strength between soil moisture and near-surface air temperature?

Several studies have tried to address some of these questions using different metrics for assessing the coupling strength. Based on a large multimodel experiment, Koster et al. (2004) provided for the first time an assessment of potential regions with pronounced land–atmosphere coupling. Their study focused on soil moisture–precipitation coupling, while the follow-up analysis (Koster et al. 2006) also provided estimates of soil moisture–temperature coupling hot spots. Seneviratne et al. (2006) analyzed the impact of soil moisture on temperature variability in both present and future (late twenty-first century) conditions, highlighting a possible geographical shift of soil moisture–atmosphere coupling hot spots with changing background climate. They also proposed to use the correlation between evapotranspiration and temperature as a measure of soil moisture–temperature coupling. This metric was found to correspond well with other estimates based, for example, on the correlation between evapotranspiration and radiation (Teuling et al. 2009; Seneviratne et al. 2010). Dirmeyer (2011) combined the sensitivity of latent heat flux to soil moisture changes with the standard deviation of soil moisture in order to create an index of surface flux sensitivity to soil moisture variability and applied it to model and reanalysis data. Another metric that includes both anomalies of sensible heat flux and temperature anomalies was introduced by Miralles et al. (2012). They used it to identify regions with strong soil moisture–temperature coupling and to assess the importance of soil moisture in the development of two heat waves in Europe and the United States. Focusing more on the biosphere, Zscheischler et al. (2015) introduced the vegetation–atmosphere coupling index. It is computed from combining anomalies in temperature with anomalies in evapotranspiration (ET) or the fraction of absorbed photosynthetically active radiation (FPAR) and can be used to identify different soil moisture–plant coupling regimes that are proxies for evapotranspiration regimes.

Recently, Gallego-Elvira et al. (2016) considered the dependence of the land surface warming rate during dry spells on antecedent precipitation for several land cover types to identify different evaporative regimes.

All of the mentioned studies developed metrics for describing different aspects of the coupling between soil moisture and atmospheric conditions. Here we provide an integrated assessment of the various steps involved in soil moisture–temperature coupling. Thereby, we apply an early concept for the dependence of evaporative fraction (the fraction of net radiation used by latent heat flux) on soil moisture. The use of evaporative fraction allows us to rule out impacts of variations in net radiation and is thus well suited for investigating the control of soil moisture on the energy partitioning. A dependence of evaporative fraction on soil moisture was already
assumed in the formulation of first-generation land surface models (e.g., Manabe 1969; Sellers et al. 1997). Koster et al. (2009b) and Seneviratne et al. (2010) extended it by the introduction of a simple framework for describing the relationship between soil moisture and evaporative fraction (Fig. 2). It distinguishes between three different soil moisture regimes. If soil moisture is very low and the water is bound in the soil pore space, no evapotranspiration is taking place (dry regime). As the water content in the soil increases, evapotranspiration sets in. As the water content in the soil increases, evapotranspiration sets in. The magnitude of evaporative fraction in this transitional regime is assumed to have a first-order approximately linear dependence on soil moisture. If soil moisture exceeds a certain threshold, the relation levels off. In this wet regime, latent heat flux depends mostly on the available energy and evaporative fraction becomes independent of soil moisture.

The presented framework has two major features that allow us to assess and quantify the impact of soil moisture on the energy fluxes. The first is the sensitivity of evaporative fraction to soil moisture in the transitional regime. A high sensitivity indicates strong soil moisture control on the energy partitioning and thus a possible high impact on near-surface air temperature. The second refers to the locations of the wilting point and critical point. The former is the soil moisture value that separates the dry soil moisture regime from the transitional regime; the latter is the threshold that separates the transitional from the wet regime.

In this study, we apply the framework to two reanalysis products and a dataset combination that includes remote sensing data and gridded station observations. In a first step, we analyze geographical patterns and seasonal variations in the coupling strength between soil moisture and evaporative fraction. To that end, we identify regions that are at least temporarily located in the transitional soil moisture regime. Subsequently, the value of the critical point at each individual location as well as the sensitivity of evaporative fraction to soil moisture changes in the transitional regime can be estimated and temporal variations can be analyzed.

The dependency between soil moisture and evaporative fraction constitutes the first part of the soil moisture–temperature coupling. The second part can be analyzed by quantifying the effect of evaporative fraction on near-surface air temperature. The combination of both parts yields the complete soil moisture–temperature coupling. Only regions in which both parts have substantial coupling strength are expected to show a strong relation between soil moisture and near-surface air temperature. Eventually, the coupling strength between near-surface air temperature and soil moisture can be combined with typical soil moisture variations to estimate the average effect of soil moisture fluctuations on temperature.

2. Methods

a. Theoretical background

Soil moisture influences atmospheric conditions by its control on the magnitude of latent heat flux (Fig. 1). This alters the partitioning of net radiation $R_{net}$ into sensible heat flux $SH$, latent heat flux $LH$, and ground heat flux $G$. To study the impact of soil moisture on the energy partitioning, it is convenient to express the fluxes as fractions of net radiation. Accordingly, the energy balance equation reads as

$$\frac{LH}{R_{net}} + \frac{SH}{R_{net}} = 1 - \frac{G}{R_{net}}. \tag{1}$$

The first term in Eq. (1) is usually referred to as evaporative fraction $EF$:

$$EF = \frac{LH}{R_{net}}. \tag{2}$$

Koster et al. (2009b) and Seneviratne et al. (2010) proposed a simple framework that describes the dependence of EF on soil moisture $\theta$ (Fig. 2):

$$EF(\theta) = \begin{cases} 0, & \text{if } \theta < \theta_{\text{wilt}} \\ \frac{\theta - \theta_{\text{wilt}}}{\theta_{\text{crit}} - \theta_{\text{wilt}}}, & \text{if } \theta_{\text{wilt}} \leq \theta \leq \theta_{\text{crit}} \\ \frac{EF_{\text{max}}}{\theta_{\text{crit}}}, & \text{if } \theta > \theta_{\text{crit}} \end{cases} \tag{3}$$

The three intervals of this piecewise linear function, which are separated by the wilting point $\theta_{\text{wilt}}$ and the critical point $\theta_{\text{crit}}$, represent three different soil moisture regimes:

- The domain $\theta < \theta_{\text{wilt}}$ is referred to as the “dry regime,” since it is characterized by very low soil

![fig2.png](http://journals.ametsoc.org/jcli/article-pdf/30/18/7105/4769130/jcli-d-16-0727_1.pdf)
moisture values and no or only very small evapotranspiration (e.g., deserts).

- The interval $\theta_{\text{wilt}} < \theta < \theta_{\text{crit}}$ is called the "transitional regime." In this regime soil moisture variations have a direct impact on evaporative fraction. The framework defined by Eq. (3) assumes a linear relationship between the two variables. Regions lying in the transitional regime are characterized by strong land–atmosphere coupling (e.g., the Sahel or the Mediterranean region during summer).

- In the "wet regime" ($\theta > \theta_{\text{crit}}$) soil moisture is plentiful and no longer the limiting factor for evapotranspiration, which is under those conditions mostly controlled by the available energy. Thus, although variations of soil moisture in the wet regime can be substantial, they hardly affect evaporative fraction. Wet soil moisture regimes are commonly found in high latitudes and tropical regions.

The location of $\theta_{\text{wilt}}$ and $\theta_{\text{crit}}$ define another important parameter, namely the slope $\partial \text{EF}/\partial \theta$ in the transitional regime. Its steepness indicates the sensitivity of evaporative fraction to soil moisture and, thus, the impact of soil moisture on atmospheric conditions. To determine the complete functional form of Eq. (3), it is sufficient to estimate the coordinates of the wilting point $\theta_{\text{wilt}}$ and the critical point $\theta_{\text{crit}}$ in the $\theta$–EF space. The method used in this study focuses thus on identifying these two "break points."

b. Model selection and coupling analysis

Soil moisture variations cause shifts along the graph of Eq. (3). This relation is depicted in Fig. 3a, where EF(\(\theta\)) is shown for an exemplary grid point (13°S, 48°W) that covers both the transitional and wet regimes. Changes in soil moisture either can occur only within one regime or can lead to transitions between different regimes. Altogether, this results in five possible models for \(\text{EF}(\theta)\), which are listed in Fig. 3b:

- a constant line in the wet regime because soil moisture never reaches values lower than $\theta_{\text{crit}}$ (model 0) [in principle a constant value could also be found if soil moisture is permanently in the dry regime, but soil moisture variations in the dry regime are usually very small (Koster et al. 2009b), and the relationship essentially collapses to a point at $\theta \approx \theta_{\text{wilt}}$];
- a sloped line in the transitional regime (model 1);
- a constant line followed by a sloped line if a transition across $\theta_{\text{wilt}}$ occurs (model 2a);
- a sloped line followed by a constant line if a transition across $\theta_{\text{crit}}$ occurs (model 2b); and
- a sloped line confined by two constant lines if a transition occurs over all three regimes (model 3).

The first part of the analysis focuses on identifying the functional relationship between soil moisture and evaporative fraction, using daily data for soil moisture, net radiation, and latent and sensible heat fluxes. Since evaporative fraction is only well defined if all variables involved in Eq. (1) (i.e., latent heat flux, sensible heat flux, and net radiation) are positive, values where any of these three variables is smaller than zero are masked. Additionally, $0 \leq \text{EF} \leq 1$ is required.

Absolute soil moisture values can vary strongly between different datasets (both models and observations) while relative soil moisture values tend to be more consistent (Koster et al. 2009a; Mittelbach and Seneviratne 2012). To enable a better comparison between the different datasets used, the soil moisture data are standardized (i.e., the transformed data have an
expected value of zero and a variance of one). The transformation is based on all daily soil moisture data of the investigated time period (i.e., 30 yr for the reanalysis datasets and 11 yr for the combined dataset; see section 2c for more details) on each grid point individually for each dataset separately. All results presented in this study are based on standardized soil moisture values, except for the critical soil moisture values, which are expressed as absolute values.

After applying these modifications, the different \( EF(\theta) \) relationships are derived from the data. The constant and sloped line are retrieved by ordinary least squares fitting, while for the curves with kinks (all curves where a transition between regimes happens, i.e., models 2a, 2b, and 3) a segmented linear regression procedure, the so-called broken stick regression method, is used (e.g., Muggeo 2003; Toms and Lesperance 2003; Hofrichter 2007). This method fits a model with a predefined number of changepoints to the data by minimizing the residual sum of squares (RSS). For the purpose of this study, the method is adjusted to only produce \( EF(\theta) \) relationships according to the models 2a, 2b, and 3. To select between the models 2a and 2b, which have the same complexity level, the convexity of \( EF(\theta) \) is determined by a parabolic fit and only the model with the same convexity as \( EF(\theta) \) is used.

The model selection is based on tenfold cross validation (Hastie et al. 2001), which gives 10 RSS values for each single model. The models are ordered by increasing complexity (constant line, sloped line, one changing point, two changing points), and mean RSS and the standard deviations are calculated for each model. Finally, the simplest model with significantly lowest RSS (i.e., the RSS values do not agree within their standard deviation) is selected (see Fig. 3c). The selected model determines the soil moisture regime and yields \( \theta_{wilt} \) and/or \( \theta_{crit} \) as well as the slope \( \partial EF/\partial \theta \) in the transitional regime.

The time scale for transitions between different soil moisture regimes usually is on the order of months. To obtain robust results, the described method is applied to 3-month subsets of data, successively for all months of the year (January–March, February–April, ..., December–February; i.e., 12 subsets per grid point) to identify seasonal changes in the coupling strength between soil moisture and evaporative fraction. If in a subset more than 75% of the data are masked due to the requirements for the calculation of \( EF \) (see above), it is not used for the calculation. This guarantees that no spurious results are obtained by the use of negative heat fluxes and/or net radiation. Because of this selection criterion the number of 3-month subsets per grid point might be reduced. The detection of \( \theta_{wilt} \) and/or \( \theta_{crit} \) is either based on the break points or obtained by intersecting constant and sloped lines from three adjoining 3-month subsets and taking the average.

Changing evaporative fraction caused by soil moisture variations impacts the partitioning of net radiation [Eq. (1)]. A decrease in \( EF \), for example, leads to higher shares of ground and sensible heat fluxes. Since for daily averages (as used here) sensible heat flux is in general much higher than ground heat flux, changes in \( EF \) mostly affect sensible heat flux. As a consequence, in the transitional soil moisture regime one can expect to find an impact on air temperature (see Fig. 1). The sensitivity of near-surface air temperature \( T \) to soil moisture can be split into two contributions:

\[
\frac{\partial T}{\partial \theta} = \frac{\partial T}{\partial EF} \frac{\partial EF}{\partial \theta}.
\]

The second term on the right-hand side, the slope \( \partial EF/\partial \theta \) in the transitional soil moisture regime, represents the coupling between soil moisture and evaporative fraction and is directly obtained from the fitted models. The stronger the relation between soil moisture and evaporative fraction is, the stronger the effect on temperature should be. The first term on the right-hand side, the sensitivity \( \partial T/\partial EF \), describes to what extent changes in energy partitioning influence air temperature. Since air temperature has a pronounced seasonal cycle in many regions, temperature anomalies are used for calculating this sensitivity. Only regions where both sensitivities \( \partial T/\partial EF \) and \( \partial EF/\partial \theta \) are high can be expected to display a detectable influence of soil moisture on air temperature.

Eventually, it is possible to estimate how strongly soil moisture variations influence near-surface air temperature. One can either consider the effect of typical soil moisture variations

\[
\Delta T_{\sigma} = \left| \frac{\partial T}{\partial \theta} \right| \sigma_{\theta},
\]

where \( \sigma_{\theta} \) is the standard deviation of soil moisture in the transitional regime in the respective 3-month subset, or the maximal effect soil moisture has on temperature

\[
\Delta T_{\text{max}} = \left| \frac{\partial T}{\partial \theta} \right| (\theta_{\text{max}} - \theta_{\text{min}}),
\]

where \( \theta_{\text{max}} \) (\( \theta_{\text{min}} \)) is the maximal (minimal) soil moisture value in the transitional regime in the respective 3-month subset. Whereas \( \Delta T_{\sigma} \) is a measure for the average effect of soil moisture variations on near-surface temperature, \( \Delta T_{\text{max}} \) is an upper bound for the impact that soil moisture can have on temperature.
c. Data

Global observation-based data of soil moisture and latent heat flux are scarce and only available for a limited number of years. One possible way to overcome these limitations is the use of reanalysis- and model-based data. On the other hand, soil moisture values produced by models show large differences between different datasets (Koster et al. 2009a) and should thus be used carefully. To obtain robust results and to be able to assess the extent of agreement between the various datasets, three different data sources (two reanalysis products and one dataset combination that includes remote sensing data and gridded station observations) are used in this study (Table 1). The two reanalysis products are ERA-Interim combined with the ECMWF reanalysis of land surface parameters (ERA-Interim/Land; Dee et al. 2011; Balsamo et al. 2015), hereinafter referred to as ERA, and Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2; Bosilovich et al. 2015). The dataset combination (hereinafter referred to as COMB) uses the Global Land Evaporation Amsterdam Model (GLEAM_v3.0b; Miralles et al. 2011; Martens et al. 2017) for latent heat flux and soil moisture, CERES for net radiation (Wielicki et al. 1996), and Berkeley Earth surface temperature for daily maximum 2-m air temperature (Rohde et al. 2013a,b). Soil moisture in GLEAM_v3.0b is calculated by combining satellite measurements of soil moisture in the top layer with a simple empirical drainage algorithm to estimate the water content in the complete root zone. Latent heat flux is calculated using an equation that is based on the Priestley–Taylor method. We want to highlight here that the GLEAM_v3.0b data, while relying on various observations, do not include direct measurements of root zone soil moisture or latent heat flux, the quantities of relevance to this study. These quantities in GLEAM_v3.0b are derived and thus, like their counterparts in the reanalysis data, limited by various modeling assumptions. For the two reanalysis products ERA and MERRA-2 the investigated time period is 1980–2009 and for the COMB it is 2003–13 (restricted by the availability of GLEAM_v3.0b and Berkeley Earth surface temperature). Moreover, GLEAM_v3.0b covers only a latitudinal range from 50°F to 50°N, since it is based on satellite data solely.

Different components of evapotranspiration are influenced by soil moisture from different soil layers (bare soil evaporation mostly by the surface layer, transpiration also by deeper layers due to plant root extraction, e.g., by trees). Since the interest of this study lies in the overall effect of soil moisture on total evapotranspiration (including all ET components, i.e., bare soil evaporation, transpiration, and interception evaporation), the analyses are performed with soil moisture down to a depth of about 1 m (i.e., the root zone soil moisture): For ERA-Interim/Land the soil moisture content of the single soil layers between 0 and 100 cm is summed up (weighted by the thickness of each layer), whereas for MERRA-2 and GLEAM_v3.0b the root zone soil moisture provided in the datasets is used (see Table 1). We only briefly discuss if the results show significant differences when using surface soil moisture.

3. Results

a. The EF(θ) framework

1) OCCURRENCE OF SOIL MOISTURE REGIMES

A first assessment of the importance of soil moisture for land–atmosphere coupling is possible by identifying the different soil moisture regimes and their spatial distribution and temporal shares (Fig. 4). The requirements for meaningful values of EF (positive energy fluxes and $0 \leq \text{EF} \leq 1$; see section 2b) lead to a latitudinal gradient in the number of considered 3-month subsets (Figs. 4a-c). Whereas near the equator all 12 three-month subsets have at least 25% valid data, the coverage decreases toward the poles. This decline is mainly caused by the occurrence of negative energy fluxes during winter months, when the incident solar radiation is very small. The results
are not sensitive to variations of the threshold for valid data between 10% and 50%.

The shares of the different soil moisture regimes reveal distinct geographical patterns (Figs. 4d–o). In the maps only regimes with significant shares are displayed (the share of all regimes for ERA, MERRA-2, and COMB can be found in Figs. S1–S3 in the supplemental material). The gray coloring indicates regions for which either no data are available (e.g., COMB for latitudes higher than 50°; see section 2c) or the data are excluded from the analysis (Greenland and Antarctica because they are mostly covered with snow and ice). Many regions

FIG. 4. Number of considered 3-month subsets and share of the different soil moisture regimes for (left) ERA, (center) MERRA-2, and (right) COMB. (a)–(c) Number of considered 3-month subsets that have at least 25% valid data (the number of subsets decreases in higher latitudes due to the selection criterion explained in section 2b), and share of (d)–(f) transitional regime, (g)–(i) passage between transitional and wet regime, (j)–(l) wet regime, and (m)–(o) dry regime. The number in parentheses in the regime labels indicates the respective model as defined in section 2b. The different maps show only soil moisture regimes with significant share (the maps for all regimes can be found in Figs. S1–S3). Regions with no data are colored in light gray.
enter the transitional regime within the course of the year (Figs. 4d–f); 33% of the latitudinal-corrected global land area (without Greenland and Antarctica) lies in the transitional regime during at least half of the year in ERA (46% in COMB and 60% in MERRA-2). The passage between the transitional and wet regimes (Figs. 4g–i) can only be observed in some limited regions. Moreover, its temporal share is rather small, showing that the transition between different regimes happens usually rather fast. The wet regime (Figs. 4j–l) is mainly found in high latitudes and in the tropics (rain forests) where water is abundant and thus available energy controls evapotranspiration. The dry regime (defined here as regions where EF is on average smaller than 0.2) prevails mostly in desert regions (e.g., the Sahara, the Arabian Peninsula, and the Gobi Desert) in which water is very scarce and soil moisture nearly constant, so that evapotranspiration varies only little from zero (Figs. 4m–o).

The geographical distribution of the single soil moisture regimes for four different seasons is shown in Fig. 5. In general, the extent of the different soil moisture regimes is similar in the different datasets (wet regime in high latitudes, dry regime in deserts, and transitional regimes in between), although the respective shares of the single regimes vary between the datasets (e.g., almost no dry regime in COMB). Whereas in winter the transitional regime is mostly confined to Latin America, Africa, Southeast Asia, and Australia, it extends much farther north on both the American and Eurasian continents during summer. However, also in lower latitudes some regions switch regimes in the course of the year, such as a
For the Sahel (SAH) and East Africa (EAF) regions (identified with asterisks in Fig. 6) that were adapted to better cover areas with strong land–atmosphere interactions. They are selected based on the results of this study and the results of Miralles et al. (2012).

In general, both ERA and MERRA-2 show higher values and much higher spatial variations than COMB, especially in the Mediterranean region, the Sahel, and India. In contrast, the values agree well in United States–Central America, Northeast Brazil, East Africa, and southern Africa.

In high latitudes, both ERA and MERRA-2 reveal some regions with (slightly) negative values of $\partial\text{EF}/\partial\theta$. For these regions the term transitional regime is inappropriate, since the negative relationship indicates atmospheric control on evapotranspiration and soil moisture (e.g., Seneviratne et al. 2006, 2010; Zscheischler et al. 2015). The temporal evolution of the spatial distributions of $\partial\text{EF}/\partial\theta$ over all 12 three-month subsets is shown in animation 1 in the supplemental material together with the spatial distributions of the different soil moisture regimes.

3) CRITICAL SOIL MOISTURE VALUE

Soil moisture only has an effect on evaporative fraction in the transitional regime (Fig. 2). The critical point $\theta_{\text{crit}}$ that separates the transitional and wet soil moisture regimes is thus an important boundary in the framework presented in Eq. (3). The global distribution of average $\theta_{\text{crit}}$ is shown in Figs. 7a–c as absolute values (i.e., not standardized). The maps show clear geographical patterns that appear rather independent from the spatial patterns of $\partial\text{EF}/\partial\theta$ (Fig. 6), indicating that $\theta_{\text{crit}}$ is mostly influenced by soil properties and vegetation. Note that $\theta_{\text{crit}}$ is highest around the equator, especially in the northern part of South America, central Africa, and Southeast Asia. Low values are found in the western part of the United States, the Sahara, central Asia, and Australia.

Although the critical soil moisture values reflect model-dependent quantities, the three different datasets agree well in the respective values for $\theta_{\text{crit}}$ and show comparable spatial variations (box-and-whisker plots on the right side of Fig. 7). On average, MERRA-2 shows slightly lower values than the other two datasets, while ERA shows the highest spatial variations.

The results for $\theta_{\text{crit}}$ depend to some extent on the considered dataset. The two reanalysis products and GLEAM-v3.0b all use an explicit dependence of soil moisture for calculating evapotranspiration, which also includes the definition of a critical soil moisture value. Comparisons between the results presented here and
the $\theta_{\text{crit}}$ values used in the single datasets are discussed in section 4c.

b. Effect on near-surface air temperature

1) Slope $\partial T/\partial EF$

The coupling between evaporative fraction and near-surface air temperature constitutes the second part of the soil moisture–temperature coupling chain [Eq. (4)]. Since latent heat flux and near-surface air temperature are anticorrelated (Fig. 1), the sensitivity $\partial T/\partial EF$ is predominantly negative, as can be seen in Fig. 8. The maps show $\partial T/\partial EF$ of the 3-month subset in which each grid point reaches the maximum slope $\partial EF/\partial u$ (consistent with the results displayed in Fig. 6). Areas with a pronounced influence of EF variations on near-surface air temperature comprise large parts of the United States and Mexico,
South America, Africa, Australia, central Asia, and central and eastern Europe. Whereas both reanalysis datasets agree in this finding, COMB shows very low sensitivities in South America and Africa, which is confirmed also by the box-and-whisker plots on the right side. The three datasets agree in their $\frac{\partial T}{\partial EF}$ estimates in United States–Central America (UCA), India (IND), and northern Australia (NAU) regions, whereas in the Northeast Brazil (NEB), SAH, and EAF regions the values for COMB are much lower than for ERA and MERRA-2.

Although most of the regions in low latitudes exhibit a negative relation between evaporative fraction and near-surface air temperature, in some regions (especially in higher latitudes) the slope $\frac{\partial T}{\partial EF}$ is positive (Figs. 8a–c). The sign is an indicator for the coupling direction: negative slopes indicate that surface fluxes control atmospheric conditions (as expected in the transitional regime) and positive slopes occur when atmospheric conditions control the surface fluxes (Seneviratne et al. 2006, 2010; Zscheischler et al. 2015). Indeed, our results reveal some regions with atmospheric control on evapotranspiration and soil moisture in high latitudes, where $\frac{\partial EF}{\partial \theta}$ is (slightly) negative (Figs. 6a,b) and $\frac{\partial T}{\partial EF}$ shows positive values.
2) SLOPE $\partial T/\partial \theta$

The combination of $\partial T/\partial \theta$ and $\partial \theta/\partial t$ according to Eq. (4) yields the overall effect soil moisture has on near-surface air temperature (Fig. 9). Again, the maps show the values of the 3-month subset in which each grid point reaches the maximum slope $\partial \theta/\partial t$ (cf. Fig. 6). The black stippling in the maps indicates regions where the slope $\partial T/\partial \theta$ is not significant at the 5% level. The $p$ values where adjusted by controlling the false discovery rate (Wilks 2016). Note that the patterns are similar if the yearly average of the 3-month subsets is used (Fig. S6 in the supplemental material).

![Maps showing the effect of soil moisture on near-surface air temperature](image)

FIG. 8. As in Fig. 6, but for $\partial T/\partial \theta$ in the transitional regime. The maps show $\partial T/\partial \theta$ of the 3-month subset in which each grid point reaches the maximum slope $\partial \theta/\partial t$ (cf. Fig. 6). The black stippling in the maps indicates regions where the slope $\partial T/\partial \theta$ is not significant at the 5% level. The $p$ values where adjusted by controlling the false discovery rate (Wilks 2016). Note that the patterns are similar if the yearly average of the 3-month subsets is used (Fig. S6 in the supplemental material).
States and around the Caspian and Aral Seas, originating from higher values of both $\partial T/\partial u$ and $\partial EF/\partial u$ compared to ERA and MERRA-2 in those regions.

Consequently, the box-and-whisker plots on the right show a similar behavior as the ones in Fig. 8. The values of the different datasets agree rather well in the UCA, Mediterranean (MED), and NAU regions. In the other regions, ERA and MERRA-2 have in general much higher variations than COMB, especially in the NEB, SAH, EAF, southern Africa (SAF), and IND regions. Moreover, the very high negative values of ERA and MERRA-2 in SAH, SAF, and IND are remarkable. They reflect the strong relationships between both $\partial T/\partial u$ and $\partial EF/\partial u$ in these regions. The global patterns of $\partial T/\partial u$ are in general similar to the patterns of $\partial EF/\partial u$. The reason for this is that regions with high $\partial EF/\partial u$ are rather confined, whereas $\partial T/\partial EF$ shows high values in much larger regions.

The slope $\partial T/\partial u$ in the wet regime calculated directly as sensitivity of temperature to soil moisture is shown in Fig. S8 of the supplemental material [Eq. (4) cannot be used, since $\partial EF/\partial u$ is zero by definition in the wet regime]. It underlines that the effect of soil moisture on
temperature is as expected much lower in the wet regime. However, the negative relation between temperature and soil moisture still holds in many parts of the globe, confirming once again the conceptual schematic of Fig. 1.

3) TEMPERATURE VARIABILITY

The combination of soil moisture variability in the transitional regime and the sensitivity of near-surface air temperature to soil moisture variations enables an estimation of the average effect that soil moisture has on temperature quantified by $\Delta T_s$ (Fig. 10, showing the dependence in the 3-month subset of maximal $\partial T_s / \partial \theta$).

The maps on the left side show clear patterns of regions with high impact of soil moisture variations on temperature. The regions agree with those, where a strong effect of soil moisture on temperature is found (Fig. 9), but the extent is larger. The strongest impacts of soil moisture variations on temperature (expressed as the 95th percentile) range between 1.1 (in COMB) and 1.3 K (in MERRA-2). All three datasets show a strong impact on temperature in the United States–Central America, Australia, India, and the Mediterranean regions. Again, COMB has much lower values in South America and Africa, which is caused by the low values of $\partial T_s / \partial \theta$ in those regions (see Fig. 9). In contrast to the other two
datasets, MERRA-2 shows a strong impact of soil moisture on air temperature also in tropical regions in South America and Africa, and partly in Southeast Asia. ERA displays high values in southern Africa.

Additionally to the average effect of soil moisture on temperature anomalies, it is possible to estimate the maximal effect soil moisture can have on temperature anomalies $\Delta T_{\text{max}}$, that is, if soil moisture drops from its maximal value in the transitional regime to its minimal value (calculated for each of the single 3-month subsets individually). The associated patterns displayed in Fig. 11 differ only marginally from those in Fig. 10, but the impact of this maximal decrease of soil moisture can affect temperature by up to 6–7 K (95th percentile). One has to keep in mind though that these are climatological values that reflect the effect of soil moisture declines happening over time scales of months (and not day-to-day variations).

It should be noted that the standardization of soil moisture does not influence (or only slightly influences) the results for temperature variability, since the normalization (by the standard deviation of soil moisture) compensates when using Eqs. (5) or (6). For this reason, the temperature variability caused by variations of soil moisture provides a robust relationship that can be used to directly compare the different datasets. This is
reflected in the fact that the magnitude of $\Delta T_s$ (and $\Delta T_{\text{max}}$) is similar in all three datasets.

4. Discussion

a. ET partitioning

The calculation of evapotranspiration in the various datasets comprises three different components: bare soil evaporation, transpiration by plants, and interception (i.e., evaporation of canopy-intercepted water). The share of the single components has an influence on the overall dependence between soil moisture and evaporative fraction in the different datasets used in this study. The latitudinal variations of ET and its components in ERA-Interim/Land (E. Dutra, ECMWF, 2016, personal communication), MERRA-2, and GLEAM_v3.0b are shown in Fig. 12. The three different datasets agree in their ET variability with ET being highest in tropical regions and decreasing toward the poles. Overall, transpiration and bare soil evaporation are the biggest shares of evapotranspiration (although in ERA-Interim/Land interception also has a relatively high share). While bare soil evaporation does not change much with latitude, transpiration and interception are highest in zones with dense vegetation cover, such as tropical rain forests, with a strong decrease toward the poles.

In contrast, the shares of the single ET components vary substantially between the different datasets. While in GLEAM_v3.0b and ERA-Interim/Land transpiration has the biggest share, in MERRA-2 bare soil evaporation is dominating. Especially around the equator, MERRA-2 has a rather high share of bare soil evaporation compared to the other two datasets. MERRA-2 uses a somewhat different hydrological model formulation than the other two products, namely a tile formulation with three different subtiles per tile (R. Koster, NASA GSFC, 2016, personal communication): 1) a saturated fraction, in which ET is energy limited, 2) a transitional fraction where ET depends on soil moisture, and 3) a wilting fraction with no evapotranspiration. Soil moisture variations affect both evapotranspiration in the transitional regime as well as the shares of the three subtiles in each tile. Furthermore, soil moisture variations influence bare soil evaporation much more strongly than transpiration in the transitional regime (not shown). This fact, combined with the observation that MERRA-2 has a relatively high fraction of bare soil evaporation, could explain the high share of transitional regimes (plus passage transitional/wet) in MERRA-2 (Fig. 4).

The overall agreement of the latitudinal ET variations (and also the variations of the single ET components) in the different datasets provides confidence that the global patterns of soil moisture–temperature coupling are robust. However, the absolute values of the coupling should be interpreted with more caution, since they depend partly on the share of the ET components, which is not completely consistent in the three different datasets.

b. Assessment of model fit

The framework used in this study for obtaining the functional relation $EF(\theta)$ is based on a limited number of models and resulting relationships. It is thus not able to capture all facets of the function $EF(\theta)$, but it constitutes a helpful approach for assessing its key features. To estimate both the power as well as the limitation of the used framework, it is important to analyze the quality of the model fits. For this purpose, the residuals of the models can be used (Fig. 13). The maps show the
mean standard deviation (averaged over the year) of the residuals of the model that best describes the EF(θ) relationship in each 3-month subset (as selected by tenfold cross validation; see section 2b and compare with Fig. 5). Low standard deviations of the residuals in regions with a strong coupling between soil moisture and evaporative fraction indicate that the framework can explain the variations of EF and that the obtained regime classification as well as the estimates for ∂EF/∂θ and θ_{crit} are reliable.

The maps reveal a clear latitudinal dependence of the residuals. The highest residuals are found in high latitudes, indicating that there the framework does not capture the driving mechanism for evaporative fraction. Indeed, energy availability and not soil moisture is the limiting factor for evapotranspiration in high latitudes. In lower latitudes the framework seems to give more robust results, but the single datasets reveal some particular differences in their residual patterns.

ERA has particularly high residuals in Australia, the western United States, and parts of Africa and central Asia. These regions coincide with areas where ERA also shows strong EF variations over the investigated time period (Fig. S11 in the supplemental material). The combined occurrence of both high residuals and high EF variations indicates that in these regions either the latter is caused by other factors or the applied linear framework is not adequate to explain the EF variations in these regions. Moreover, the slopes ∂EF/∂θ are in general rather low in these regions (see Fig. 6), pointing out that the effect of soil moisture on evaporative fraction is not captured accurately. [Note, however, that the EF(θ) framework captures the EF variations well if only the surface soil moisture is used (not shown). This indicates that in ERA the evaporative fraction in these regions depends mostly on the soil moisture in the upper soil layers, whereas the lower layers are rather decoupled with regard to evapotranspiration.]

MERRA-2 does not show such distinct regions with high residuals. It has rather small residuals in lower latitudes, indicating that the applied framework explains the relation between soil moisture and evaporative fraction satisfactorily. Around the Mediterranean region and in the United States, which are both regions that enter the transitional regime during a substantial part of the year (Figs. 4e,h) the residuals are, however, rather high. Similar conclusions apply to COMB, except that it has smaller residuals than MERRA-2 in the United States. Both ERA and COMB have slightly higher residuals than MERRA-2 in densely vegetated areas (e.g., rain forests). Again, this can be expected since evapotranspiration in these wet regions is not dependent on soil moisture availability.

In general, regions in which the slope ∂EF/∂θ is high have rather low residuals. This indicates that in the transitional regime, EF variations are indeed mostly explained by variations of soil moisture.

c. Comparison of critical soil moisture values

ERA-Interim/Land, GLEAM_v3.0b, and MERRA-2 all include an explicit dependence of evapotranspiration on soil moisture in their models (ECMWF 2012; Martens et al. 2017; R. Koster, NASA GSFC, 2016, personal communication). The calculation includes the definition of a critical point similar to the one that we obtain as a result of applying the framework described
in Eq. (3). To investigate the extent to which the results of this study are influenced by the formulations in ERA-Interim/Land, MERRA-2, and GLEAM_v3.0b, the $\theta_{\text{crit}}$ values implemented in those models are compared to the findings of our analysis (Fig. 14). Both ERA-Interim/Land and GLEAM_v3.0b show similar $\theta_{\text{crit}}$ values in large parts of the world (Fig. 14, top), but GLEAM_v3.0b has lower values in the Sahara and central Asia. MERRA-2 has on average lower $\theta_{\text{crit}}$ values, but the geographical variations agree well with the other two datasets. Accordingly, the differences between the critical values calculated in this study and critical points implemented in the models are larger for ERA than for MERRA-2 and COMB (Fig. 14, bottom). The $\theta_{\text{crit}}$ values used in the models seem to provide an upper boundary for the values calculated in this study. Only in tropical regions the calculated values are slightly higher than the ones implemented in the models. The rather high differences between calculated and implemented critical points indicate that the results of this study are not merely a consequence of the model formulations, but provide an effective estimate of the influence of soil moisture on evaporative fraction.

5. Conclusions

Soil moisture plays an important role for climate and atmospheric conditions in multiple regions of Earth. However, the geographical extent and the strength of the effect vary throughout the year. The EF($\theta$) framework used in this study takes both the spatial and temporal variations of the coupling between soil moisture and evaporative fraction into account. It is especially useful to investigate the coupling strength of the two variables and to identify the critical soil moisture value that constitutes the boundary between the wet and transitional soil moisture regimes.

The analysis is based on three different data sources (two reanalysis datasets and one dataset combination that uses remote sensing data and gridded station observations to derive estimates of root zone soil moisture and latent heat flux). The results reveal that many regions, mostly located in low latitudes, enter the transitional regime in the course of the year and about 30%–60% (depending on the dataset) of the latitudinal-corrected global land area (without Greenland and Antarctica) lies in the transitional regime during at least half of the year (Fig. 4). On the other hand, the impact of soil moisture on evaporative fraction, quantified by the slope $\partial \text{EF}/\partial \theta$ varies notably between different regions (Fig. 6). Moreover, the coupling strength also changes over the course of the year (see the animations in the supplemental material). Overall, the regions with strong land–atmosphere coupling identified here agree with the results of other authors (Koster et al. 2004; Seneviratne et al. 2006; Miralles et al. 2012).

Although the three different data sources generally agree in their geographical patterns for the several relationships analyzed in this study, the absolute strengths of the relationships show partially large differences. Since none of the used data sources can be seen as a reference, the extent to which the datasets agree (or
disagree) provides an indication for the reliability of the results and of the presented relationships. Larger spread between the different datasets indicates that the processes involved in land–atmosphere coupling in those regions are still uncertain and require more research.

The results presented in this study are obtained using soil moisture down to about 1-m depth (i.e., the root zone layer), but they also stay qualitatively the same when using only surface soil moisture (not shown). This indicates that the soil moisture impact on evaporative fraction is rather insensitive to the soil moisture layer used [at least for mean climatological conditions; for very dry conditions Hirschi et al. (2014) found contrasting results], an outcome that is in line with the findings of Qiu et al. (2016).

Another important result of the analysis is the identification of the critical soil moisture value that constitutes the boundary between the transitional and wet soil moisture regimes (Fig. 7). Although the models analyzed in this study use implemented critical soil moisture values, the critical values resulting from our analysis differ from the implemented model values. This indicates that our analysis is resulting in emerging, effective critical soil moisture values, while the implemented model values can be seen as an upper boundary.

The effect of soil moisture on near-surface air temperature, which results jointly from 1) the effect of soil moisture on evaporative fraction (Fig. 6) and 2) the effect of evaporative fraction on temperature anomalies (Fig. 8), is particularly high in certain confined regions in lower latitudes (Fig. 9). In high latitudes, where evapotranspiration is mainly energy limited, soil moisture does not have an effect on air temperatures (as expected from previous analyses; e.g., Seneviratne et al. 2006, 2010).

Combining the sensitivity of near-surface air temperature on soil moisture with typical soil moisture variations allows the estimation of the average effect soil moisture has on near-surface air temperature. It impacts temperature in multiple regions of the world (Fig. 10) by up to 1.1–1.3 K (95th percentile, depending on the dataset).

The different results of this study highlight that climatic conditions in various regions are significantly impacted by varying soil moisture conditions. The presented analyses may be applied in the future for the investigation of soil moisture–temperature coupling in climate models.

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