Soil moisture-temperature coupling: A multiscale observational analysis

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Received 28 August 2012; revised 3 October 2012; accepted 4 October 2012; published 6 November 2012.

[1] Land-atmospheric interactions are complex and variable in space and time. On average soil moisture-temperature coupling is expected to be stronger in transition zones between wet and dry climates. During heatwaves anomalously high coupling may be found in areas of soil moisture deficit and high atmospheric demand of water. Here a new approach is applied to satellite and *in situ* observations towards the characterization of regions of intense soil moisture-temperature coupling, both in terms of climatology and anomalies during heatwayes. The resulting average summertime coupling hot spots reflect intermediate climatic regions in agreement with previous studies. Results at heatwave-scale suggest a minor role of soil moisture deficit during the heatwave of 2006 in California but an important one in the 2003 event in Western Europe. Progress towards near-real time satellite products may allow the application of the approach to aid prediction and management of warm extremes. Citation: Miralles, D. G., M. J. van den Berg, A. J. Teuling, and R. A. M. de Jeu (2012), Soil moisture-temperature coupling: A multiscale observational analysis, Geophys. Res. Lett., 39, L21707, doi:10.1029/2012GL053703.

1. Introduction

[2] Since the unprecedented 2003 event in Europe heatwaves have drawn extensive attention from science and media. Thereafter our understanding of the physical processes behind heatwaves has improved [e.g., Schär et al., 2004; Ciais et al., 2005; Teuling et al., 2010a]. The presence of anticyclonic atmospheric conditions - with a clear sky, warm advection and high soil temperatures - is commonly recognized as a requirement of occurrence [Meehl and Tebaldi, 2004]. Stable atmospheric conditions may favor control by the local energy balance and, consequently, the surface sensible heat flux becomes the driver of air temperature. These conditions enable a positive feedback from land: as soil moisture deficit increases as a consequence of the high atmospheric demand of water, evaporative cooling decreases leading to a further raise in air temperature. The effect of soil moisture on air temperature is

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commonly referred as soil moisture-temperature coupling [Seneviratne et al., 2010].

[3] Nowadays it is accepted that droughts increase the probability of occurrence of extreme warm events [Hirschi et al., 2011]. In regions like Europe or North America, where climate change is expected to increase the frequency and intensity of heatwaves [Meehl and Tebaldi, 2004; Alexander et al., 2006], several studies have pointed to the role of soil-atmosphere interactions to help explain projected changes in summer climate [Seneviratne et al., 2006; Vidale et al., 2007]. Acknowledging that the level of coupling depends on climatic conditions, modeling activities have focused on the portrayal of the world's regions of intense coupling [Koster et al., 2006] and how these regions may shift as climate changes [Seneviratne et al., 2006]. The GLACE experiments exposed the land-atmosphere coupling hot spots at the global scale [Koster et al., 2006], and different activities have recently contributed to the better understanding of coupling at the regional [e.g., Taylor et al., 2011] and local scales [e.g., Santanello et al., 2011].

[4] Soil moisture-temperature coupling experiments looking at the short-scale of heatwaves have mainly focused on climate modeling activities [*Fischer et al.*, 2007] and statistical analysis of meteorological data [*Teuling et al.*, 2010a; *Hirschi et al.*, 2011]. Traditional coupling diagnostics used in climate modeling experiments are unambiguous in terms of causality. However, the degree of land-atmospheric coupling varies greatly from model to model and, additionally, results cannot be replicated or validated using field measurements [*Seneviratne et al.*, 2010]. On the other hand, ground measurements are accurate, but even when appropriate coupling diagnostics are applied, the limited spatial coverage of current meteorological networks precludes global-scale analyses.

[5] Despite being uncertain, satellite information allows the study of soil moisture-temperature coupling from an observational perspective and at the global scale. The recent rise of a series of remote sensing-based global land surface products, including soil moisture [e.g., *Owe et al.*, 2008] and evaporation [e.g., *Fisher et al.*, 2008; *Zhang et al.*, 2010; *Miralles et al.*, 2011a], enables progress in this direction. The daily temporal resolution of some of these datasets allows the study of coupling at the short time scales relevant for heatwaves, while the multi-decadal record length can be used to analyze climatological means of land-atmospheric coupling.

[6] Here we propose the first use of one of these satellitebased datasets, in combination with unique diagnostics, to study soil moisture-temperature coupling. The suggested diagnostics aim to fill the gap between climatological and event studies of soil moisture-temperature coupling by focusing on both contrasting timescales. In the following

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sections the approach is described and applied. Global summertime coupling *hot spots* are illustrated and compared to in situ-based estimates. Subsequently, we explore the role of soil moisture during the 2003 heatwave in Western Europe and the 2006 event in California.

2. Methods

2.1. Coupling Metrics

[7] The rationale is the estimation of two energy balances – one based on actual evaporation (*E*) and one based on potential evaporation (E_p) – and their differential skill in explaining the dynamics of near-surface air temperature (*T*). When long-term time series of daily surface net radiation (R_n), *E* and *T* are available, we propose the following metric (II) to capture soil moisture-temperature coupling at long (climatic) timescales

$$\Pi = \rho(H, T) - \rho(H_{\rm p}, T) \tag{1}$$

where ρ is the Pearson's correlation coefficient, $H = R_n - \lambda E$ and $H_p = R_n - \lambda E_p$. The latent heat of vaporization (λ) can be calculated as a function of *T*, and E_p can be estimated as a function of R_n and *T* [*Priestley and Taylor*, 1972]. Positive values of II are obtained for those regions where considering soil moisture restrictions in the partitioning of surface energy helps explain a larger fraction of the variability of *T*. The ground heat flux (*G*) is not included in equation (1); the effect of this omission is examined in the auxiliary material (Table S1 and Figure S2).¹

[8] Considering σ_T , σ_H and σ_{H_p} the standard deviations of *T*, *H* and *H*_p, equation (1) can also be expressed in terms of covariances (*cov*):

$$\Pi = \frac{1}{\sigma_T} \left(\frac{cov(H, T)}{\sigma_H} - \frac{cov(H_p, T)}{\sigma_{H_p}} \right)$$
(2)

Since Π is based on long-term correlations, it has to be modified for its application at the short time-scales of heatwaves, where the focus is on anomalously high temperatures. Keeping the same rationale as in equations (1) and (2), we define the daily coupling metric (π) for day *i* as

$$\pi_{i} = \frac{T_{i} - \overline{T}}{\sigma_{T}} \left(\frac{H_{i} - \overline{H}}{\sigma_{H}} - \frac{H_{p,i} - \overline{H_{p}}}{\sigma_{H_{p}}} \right)$$
(3)

where \overline{T} , \overline{H} and $\overline{H_p}$ are the averages of the long-term series of T, H and H_p . If we drop the subscript '*i*', the notation can be simplified as:

$$\pi = \left(H' - H_{\rm p}'\right)T' \tag{4}$$

where T, H' and H'_p indicate the daily anomalies of T, Hand H_p expressed in the number of standard deviations relative to the expectation. Note that π consists of an energy term $(H' - H'_p)$ and a temperature term (T'). The energy term represents the contribution of soil moisture deficit to H, i.e., the short-term potential of soil moisture to affect T. This term will be zero when soil moisture is sufficient to meet the atmospheric demand, and under dry conditions it will increase as the atmospheric demand increases. Only if this potential of soil moisture to affect T concurs with an anomalously high value of T (a large T), the local energy balance may be controlling air temperature and π will be large.

[9] It can be noted that the scale of Π and π differ, as the latter expresses how anomalous measurements on a single day are (in terms of standard deviations), while the former comprises the long-term record (in terms of correlation coefficients). Values of Π and π smaller or equal to zero denote no coupling, whereas higher values indicate higher coupling.

2.1. Data

[10] Estimates of H and H_p are derived from GLEAM (Global Land-surface Evaporation: the Amsterdam Methodology) as described by Miralles et al. [2011a, 2011b]. GLEAM calculates E by combining estimates of E_p and evaporative stress. $E_{\rm p}$ is estimated via Priestley and Taylor equation using R_n (GEWEX Surface Radiation Budget 3.0 [Stackhouse et al., 2004]) and T (Atmospheric InfraRed Sounder (AIRS) and ISCCP [Rossow and Schiffer, 1999]). The evaporative stress is based on the root-zone soil moisture calculated via data assimilation of microwave surface soil moisture [Owe et al., 2008] into a multilayer profile driven by observations of precipitation. GLEAM datasets have been extensively validated and are available at daily time steps, for 1984–2007, at 0.25° spatial resolution. In this study the interception loss of GLEAM is not considered. provided both its independency from the soil water content and the current discrepancies about the sources of energy driving this flux in nature [see Holwerda et al., 2012].

[11] For the estimates of *T* in equations (1) and (4) we use the daily average screen-level (2 m) air temperature from ERA-Interim [*Dee et al.*, 2011]. ERA-Interim is the latest reanalysis by the European Centre for Medium-Range Weather Forecasts (ECMWF). The dataset is global, spans from 1979 to near-real time and has a 1.5° resolution. ERA-Interim *T* has been chosen for its overall accuracy [see *Mooney et al.*, 2010]. Estimates have been downscaled to 0.25° by inverse distance weighted interpolation.

[12] Meteorological measurements from the Opened synthesis FLUXNET dataset [*Baldocchi et al.*, 2001] are used for the *in situ* comparison. Only stations in Europe and North America with at least 100 daily summertime observations and less than 20% mismatch in their energy closure are considered. This adds a total of 41 sites that cover a large variety of ecosystems. The list of stations is presented in Table S1.

3. Global Coupling Regions

[13] Overall land's control over air temperature is larger than its control over precipitation, but both are expected to occur in similar climatic regions [*Koster et al.*, 2006]. Here we examine the location of the global *hot spots* of soil moisture-temperature coupling by using the metric described in equation (1) and applying it to GLEAM (*H* and H_p) and ERA-Interim (*T*) data.

[14] Figure 1a shows Π for boreal summer. Only data from the months of June, July and August (JJA) during 1984–2007 are used to calculate the metric. As expected, intermediate regions between dry and wet climates are highlighted – these are regions in which soil moisture limits

¹Auxiliary materials are available in the HTML. doi:10.1029/2012GL053703.



Figure 1. Global soil moisture-temperature coupling (Π) over the period 1984–2007, as calculated using GLEAM (*H* and *H*_p) and ERA-Interim (*T*) for: (a) boreal summer (JJA), and (b) austral summer (DJF). (c) The results of the FLUXNET comparison study for JJA. The size of the markers is directly proportional to the length of the record of measurements at the station. Stations falling within adjacent pixels have been combined to aid visualization. The background in Figure 1c reflects the gridded results for JJA over period 1995–2007.

evaporation and is still sufficiently dynamic to significantly affect the variability of air temperature. Results are in agreement with the coupling *hot spots* displayed by *Koster et al.* [2006], and in even closer correspondence with the results from *Seneviratne et al.* [2006, 2010] – i.e., in the latter the Mediterranean region was portrayed as an area of strong coupling but Eastern China was not. Figure 1b illustrates II for austral summer (December, January and February, DJF). *Hot spots* concentrate in different regions than in Figure 1a but still expose transitional climate zones.

[15] Since the analysis focuses on summer periods only, the role of the seasonal cycles of H, H_p and T in the correlations calculated in equation (1) is rather unimportant. This is demonstrated by the similarities between Figures 1 and S1. In Figure S1 the seasonal expectations at each day of the year – calculated based on a 31-day moving window and the entire multiyear dataset – have been removed prior to equation (1).

[16] In order to validate the results from Figures 1a and 1b with fully independent observations, Π has also been derived using measurements of R_n , λE and T from FLUX-NET as input to equation (1). Figure 1c shows the correspondence between pixel and point estimates for North America and Europe during JJA. Pixel estimates in the background correspond to 1995–2007 (i.e., the period with FLUXNET data) using GLEAM (*H* and H_p) and ERA-Interim (*T*) data. Despite point-to-pixel errors and differences in the recording periods at each of the stations, Figure 1c shows correspondence between point and pixel inferences, with a correlation coefficient of 0.66 and a positive bias of 0.14, mainly due to the overestimation of coupling in the Southern Great Plains and Western Mediterranean.

[17] Table S1 presents the results at each of the stations, and Figure S2 recreates Figure 1c but considering the *G* measured at the stations in the calculations of *H* and H_p . Comparison between Figures 1c and S2 suggests an effect of *G* in dry regions, but a limited influence in terms of the overall spatial variability of the metric (with a correlation coefficient between II estimates with and without *G* of 0.94). Even though the results of this validation do not prove the skill of the metric to detect coupling, they indicate that the *hot spots* can be replicated using higher quality independent observations. This provides an insight on to what extent Figures 1a and 1b are affected by the choice of GLEAM and ERA-Interim as input to equation (1).

4. Coupling Anomalies During Heatwaves

[18] In the summer of 2003 in Western and Central Europe, temperatures exceeded the 1961-90 mean by up to 5 standard deviations [Schär et al., 2004]. Anomalies culminated with unprecedentedly high values from August the 3rd to August the 12th [Fischer et al., 2007; García-Herrera et al., 2010] resulting in a dramatic increase in mortality rates, especially in France. Figure 2a illustrates the average daily soil moisture-temperature coupling (π) during this period as calculated via equation (4). The long-term expectation and standard deviation of H, H_p and T in equation (4) are based on the months of JJA between 1984 and 2007. High coupling concentrated over France, despite the fact that extreme temperatures extended across the majority of Europe. The spatial variability of coupling agrees with the results by Fischer et al. [2007] using a regional climate model, even though their analyses focused on longer-term impacts of soil moisture deficits, which can also act via their potential impact on atmospheric circulation.

[19] Figure 3a illustrates the evolution of the two terms of π in equation (4), i.e., the temperature term (*T*) and the energy term ($H' - H'_p$). The left panel indicates that over France there was evaporative stress even before the heat-wave arrived – this is revealed by the anomalous levels of energy contributed by soil moisture deficit (i.e., large values of $H' - H'_p$). Despite areas like the British Isles, the Alps or the Benelux experiencing extreme anomalies in *T* during the event (shown by the orange and red colors) the effect of soil moisture deficit on the energy balance was not anomalous. It is mainly France where high values of *T* and $H' - H'_p$ concurred during the heatwave, and in addition, the only region where the *T* reached up to 4 standard deviations over the summer expectation. Figure 3b presents the time series of *T*



Figure 2. Soil moisture-temperature coupling (π) during heatwaves of (a) Europe 2003 (from August 2nd to 12th) and (b) USA 2006 (from July 16th to 25th).

and $H' - H'_p$ during July and August in the epicenter of the heatwave in France (coarsely marked by a white dash line in the second panel of Figure 3a) – grey boxes in Figure 3b highlight the periods considered in Figure 3a. During the heatwave event both T' and $H' - H'_p$ reached their maximum; the fact that H_p is not anomalous indicates that the peak in Tdoes not respond to anomalies in R_n .

[20] Not as severe and dramatic as the 2003 event in Europe, the summer 2006 event in North America still caused considerable morbidity and mortality [Gershunov et al., 2009]. By mid July, unprecedented high temperatures extended across the majority of USA and added to the drought in the Great Plains and the Southeast. The heatwave hit the West Coast around July 21st, setting new temperature records in parts of California. At the end of July, the high temperatures moved back to the Great Plains and dissipated through the east at the beginning of August [National Climatic Data Center, 2006]. Studies of this event have focused on the effect of greenhouse emissions, El Niño teleconnections and atypical high-pressure systems [e.g., Hoerling et al., 2007; Kozlowski and Edwards, 2007]. Figure 2b indicates that anomalous soil moisture-temperature coupling (i.e., $\pi > 0$) also occurred over the Midwest during the event (here considered as July 16th to 25th). Over the West Coast however, the coupling was lower – the values of $\pi \sim 1.5$ found in California during the event contrast with the π up to 8 of the 2003 event in France (see Figure 2a).

[21] Figures 3c and 3d take a closer look at the heatwave. The low H'_p during the peak of the event in California (corresponding to the third panel in Figure 3c) suggests that the maximum *T* did not respond to anomalies in R_n , and the low $H' - H'_p$ suggests that the soil moisture restriction over evaporation was not anomalously high. This supports the hypothesis that ocean advection and decreased upwelling of cooler waters played a major role in the event [*Gershunov* et al., 2009; Kozlowski and Edwards, 2007].

[22] Results in Figures 2 and 3 can be impacted by errors in the parameterizations of root-zone soil moisture and

evaporative stress in GLEAM, as well as the uncertainties of the satellite observations used to estimate E and E_p . Uncertainties in ERA-Interim T also affect the results. Some of the stations from Table S1 are located close to the epicenters of the two heatwaves and reported T and energy fluxes during the events. Figure S3 analyses the correspondence between the gridded estimates of H, H_p and T and the measurements from stations in France (FR–SRb) and California (US–Ton and US–Var) in the course of the two heatwaves. The general agreement between gridded and station-measured variables shown in Figure S3 adds confidence to the results above presented.

4. Conclusion

[23] Recently developed global satellite-based evaporation products open new opportunities for the observational analyses of land-climate interactions. Here soil moisture-temperature coupling is analyzed at different scales by combining satellite-based estimates of evaporation and temperature using new diagnostics. Resulting global coupling *hot spots* agree with the regions depicted by *Koster et al.* [2006] and *Seneviratne et al.* [2006, 2010], illustrating zones of transitional climate, both for boreal and austral summer. Comparison with *in situ*-based inferences suggests that the results are robust and not a product of the current choice of input data.

[24] Soil moisture-temperature coupling during the 2003 European heatwave and the 2006 USA heatwave has been explored. The resulting spatial patterns of coupling correspond well with experiments using regional climate models [*Fischer et al.*, 2007] and support the conclusions of previous studies about the drivers behind the two heatwaves. Maximum coupling is found over France during the 2003 heatwave, even prior to the peak in temperatures. Lower coupling is found in the 2006 event in USA; soil moisture is suggested to have played an insignificant role in California.

[25] It is worth stating that the effect of soil moisture in the local energy balance is simplified in this study. Soil moisture



Figure 3. Evolution of temperature and energy anomalies [relative to JJA from 1984–2007] for: (a and b) the 2003 event in Europe, and (c and d) the 2006 event in USA. Time series of T, H', H'_p and $H'-H'_p$ for the months of July and August are shown in Figures 3b and 3c for the 2003 and 2006 event respectively. The regions contributing to the time series are marked with white dash lines in Figures 3a and 3c. Grey boxes in Figures 3b and 3d mark the heatwave periods considered in Figures 3a and 3c respectively. Bivariate colormaps have been created using the technique described in *Teuling et al.* [2010b].

effects on albedo and ground heat flux are assumed to be negligible compared to the role of soil moisture in the partitioning of radiation between latent and sensible heat flux. In addition, other potential soil moisture feedbacks on atmospheric dynamics (e.g., enhanced cloudiness, long-term changes in atmospheric circulation), which may affect the potential and actual rates of evaporation, are not directly accounted for. Given the range of assumptions, the proposed diagnostics cannot provide ultimate proof of causality. Nevertheless, the simplicity of the approach is key for its applicability, and results suggest that it can potentially isolate the contribution of soil moisture deficit to air temperature in an efficient manner.

[26] Different lag times can also be applied to our proposed metrics to increase their potential for seasonal forecasting. Given the memory associated with soil moisture, efforts towards the better understanding of land-atmosphere interactions may help improve weather forecasts [*van den Hurk et al.*, 2012; *Mueller and Seneviratne*, 2012]. Recent progress in satellite remote sensing towards providing data products at near-real time sets an opportunity for approaches like ours to aid prediction and management of warm extremes. [27] Acknowledgments. We thank FLUXNET and in particular the PIs of the sites listed in Table S1. We also thank Wade Crow for his valuable feedback. AJT acknowledges financial support from The Netherlands Organization for Scientific Research through Veni grant 016.111.002. This work is partially supported by the European Space Agency contract no. 4000106711/12/I-NB.

[28] The Editor thanks the anonymous reviewer for his/her assistance in evaluating this paper.

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