Sensitivity of afternoon precipitation to evaporative fraction in eastern Asia based on ERA-Interim datasets

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1 INTRODUCTION

Soil moisture–precipitation feedback has long been a topic of interest. Previous studies have recognized that soil moisture could modify atmospheric processes on a range of spatial and temporal scales (Seneviratne et al., 2010; Taylor et al., 2011), potentially leading to cloud formation (Ek and Holtslag, 2004) and precipitation (Pielke, 2001; Taylor et al., 2012). Quantifying soil moisture–precipitation coupling has implications for improving our ability to forecast weather and predict climate. Under future climate scenarios, with higher values of evaporation driving increased soil moisture deficits in many places, soil moisture–precipitation feedbacks may play an increasingly important role (Dirmeyer et al., 2013; Seneviratne et al., 2013). However, because of incomplete representations of complex physical systems in models and lack of observational data around the world, the nature of soil moisture–precipitation feedback is still debated (Seneviratne et al., 2010; Berg et al., 2013; Guilgod et al., 2015; Tuttle and Salvucci, 2016; Welty and Zeng, 2018).

To facilitate a comprehensive investigation and effective quantification of complicated interactions and feedbacks between soil moisture and precipitation, a land–atmosphere coupling process chain was proposed, which is composed of the sensitivity of evaporative fraction (EF; \( \text{EF} = \lambda E / (H + \lambda E) \)) with \( H \) and \( \lambda E \) representing the surface sensible and latent heat flux, respectively, to soil moisture, the planetary boundary layer (PBL) evolution to the surface flux (Santanello Jr. et al., 2011).

The first part of the process chain may be the most significant in dry-wet climate transitional regions, where evapotranspiration is mainly determined or limited by soil moisture (Koster et al., 2004; 2006; Guo et al., 2006). The second part is how both the moisture and heat fluxes into the atmosphere impact the development of PBL and thereby the initiation and
intensity of precipitation, which is the most uncertain part in the chain (Van Heerwaarden et al., 2009; Findell et al., 2011; Santanello Jr. et al., 2018). One important reason for the controversy about the sensitivity of precipitation to evaporation is that the precipitation process is very complicated. In addition, the lack of sufficient observations is also an important obstacle so many studies are carried out with different reanalysis datasets. For examples, Findell et al. (2011) and Aires et al. (2014) found that local surface fluxes represent an important trigger for convective rainfall in eastern United States and Mexico during summer, leading to a positive evaporation–precipitation feedback using the North American Regional Reanalysis dataset (NARR) with a metric of triggering feedback strength (TFS). Guillod et al. (2014) agreed with Findell about the positive feedback over the eastern and southwestern United States using a modified metric version of triggering feedback strength (TFS*).

In the previous studies, investigations of feedback strength between evaporation and precipitation have mainly focused on analyses in the United States and particularly in the Great Plains, little information has been known in Asia, where topography and climate are diverse, and the nature of the coupling may be quite different. Based on this consideration, this paper quantifies the relationship between before-noon EF and afternoon convective precipitation across eastern Asia regions via TFS and amplification feedback strength (AFS).

2 DATA AND ANALYSIS METHOD

Compared with the observations of soil moisture and precipitation, the European Centre for Medium-Range Weather Forecasts Interim reanalysis (ERA-Interim) and ERA-Interim/LAND datasets have good applicability on the scales of daily, monthly and annual (Betts et al., 2009; Albergel et al., 2013; Balsamo et al., 2013). In addition, Sato and Xue (2013), Liu et al. (2016) and Sathyanadh et al. (2016) evaluated several sets of reanalysis data and found that the ERA reanalysis data were relatively reliable in East Asia. Therefore, the ERA-Interim and ERA-Interim/LAND are used in this study. ERA-Interim provides 4 times daily the air temperature, pressure and specific humidity in model-level at the resolution of 0.125 × 0.125° covering the summer period of 1979–2010, and ERA-Interim/LAND provides the 3-hourly integration of the sensible heat flux, latent heat flux and 6 times daily total precipitation at the 0.125 × 0.125° resolution from 1979 to 2010. For each grid, 2,944 summer-time days are available for analysis using ERA datasets (32 years, 92 days in June–July–August [JJA]).

Two metrics are used to assess the early morning weather potential for convective development: convective triggering potential (CTP) and low-level humidity deficit (HI_low) (Findell et al., 2011). CTP is a measure of energy available for convection in an area of the atmosphere 100–300 hPa above land surface, which is pressure interval likely to be critical to the development of daytime boundary layer, and HI_low is defined as the sum of dew-point depressions 50 and 150 hPa above land surface (Findell and Eltahir, 2003a; 2003b). Accounting for these two variables is expected to reduce confounding effects of atmospheric conditions. CTP and HI_low are calculated as follows (McCabe et al., 2008):

\[
CTP = \int_{100mb}^{300mb} \frac{g}{T_{\text{parcel}} - T_{\text{env}}} dz,
\]

where \( z \) is pressure level above the surface, \( g \) is gravity, \( T_{\text{ parcel}} \) is parcel temperature at 100 mb and \( T_{\text{env}} \) is environment temperature.

\[
\text{HI}_{\text{low}} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}),
\]

where \( T_{p} \) and \( T_{d,p} \) are temperature and dew-point temperature at pressure level \( p \), respectively. Since the study area spans five time zones (from the East 5th District to the East 10th District, and the local standard time [LST] equals Greenwich Mean Time [GMT] plus five to ten hours) and the reanalysis dataset are taken at a 3-hour interval, CTP and HI are calculated for each time zone using data at around 0600 LST. The LST mentioned below is the same.

EF is used to assess the energy partitioning at land surface before noon, because it has been shown to be relatively constant and almost not affected by turbulence variability in contrast to the large variation of surface sensible and latent heat flux (Cragoa and Brutsaert, 1996). The sensible and latent heat flux data from ERA-Interim/LAND at 1200 LST are used. Afternoon rainfall is defined from 1200 to 1800 LST.

To constrain our analyses to the local impact of before-noon surface heat fluxes on subsequent convective development and prevent a probable influence of synoptic systems, three preconditioning measures are taken. First, remove days with negative CTP when early morning condition is too stable to support afternoon convection. Second, remove days with rain already occurring before noon to prevent influencing from a long-duration stratiform rainfall event, and third, delete the days with rain on the day before or after to restrain the impact of precipitation persistence (Guillod et al., 2014). Rainfall is “triggered” only when afternoon rainfall exceeds a small threshold value, currently set to 1 mm. Through these restrictions, about 51% of days are excluded.

To quantify the impacts of EF on frequency and intensity of summer convective rainfall, two measures are adopted: TFS and AFS (Findell et al., 2011); they reflect respectively how afternoon rainfall frequency varies with EF and how accumulated rainfall varies with EF when afternoon rainfall occurs, defined as follows:

\[
\text{TFS} = \sigma_{\text{EF}} \frac{\partial \Gamma(r)}{\partial \text{EF}},
\]

\[
\text{AFS} = \sigma_{\text{EF}} \frac{\partial \text{E}(r)}{\partial \text{EF}},
\]
where $\sigma_{\text{EF}}$ is standard deviation of EF, $\Gamma(r)$ is probability of afternoon (noon–1800 LST) rain exceeding the 1 mm threshold and $E(r)$ denotes expected value of afternoon rainfall amount. We use the data in AFS calculation only on days when afternoon rainfall does occur. In order to ensure the accuracy of statistics, grids with less than 100 rainfall days when afternoon rainfall does occur. In order to ensure the accuracy of statistics, grids with less than 100 rainfall times are not involved in AFS calculations.

To calculate the partial derivatives in (3) and (4), EF, HIlow and CTP are all treated as discrete random variables with their parameter space divided into different discrete bins, respectively. With reference to the CTP–HIlow framework (Findell and Eltahir, 2003a; 2003b), CTP is divided into three intervals of $(0, 50)$, $(50, 200)$ and $(200, +\infty)$ in units of J/Kg, and HIlow is split into four intervals of $(0, 5)$, $(5, 10)$, $(10,15)$ and $(15, +\infty)$ in units of K, respectively. The observed EF data are divided into 10 bins with each containing an equal number of data points. Unlike the CTP bins and HIlow bins having the pre-defined intervals, the EF bins are not identical at different grid points. For detailed calculation of TFS and AFS, refer to Findell et al. (2011, eqs. 2–4).

3 | RESULTS

The TFS map (Figure 1a) shows that the pattern with significantly large positive TFS values appears in the northern Indian Peninsula, parts of the Mongolian Plateau, and most parts of Myanmar, with peaks over about 16% in the northern Indian Peninsula. That is to say, higher EF enhances the probability of afternoon convection rainfall triggering by up to 16% over there. It is consistent with the results presented in Findell et al. (2011), though the peak signal is weaker than the 25% found in Florida in the United States in that study. In addition to the positive feedback between EF and subsequent afternoon precipitation, negative feedback also appears in south China and Indo-China Peninsula. Obvious negative TFS region did not exist in the United States in summer (Findell et al., 2011; Berg et al., 2013), but negative correlation was reported between morning soil moisture and afternoon convective precipitation accumulations over the U.S. Southern Great Plains under relatively limited synoptic influence with low water vapor convergence (Welty and Zeng, 2018). The AFS map (Figure 1b) indicates that, once triggered, the afternoon rainfall is rather insensitive to EF with the variation of rainfall intensity no more than 1 mm in all regions. What is more, similar results have been obtained by using the TFS* metric, a simplified version of TFS, proposed by Guillod et al. (2014).

To find out possible reasons for the different spatial distribution of TFS, the mean EF (EF) and its standard deviation ($\sigma_{\text{EF}}$), the partial derivative of rainfall probability with respect to EF $\frac{\partial \Gamma(r)}{\partial \text{EF}}$ (hereafter refer to the sensitivity term) and the mean probability of afternoon precipitation $\overline{\Gamma(r)}$ are shown in Figure 2. In the Mongolian Plateau and India, although the mean EF is not large (Figure 2a), especially in Mongolian Plateau and the Indian desert (northwestern-most corner of India), the sensitivity term is large (Figure 2c) and EF varies in a large range with $\sigma_{\text{EF}}$ approximating 0.3 (Figure 2b), resulting in a large positive feedback between the before-noon EF and the occurrence frequency of afternoon precipitation (Figure 1a).

In contrast to the Mongolian Plateau and India, in Myanmar both mean EF (Figure 2a) and probability of afternoon precipitation are higher (Figure 2d), but the EF variability is smaller (Figure 2b). Over this region, the strong triggering strength (Figure 1a) stems from higher sensitivity term (Figure 2c). In other words, the larger sensitivity term may offset the smaller standard deviation of EF. In south China and east Indochinese peninsula, EF is also high, but both the variance of EF and the probability of afternoon precipitation

![FIGURE 1](image_url) The sensitivity of convective triggering and rainfall depth to evaporative fraction. (a) TFS (units of probability of afternoon rain) and (b) AFS (units of millimeters of afternoon rain) in summer. Shading indicates that the mean of the 50 samples is significantly different from zero based on the two-sided t test at the 95% significance level.
are smaller than those in Myanmar, especially the largely reduced probability of afternoon precipitation in south China (Figure 2d). It is the summer East Asian monsoon that a great quantity of moist air is advected over those regions and thus the top of PBL becomes moistening, which provides a favorable background for the formation of deep convective rain at low EF values (Findell and Eltahir, 2003b; Berg et al., 2013).

In order to investigate the relative magnitudes of the triggering and amplification metrics, TFS and AFS are

FIGURE 2 (a) The mean, (b) standard deviation of before-noon EF, (c) the partial derivative of rainfall probability with respect to EF, (d) the mean probability of afternoon precipitation. Note an artifact anomaly at 112.5°E in 2D due to converting the 3-hr dataset to the local time.

FIGURE 3 Normalized, non-dimensional versions of sensitivity map. (a) normTFS and (b) normAFS.
normalized by the ratio of two mean values, $\bar{f}_E$ and $\bar{f}_R$, respectively (Findell et al., 2011) (hereafter normTFS and normAFS, Figure 3). The pattern of normTFS map is similar with TFS map, except in the Myanmar, where the mean probability of afternoon precipitation is very high (Figure 2d). Like the AFS signal (Figure 1b), the signal of norm AFS map is still weak. Obviously, the magnitude of triggering metric is much larger than amplification metric. That is to say, once triggered, the afternoon rainfall intensity is insensitive to EF in eastern Asia, which is consistent with that in United States (Findell et al., 2011).

4 | SUMMARY AND DISCUSSIONS

In this study, two metrics (TFS and AFS) are applied to quantify the sensitivity of afternoon precipitation to EF in eastern Asia regions using ERA datasets. The regions with positive TFS appear in the northern Indian Peninsula, parts of the Mongolian Plateau, and most parts of Myanmar. In addition, over south China and Indo-China peninsula, there exists a negative feedback between EF and the probability of the rainfall occurrence. What is more, EF mainly controls on the frequency of afternoon convection. However, once rainfall is triggered, EF impacts the rainfall amount with a very small scale (<1 mm), which is consistent with the results of Findell et al. (2011) over United States.

We only evaluate the impact of before-noon EF on afternoon precipitation frequency and intensity based on ERA-Interim/LAND dataset. In future work, other datasets will be used to verify present results. In addition, comparison to other land-atmosphere coupling metrics using common datasets would also be useful in fully understanding the characteristics of land-atmosphere coupling over eastern Asia.

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