Climate-induced hotspots in surface energy fluxes from 1948 to 2000

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Abstract
Understanding how land surfaces respond to climate change requires knowledge of land-surface processes, which control the degree to which interannual variability and mean trends in climatic variables affect the surface energy budget. We use the latest version of the Community Land Model version 3.5 (CLM3.5), which is driven by the latest updated hybrid reanalysis-observation atmospheric forcing dataset constructed by Princeton University, to obtain global distributions of the surface energy budget from 1948 to 2000. We identify climate change hotspots and surface energy flux hotspots from 1948 to 2000. Surface energy flux hotspots, which reflect regions with strong changes in surface energy fluxes, reveal seasonal variations with strong signals in winter, spring, and autumn and weak ones in summer. Locations for surface energy flux hotspots are not, however, fully linked with those for climate change hotspots, suggesting that only in some regions are land surfaces more responsive to climate change in terms of interannual variability and mean trends.

Keywords: hotspots, climate change, surface energy flux change

1. Introduction

Global surface air temperatures rose and precipitation patterns changed during the 20th century (e.g. Intergovernmental Panel on Climate Change (IPCC) 2007). The consequence of climate change for the global and regional surface energy budget and hydrological cycle remains unclear because changes in atmospheric forcings are not evenly distributed in the globe and because the land–atmosphere system is nonlinearly coupled. Land surfaces have direct effects on climate through biophysical processes, which determine the absorption and partitioning of energy at land surfaces, and biogeochemical processes, which alter the chemical composition of the atmosphere (e.g. Bonan 2002, Feddema \textit{et al} 2005). Such land-surface processes are, to some extent, determined by a nonlinearly combined consequence of land-surface conditions (e.g. soil moisture and vegetation properties) and atmospheric forcings through the land–atmosphere system (Douville 2003, Koster \textit{et al} 2004, Seneviratne \textit{et al} 2006).

Previous studies indicate that land-surface processes and surface energy fluxes have changed in response to climate-induced changes in atmospheric forcings and soil moisture in the past decades. Such changes in surface energy fluxes are actually reflected, for example, by the decline in pan evaporation in many regions of the world since the 1950s (e.g. Ohmura and Wild 2002). It is believed that the global hydrological cycle, one of the indicators of change in surface energy fluxes, has accelerated in the past decades and varied in both intensity and location (Roderick and Farquhar 2002, Walter \textit{et al} 2004, Stocker and Raible 2005). A recent offline simulation using the Community Land Model (CLM3; Qian \textit{et al} 2007) indicates an increasing trend in terrestrial evaporation and a decreasing trend in sensible heat fluxes over the Mississippi River. Variations in surface energy fluxes in this region have been attributed to changes in climatic forcings: a decreased trend in net radiation in association with observed...
increases in cloudiness and an increase in the basin-averaged precipitation, which is compensated by increases in both runoff and evaporation from 1948 to 2000. In addition, there have been large spatial variations in the surface energy fluxes and hydrological budget in the Mississippi River basin and the contiguous United States (Qian et al. 2007), demonstrating different responses of land surfaces to large-scale atmospheric forcings. It remains unclear, however, to what degrees different land surfaces may respond to interannual variability and the mean trends of climatic forcings through land-surface processes.

Climate change hotspots are defined as the most responsive regions to projected climatic variable change under A1B, A2, and B1 IPCC emission scenarios (e.g. Giorgi 2006). Based on this definition, we hypothesize that these kinds of climate change hotspots should have existed when the same approaches as those in Giorgi (2006) were applied for the historical climate data in the past decades. One of the objectives in this study is to identify climate change hotspots using the global historical climate data from 1948 to 2000 and compare them to those identified from the projected climate data in Giorgi (2006). We also quantify and identify the regions where changes in surface energy fluxes are strongest, i.e., surface energy flux hotspots (hereafter flux hotspots), which are reflected by net radiation hotspots (Rn hotspots), sensible heat flux hotspots (H hotspots), and latent heat flux hotspots (LE hotspots) using results produced by a comprehensive land-surface model (i.e., the National Center for Atmospheric Research Community Land Model version 3.5; CLM3.5) driven by historical climate data from 1948 to 2000. Another objective of this study is to explore possible links between historical climate change hotspots and flux hotspots.

2. Methodology

Three hourly, global hybrid reanalysis-observation forcing data from 1948 to 2000 are available from the Land Surface Hydrology Research Group at Princeton University (the Princeton climate data, Sheffield et al. 2006), with several critical adjustments and updates to reduce errors and uncertainties (Sheffield 2008).

In our study, CLM3.5 is used to simulate components of the surface energy budget and hydrological cycle. CLM3.5, which is an updated version of CLM3 and has considerable improvements in simulating hydrological cycle, is well documented in Oleson et al. (2004, 2008) and has been extensively evaluated by offline comparisons with FLUXNET data (CLM3.0 and CLM3.5, e.g. Stöckli et al. 2008), offline studies on catchment-scale climate-hydrology feedbacks (CLM3.0, e.g. Qian et al. 2007), and coupled experiments (CLM3.0, e.g. Dickinson et al. 2006), indicating consistently good performance in reproducing land-surface fluxes. Driven by the Princeton climate data, our simulations are conducted globally at the resolution of T85 (approximately 1.41° × 1.41°) after 18 year spin-up using the 1948 data to repeat simulations 18 times to ensure that soil moisture has reached a long-term equilibrium.

In our study, we identify climate change hotspots and flux hotspots from 1948 to 2000 by quantifying the climate change index (CCI) and surface energy flux index (SEFI). CCI is derived from seasonal variations in temperature and precipitation over the global land surface, following Giorgi (2006),

\[ CCI = \sum_{season} [n(\Delta P) + n(\sigma_P) + n(WAF) + n(\sigma_T)], \]

where \( n \) is a classification coefficient and taken with the numbers in Giorgi (2006). CCI consists of four seasons: spring (March, April, and May), summer (June, July, and August), autumn (September, October, and November), and winter (December, January, and February). CCI for each season includes four parts: \( \Delta P \) is the change of precipitation for each year relative to the mean precipitation (% relative to mean precipitation) so that the trend contribution of precipitation is included; \( \sigma_P \) is the unbiased standard deviation of precipitation (normalized by the 53 year mean) for each grid after linearly de-trending over the 53 year period; WAF (warming amplification factor) is the change in the mean surface air temperature for each grid relative to the global mean temperature change so that the trend contribution of temperature is included; and \( \sigma_T \) is the unbiased standard deviation in temperature for each grid after linearly de-trending over the 53 year period (similar to \( \sigma_P \) but not normalized).

Besides the CCI, the temperature change index (TCI), precipitation change index (PCI), and vapor pressure deficit change index (VPDCl) are also individually quantified to identify hotspots for each variable. Additionally, the SEFI is reflected in this study by three components: the net radiation change index (NRCI), the sensible heat flux change index (HCI), and the latent heat flux change index (LECI). All these indexes are defined in the same way as CCI:

\[ TCI = \sum_{season} [n(WAF) + n(\sigma_T)], \]
\[ PCI = \sum_{season} [n(\Delta P) + n(\sigma_P)], \]
\[ VPDCl = \sum_{season} [n(\Delta VPD) + n(\sigma_{VPD})], \]
\[ NRCI = \sum_{season} [n(\Delta R_n) + n(\sigma_{R_n})], \]
\[ HCI = \sum_{season} [n(\Delta H) + n(\sigma_H)], \]
\[ LECI = \sum_{season} [n(\Delta LE) + n(\sigma_{LE})]. \]

The classification coefficient \( n \) in equations (2)–(7) is taken as 0, 1, 2, and 4 when changes in variables (i.e., \( \Delta H \), \( \Delta VPD \), \( \Delta LE \), and \( \Delta R_n \)) or their standard deviations (i.e., \( \sigma_H \), \( \sigma_{LE} \), \( \sigma_{VPD} \), and \( \sigma_{R_n} \)) are 0–5%, 5–10%, 10–15%, and >15%, respectively. Based on the definitions of the CCI (i.e., equation (1)) and the SEFI, trend and interannual variability are two essential factors in determining the CCI and SEFI.
Figure 1. Global distributions of (a) climate change index (CCI), (b) temperature change index (TCI), (c) precipitation change index (PCI), and (d) vapor pressure deficit change index (VPDCI). These indexes are directly derived from the climate forcing data and calculated for each grid (180 × 360 grids globally) for the period from 1948 to 2000.

3. Results

3.1. Climate change hotspots and seasonal variations

Figure 1 shows the globally distributed patterns of the CCI, TCI, PCI, and VPDCI produced from the Princeton climate data for 1948 to 2000. The areas with large indexes (highlighted with colors in figures) represent the most prominent hotspots. In general, the climate change hotspots are primarily located in most of the northern high latitudes of North America (i.e., the areas to the north of 50°N), Greenland, northern Europe, central Siberia, central Asia, and central South America (figure 1(a)). It is noted that, however, these climate change hotspots are not fully consistent with those produced in Giorgi (2006) from the latest set of climate change projections by climate models for different emission scenarios. Numerically, the hotspots in Giorgi (2006) are quantified as the difference of the CCI (i.e., the difference in slopes that represent trends, and standard deviations that represent interannual variability of climatic variables) between two periods (i.e., future versus present); while the hotspots in this study are quantified as the CCI for the past 53 years. In this study, it is calculated that temperature change (TCI in figure 1(b)) contributes about 61% to the climate change hotspots over the global land; the contribution from temperature change can be up to 83% in high latitudes; the contribution from precipitation to climate change hotspots is relatively small and highly region dependent, but this contribution can be up to 70% in some small areas in central USA, western Mexico, central South America, and central Asia (i.e., the red areas in figure 1(c)). The precipitation hotspots are located in the central and western USA, Mexico, central South America, and areas from Saudi Arabia to India, followed by the primary hotspots in northern Africa, central Africa, and southern China (figure 1(c)). The most VPD hotspots are located in northern North America, the southern USA, Greenland, northern Europe, central Siberia, central Asia, and western Australia (figure 1(d)). A comparison of the PCI in figure 1(c) and the VPDCI in figure 1(d) shows that changes in VPD are not correlated directly to changes in precipitation. For the PCI and VPDCI, our results indicate that interannual variability contributes to the changes in hotspots more than the long-term trend of the contributing variables, which is consistent with Diffenbaugh et al (2008). For the TCI, however, long-term trends are more important than interannual variability.

Hotspots for climatic variables (temperature, precipitation, and VPD) present seasonal variations in both locations and strengths (figure 2). There is a wider range of temperature hotspots in winter than any other seasons, located in northern North America, Greenland, and on the northern Eurasian continent. These two large hotspots shrink in size in spring, autumn, and summer and show a decrease in strength (figures 2(a)–(d)). Seasonal variations in PCI reveal several distinct precipitation hotspots in the midwestern and southwestern USA, in southwestern Mexico, central South America, northern Africa, and southwestern Asia (from Saudi Arabia to India), with a shift in locations and a change in strengths (figures 2(e)–(h)). The VPDCI shows major VPD
Figure 2. Seasonal variations in the globally distributed temperature change index (TCI) for (a) spring (MAM), (b) summer (JJA), (c) autumn (SON), and (d) winter (DJF); the precipitation change index (PCI) for (e) spring, (f) summer, (g) autumn, and (h) winter; and the VPD change index (VPDCI) for (i) spring, (j) summer, (k) autumn, and (l) winter.
hotspots in the Arctic and boreal regions in North America and parts of Greenland for all seasons, and some scattered hotspots in central Asia and northwestern Australia in spring, autumn, and winter, and northern Russia in summer.

3.2. Surface energy fluxes hotspots and seasonal variations

Figure 3 shows the global maps of seasonal NRCI, HCI, and LECI, based on the radiative budget and surface energy fluxes generated by CLM3.5 driven by the Princeton climate data.
from 1948 to 2000, and reveals distinct hotspots. We notice that three zonal bands with large NRCl (Rn hotspots) are located in the Arctic region in spring (figure 3(a)), high-latitude regions around 55°N in both North America and Eurasia in autumn (figure 3(c)), and mid-latitude regions around 40°N in both North America and Eurasia in winter (figure 3(d)). No obvious Rn hotspots are observed in summer (figure 3(b)). The three zones in the NRCl correspond to interannual variability in snow cover that leads to variations in the surface albedo and thus in net radiation. In spring (figure 3(a)), the Rn hotspot in the Arctic region is sensitive to interannual variability in the timing of spring snowmelt due to past decades of advanced high-latitude warming (e.g. Chapin et al 2005). In summer (figure 3(b)), however, the NRCl is small because interannual variability in land-surface characteristics is small (thus the surface albedo is invariant). The Rn hotspot in the latitude around 55°N in figure 3(c) is associated with interannual variability in the timing of the first autumn snowfall, which has been observed to have changed in the past decades (e.g. Euskirchen et al 2007, Dye 2002). The more southerly, mid-latitude Rn hotspot in winter (around 40°N in figure 3(d)) reflects interannual variability in the spatial extent of snow events in the southern parts of the Northern Hemisphere, which is found to have varied in the past few decades (e.g. Brown 2000).

Changes in net radiative budgets (i.e., Rn hotspots) as well as in other climatic variables lead to changes in partitioning of the surface energy fluxes, which are reflected by variations in the HCI and LECl (i.e., figures 3(e)–(l)). One primary feature is that for each season, changes in the HCI and LECl do not fully correspond to changes in the NRCl or in theCCI that is composed of the TCI, PCI, and VPDCl, indicating a complexity of coupling of the land–atmosphere system. Horizontal advection of heat and moisture is believed to contribute to the global distribution of climate change and, consequently, to climate change hotspots in some regions. Due to the inherent nature of the non-linear relationship between climate change and flux hotspots, the impact of horizontal advection on flux hotspots requires coupled simulations using global circulation models.

In spring, vegetation in most northern high latitudes is still dormant (or partly active in late spring), and sensible heat exchange dominates over the evaporation process. Therefore, changes in net radiation in these regions are balanced primarily by changes in sensible heat fluxes. Under this circumstance, the H hotspots on the northeast coast of Canada, the north coast of the Eurasian continent, and eastern Siberia are roughly in correspondence to the Rn hotspots over these two regions, respectively (figure 3(e) versus figure 3(a)). This explanation is applied for the H hotspots and the Rn hotspots over the Tibetan Plateau. A small H hotspot is also present at the southern end of South America (figure 3(e)). In summer, a few small H hotspots are found on the west coast of northern South America, the east coast of southern South America, and the southeast coast of Australia. The H hotspots in northwestern North America and central North America and the one over Eurasia (figure 3(g) for autumn and figure 3(h) for winter) reflect the influence of changes in net radiation (i.e., the NRCl in these regions in figures 3(c) and (d), respectively) on sensible heat exchange, though the sizes expand over these regions. Note that vegetation over the subtropical and tropical regions is active in spring and hence strongly modulates partitioning of the surface energy budgets. Under this circumstance, the LE hotspots over the west coast of Mexico, northern Africa (including the Sahel), and areas from Saudi Arabia to India are more likely to be associated with the precipitation hotspots over these regions, not only for spring but also for other seasons (figures 3(i)–(l)). In addition, LE hotspots are also identified in areas such as central Asia, Chile, southern Africa, and Australia in all seasons. In winter, these spatial patterns are extended to larger areas on the northern coast of eastern Europe and northwestern Asia. Direct mechanisms for these LE hotspots are, however, unclear, but might be due to complex, non-linear interactions between vegetation and environmental controls (e.g. soil moisture, VPD, and temperature). In low to midlatitudes, for example, interannual variability in LE is largely associated with variations in soil moisture availability (Bonan 2002). In high latitudes, however, interannual variability in LE is considerably affected by variations in air temperature that lead to changes in the lengths of growing seasons and changes in ecosystem functioning (e.g. Euskirchen et al 2007).

4. Discussion and conclusions

Under the past climate change (as indicated by climate change hotspots), changes in land-surface fluxes reflect responses of specific land surfaces to changes in climatic forcings (e.g. their interannual variability and mean trends). Moreover, land-surface characteristics such as soil moisture, ice and snow cover, in some regions also experienced interannual variability due to changes in precipitation and temperature patterns. Interannual variability in precipitation patterns, for example, affects soil water availability for evapotranspiration, while changes in temperature patterns alter variations in snow and ice cover, and thus land-surface albedo, as well as lead to changes in ecosystem functioning. This causes proportional changes in radiative budgets and partitioning of the surface energy budget through land-surface processes. These land-surface processes are, however, regionally and seasonally dependent and non-linearly sensitive to the combined effects of soil moisture and atmospheric forcings for specific land surfaces.

Seasonal variations in the Rn hotspots are observed, and believed to be associated strongly with changes in snow-albedo feedbacks in different latitudes for different seasons. However, patterns of hotspots for H and LE do not correlate fully with the hotspots for climatic forcings (i.e., temperature, precipitation, and VPD) and Rn. The H and LE hotspots show a seasonal shift in both locations and intensity, demonstrating different responses of individual land surfaces to seasonal changes in climatic forcings (i.e., a combined effect of atmospheric forcings and land-surface conditions including vegetation properties and soil moisture). Through comparisons between the spatial distributions of hotspots and the spatial distributions of dominant plant functional types over the global land, our analysis suggests that occurrences of flux hotspots are not
specifically linked with some certain biomes over the global land. Since sensible heat provides boundary layer heating and evaporation affects boundary layer moistening, therefore, regions with H and LE hotspots (i.e., strong land–atmosphere feedbacks) could have more extreme weather activities (e.g. moisture convection precipitation events, summer heat waves) on shorter timescales and have more significant contributions to strong seasonal or interannual variability in climate on longer timescales. However, regions with climate change hotspots could, on longer timescales, have a large impact on land-surface properties (e.g. high-latitude albedo change as a result of advanced spring warming that alters snow cover) and even ecosystem structures (e.g. warming-induced enhanced shrub growth and enhanced greenness in the northern high latitudes) (Sturm et al. 2001, Nemani et al. 2003, Chapin et al. 2005). Such changes in land-surface properties have contributed considerably to climate change through land and climate feedback (e.g. Chapin et al. 2005).

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