The Role of Land–Atmosphere Interactions for Climate Variability in Europe

S. I. Seneviratne[1] **and R. Stöckli**[1,2]

Abstract We provide here a brief review on the role of land–atmosphere interactions for climate variability, with a special focus on the European continent. First, an overview of the land energy and water balances and of the underlying physical, biophysical, and biogeochemical soil–vegetation–atmosphere processes is presented. Further, we highlight how land–atmosphere feedbacks can impact seasonal to interannual climate variability in transitional climate zones and midlatitude regions along three main paths: Soil moisture–temperature interactions, soil moisture–precipitation interactions, and vegetation–climate interactions. In this context, we discuss recent results based on findings from terrestrial observational networks, satellite observations, and numerical climate models across a number of spatial and temporal scales. These results illustrate the extent to which land-surface processes, land–atmosphere interactions, and associated memory effects can modulate the dynamics of the climate system. Finally, the concluding section addresses current areas of uncertainty and open questions for research in this field.

1 Introduction

The importance of land–atmosphere interactions and all processes they involve for the climate system is increasingly being recognized. Similar to the oceans, land areas provide the lower boundary for the atmosphere, with which they exchange energy, water and chemical compounds such as CO_2 (Fig. 1). Storage of water on land (e.g., as soil moisture, groundwater, snow, surface water or ice) constitutes a significant memory component within the climate system, similar in many ways to heat storage in the oceans. Moreover, anomalies of soil moisture (positive or negative)

[1] Institute for Atmospheric and Climate Science, ETH Zurich, Zurich, Switzerland
sonia.seneviratne@env.ethz.ch

[2] Department of Atmospheric Science, Colorado State University, Fort Collins, CO, USA
reto.stoeckli@env.ethz.ch

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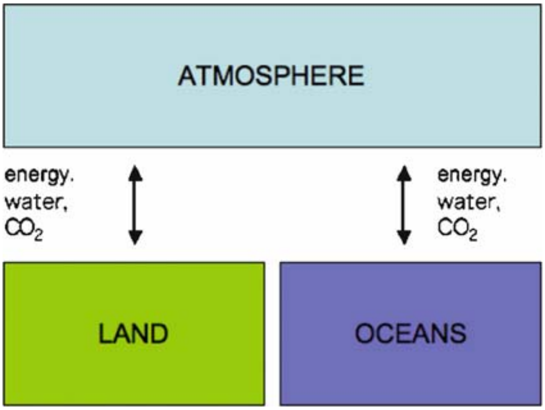

Fig. 1 Land–atmosphere and oceans–atmosphere interactions.

have strong effects on the land energy and water balances in regions where evapotranspiration is limited by soil moisture availability. Consequently, also vegetation processes are critical in constraining the uptake of soil moisture for evapotranspiration.

While the role of land for climate variability has often been neglected in the past, recent studies have highlighted how land–atmosphere interactions can be critical in modulating variations in climate on a range of temporal (seasonal to centennial) and spatial (local to global) scales. In particular, the role of soil moisture for precipitation (e.g., Betts et al. 1996; Beljaars et al. 1996; Eltahir 1998; Schär et al. 1999; Betts 2004; Koster et al. 2004a) and temperature (e.g., Seneviratne et al. 2006a) in midlatitude and transitional climate zones has been highlighted in several investigations. Soil moisture is also an important memory component of the climate system (e.g., Koster and Suarez 2001; Seneviratne et al. 2006b) and thus a useful source of persistence for seasonal forecasting (e.g., Koster et al. 2004b; Ferranti and Viterbo 2006). Further relevant land–atmosphere interactions in the framework of climate change involve interactions with the carbon cycle and in particular links between CO_2 assimilation and water use in plants (e.g., Field et al. 1995; Körner 2000; Gedney et al. 2006).

In this review, we focus more particularly on the role of land–atmosphere interactions for the seasonal-to-interannual variability of the European summer climate. However, land–atmosphere interactions have been shown to be relevant for several other regions and time scales. The recent Global Land-Atmosphere Climate Experiment (GLACE, Koster et al. 2004a, 2006) pinpointed that land–atmosphere interactions tend to be particularly important in transitional zones between dry and wet climates. For present climate, this applies, e.g., to the Great Plains of North America, the Sahel, equatorial Africa, and Northern India (Koster et al. 2004a), but also to the Mediterranean region (Seneviratne et al. 2006a). Moreover, these “hot spots” of land–atmosphere coupling are also inherently modified with shifts in climate regimes, for instance due to climate change (Seneviratne et al. 2006a). They can thus be displaced on longer time scales. Finally, long-term vegetation dynamics and human-induced land use changes can also interact with the rest of the climate system on decadal-to-centennial time scales (Cramer et al. 2001; Claussen et al. 2004; Pielke 2005). These longer-term feedbacks will not be treated in detail as part of the present review.

The structure of this chapter is as follows. Section 2 provides a short overview of the processes governing land energy and water balances, and of their interconnections; Sect. 3 presents soil moisture–temperature interactions; Sect. 4, soil moisture–precipitation interactions; and Sect. 5, vegetation–climate interactions. Finally, we present a summary and an outlook in Sect. 6.

2 Land Energy and Water Balances

The land energy balance for a surface soil layer (including possible snow or ice cover) can be expressed as:

$$\frac{dH}{dt} = R_n - \lambda E - SH - G \tag{1}$$

where dH/dt is the change of energy within the considered surface layer (e.g., temperature change, phase changes), R_n is the net radiation, λE is the latent heat flux (latent heat of vaporization λ times the evapotranspiration E), SH is the sensible heat flux and G is the ground heat flux to deeper layers (Fig. 2, left).

Similarly, the land water balance for a surface soil layer (including a possible snow or ice cover) is expressed as:

$$\frac{dS}{dt} = P - E - R_s - R_g \tag{2}$$

where dS/dt is the change of water content within the considered layer (e.g., changes in soil moisture, snow content, ice content, surface water, groundwater), P is the precipitation, E is the evapotranspiration, R_s is the surface runoff, and R_g is (depending on the soil depth considered) the drainage or groundwater runoff (Fig. 2, right).

Equations (1) and (2) show that the land energy and water balances are coupled through the evapotranspiration term. If soil moisture is lacking, then no evapotranspiration can take place and most of the incoming energy (net radiation) goes into sensible heat flux, thus strongly enhancing air temperature. Reversely, if water is available in ample supply (moist surface or water body), then a large amount of energy will be used for evapo(transpi)ration thus effecting a net cooling compared to dry surfaces. These effects are, however, only important in regions where soil moisture is the main controlling factor for evapotranspiration. In high-latitude regions, for instance, evapotranspiration is limited by net radiation and the length of the growing season.

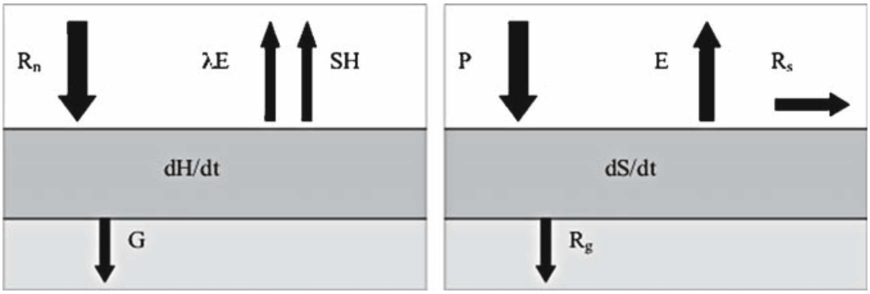

Fig. 2 Land energy (left) and water (right) balances.

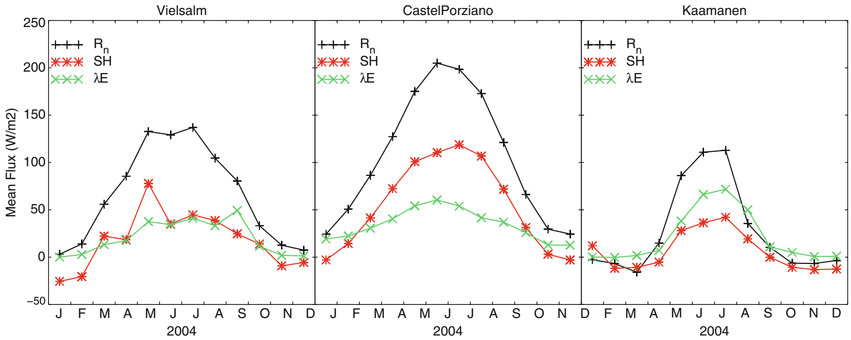

Fig. 3 Monthly net radiation (R_n), latent heat flux (λE) and sensible heat flux (SH) during 2004 at three CarboEurope flux tower sites covering a wide range of climate zones: Vielsalm, Belgium (temperate mixed forest, Aubinet et al. 2001); Castelporziano, Italy (summer-dry Mediterranean evergreen forest, Reichstein et al. 2002); and Kaamanen, Finland (Arctic tundra, Aurela et al. 2002). Fluxes were averaged from the original Level 2 CarboEurope data set.

Surface energy and water exchanges can for instance be analyzed from flux tower measurements provided by CarboEurope (http://www.carboeurope.org/). As an example, we display in Fig. 3 surface fluxes from three sites encompassing a wide range of climate zones: Vielsalm, Belgium (temperate mixed forest); Castelporziano, Italy (summer-dry Mediterranean evergreen forest); and Kaamanen, Finland (Arctic tundra). The temperate forest (left) has a well-balanced distribution between the monthly λE and SH fluxes. At the Mediterranean site (middle), lack of precipitation resulting in low soil moisture severely constrains vegetation activity; this limits the λE flux, thus inducing an enhanced SH flux. The arctic Tundra site's (right) seasonal courses of SH and λE are controlled by the short growing season and the low magnitude of available energy R_n.

Finally, the land energy and water balances are themselves linked with the terrestrial carbon as carbon assimilation and evapotranspiration are tightly coupled. These aspects will be discussed in more detail in Sect. 5.

3 Soil Moisture–Temperature Interactions

The main mechanism by which soil moisture can impact air temperature has been discussed in the previous section. Namely, soil moisture exerts a strong control on the partitioning of incoming surface energy (net radiation) in the latent and sensible heat fluxes in any region where it is the limiting factor for evapotranspiration. The fact that these regions are often transitional zones between dry and wet climates (e.g., Koster et al. 2004a) can be understood in the following way: In constantly wet regions, soil moisture is not a limiting factor for evapotranspiration and thus will not have a strong impact on the land energy balance; in constantly dry regions (deserts), there is too little soil moisture to allow significant evapotranspiration, independent of the season or of interannual variations in further climate variables. However, in regions where soil moisture can vary seasonally and interannually between dry and wet conditions, it will necessarily be an important factor impacting temperature variability.

For Europe, Seneviratne et al. (2006a) recently investigated how land–atmosphere coupling impacts summer temperature variability in present- and future-climate Regional Climate Model (RCM) simulations. For present climate, land–atmosphere coupling is found to have a significant impact in the Mediterranean region, where ca. 60% of the simulated interannual variability of summer temperature is due to interannual variations in soil moisture content (Fig. 4). Observational evidence for a link between spring precipitation deficits and summer temperature in the Mediterranean region for present climate (Della-Marta et al. 2007) lends support to these modeling results. Interestingly, the GLACE study (Koster et al. 2004a, 2006) did not identify a strong impact of soil moisture for temperature or precipitation in this region. This may be due to the setup of this study, for instance, because the impact of interannual variations in sea surface temperatures was not considered (see also discussion in Seneviratne et al. 2006a).

In future climate, the same study finds strong soil moisture–temperature coupling in most of Central and Eastern Europe (Fig. 5). This is due to a gradual shift of climatic regimes within the continent, whereby the transitional climate zone is shifted northward from the Mediterranean region to Central and Eastern Europe. This shift also appears responsible for the very large increase in summer temperature variability projected in this region (e.g., Schär et al. 2004). A further analysis of the correlation between evapotranspiration and temperature (which can be seen as an indirect measure of coupling) in the RCM experiments and in Global Climate Model (GCM) simulations from the Intergovernmental Panel on Climate Change (IPCC) 4th Assessment Report (AR4) confirmed this result (Seneviratne et al. 2006a).

4 Soil Moisture–Precipitation Interactions

The possible existence of strong soil moisture–precipitation feedbacks has been the topic of several investigations, based either on observations (e.g., Betts et al. 1996) or modeling studies (e.g., Beljaars et al. 1996; Schär et al. 1999; Pal and Eltahir

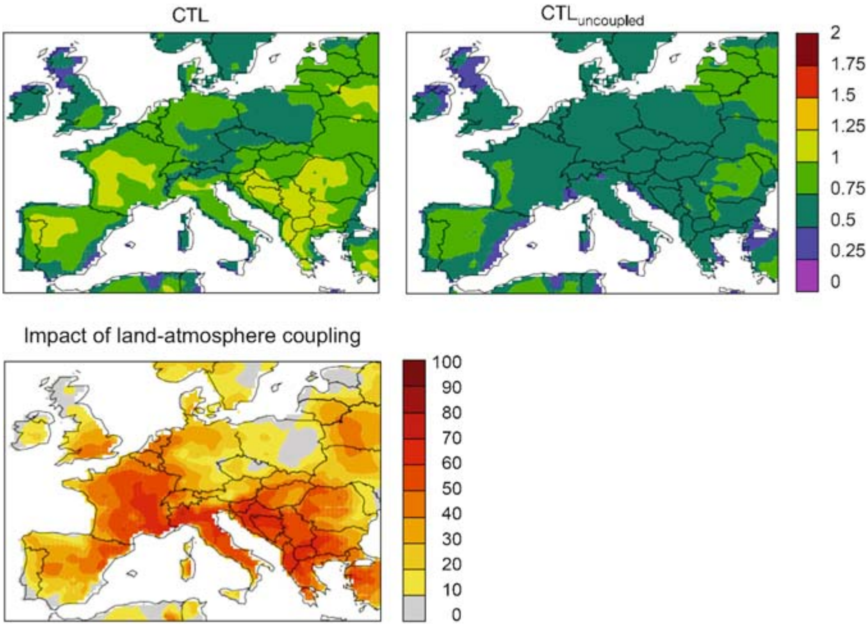

Fig. 4 Impact of land–atmosphere coupling for interannual summer (JJA) temperature variability in simulations for the time period 1970–1989. (Top) Standard deviation of the JJA 2 m temperature in the "coupled" (left) and "uncoupled" (right) experiment (K). (Bottom) Percentage of interannual JJA temperature variance due to land–atmosphere coupling (%) (from Seneviratne et al. 2006a).

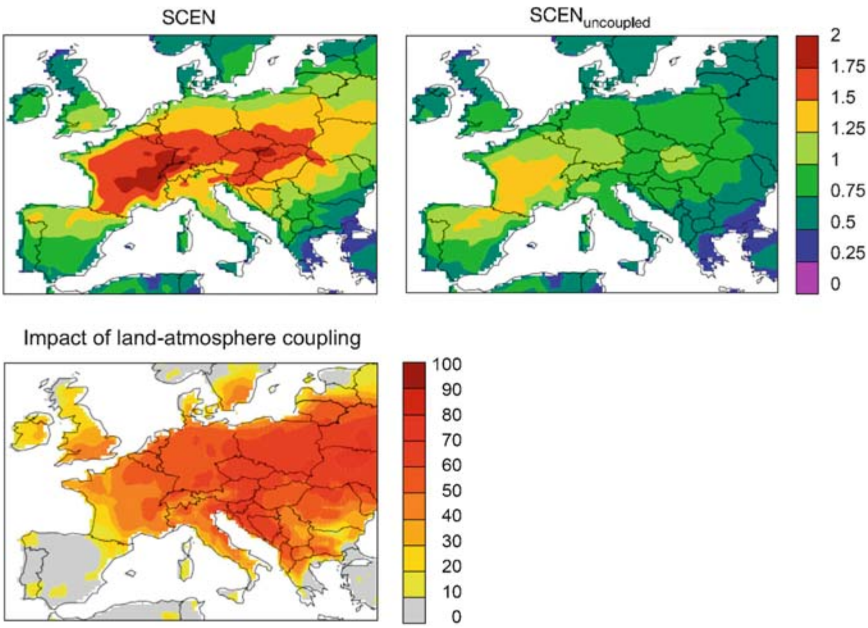

Fig. 5 Same as Fig. 4 for the time period 2080–2099 (from Seneviratne et al. 2006a).

2001; Koster et al. 2004a). The processes involved are more complex than a simple recycling mechanism by which additional moisture input from evapotranspiration to the atmosphere would lead automatically to additional precipitation. Rather, it appears that the suite of processes leading to a positive feedback loop between soil moisture content and subsequent precipitation involves modifications of the boundary layer structure and of the atmospheric stability profile (e.g., Eltahir 1998; Schär et al. 1999; Betts 2004).

As for temperature, soil moisture mostly appears to impact subsequent precipitation in transitional regions between dry and wet climate (Koster et al. 2004a). Again, this is due to the fact that soil moisture is not a limiting factor for evapotranspiration in wet climate, and that evapotranspiration is too limited in dry climate to significantly impact the regional climate system. For Europe, the GLACE study did not identify strong soil moisture–precipitation coupling (Koster et al. 2004a) in this region. However, as for soil moisture–temperature coupling, this could possibly be due to the setup of the numerical experiments (see preceding section). While investigations of possible soil moisture–precipitation coupling from observations is difficult in Europe due to the lack of soil moisture observations, modeling studies did find some impact of soil moisture for subsequent precipitation (e.g., Schär et al. 1999; Fischer et al. 2007). For instance, Schär et al. (1999) investigated the impact of initial soil moisture in 2-month-long RCM simulations for the months of July 1990 and July 1993. Their findings suggest that soil moisture anomalies can have strong impact on subsequent precipitation in Spain, France and Central Europe (Fig. 6). Soil moisture–precipitation feedbacks also appear relevant for future increases in precipitation variability in Europe, mostly in the Alpine region (Seneviratne et al. 2006a).

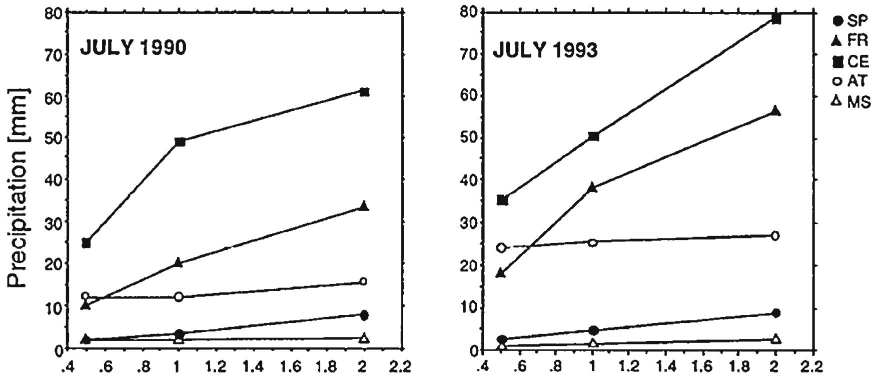

Fig. 6 Impact of initial soil moisture content on subsequent precipitation in various European regions (SP: Spain, FR: France, CE: Central Europe, AT: Atlantic, MS: Mediterranean Sea). Displayed is the simulated total precipitation in month-long regional climate model experiments for July 1990 and 1993, as function of the factor applied to the initial moisture content (from Schär et al. 1999. Copyright AMS).

5 Vegetation–Climate Interactions

In vegetated landscape water flux from the soil to the atmosphere mostly occurs through leaf stomates during the growing season (Sellers et al. 1997; Bonan 2002) and to a lesser extent through soil evaporation. During photosynthesis plants open stomates to take up CO_2. This chemical process is primarily driven by the light energy but its rate depends on various environmental conditions and the plant's biophysical and biochemical state. While stomates are open, water leaks from the saturated leaf interior (Farquhar et al. 1980; Ball et al. 1987) and has to be redrawn from the soil through the plant's root system in order to avoid dessication. Thus, evapotranspiration from vegetated surfaces occurs mostly as a by-product of photosynthesis and is constrained by this process. Sunlight, soil moisture, atmospheric vapor pressure, temperature and carbon dioxide concentration are the main physical environmental regulators for photosynthesis (Dickinson 2001), but it is also modulated by a number of biotic and abiotic factors such as tree age, nutrient availability, pests and the phenological state of the plant.

Hence, while vegetation cover, plant growth and photosynthesis are obviously strongly constrained by regional climate, they can also have a significant impact on climate on seasonal to interannual time scales. Many observational and modeling studies document such effects in Europe. For instance, in a study for the summer 2003 heat wave in France, Zaitchik et al. (2006) show how air temperature can be sensitive to vegetation cover in this region: Using satellite data, they identify that temperature differences between August 10, 2003 (at the peak of the heat wave) and a normal August day in 2000 (August 1, 2000) are much higher over pasture/active crops (+20°C) than forest (+11°C) areas. This complex spatial pattern is even more evident in Fig. 7 where the highest anomalies of MODIS radiative land surface temperatures for summer 2003 are concentrated in the predominantly agricultural areas of central France.

A better resilience of the forest areas to the heat wave conditions similarly shows up in Normalized Difference Vegetation Index (NDVI) measurements (a measure for vegetation activity): For the pasture/active crops areas, NDVI on August 10, 2003 corresponds to 50% of the value in 2000, while forested areas show no differences (Zaitchik et al. 2006). Ciais et al. (2005) found that in general drought-tolerant ecosystems in the Mediterranean had a lower response to the unusual 2003 conditions than more drought-susceptible temperate vegetation in central Europe. A complex pattern of impacts was seen for Alpine ecosystems (Jolly et al. 2005): Longer-growing seasons were observed at high elevations due to a longer snow-free period while lower elevations were experiencing a shorter growing due to temperature and moisture stress.

These temporal variations in vegetation greenness feed back on the hydrological cycle over land through modifications of the surface heat, water (Guillevic et al. 2002) and carbon balances (Schaefer et al. 2005). Start and length of the growing season over Europe derived from 20 years of NDVI data reveal a significant interannual variability in European phenology (Stöckli and Vidale 2004; Studer et al.

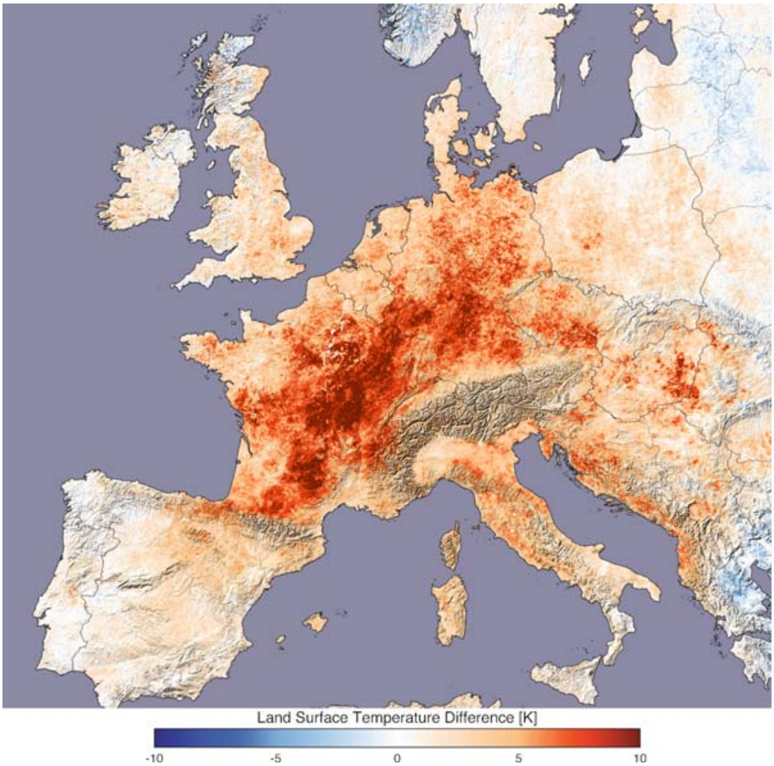

Fig. 7 MODIS (MODerate resolution Imaging Spectroradiometer) radiative land surface temperature (LST) anomaly over Europe for 2003: LST difference (all cloud-free pixels during July 20–August 20) for the years 2000, 2001, 2002, 2004, 2005 and 2006 subtracted from 2003. (Visualization by R. Stöckli as published in Allen and Lord 2004 but with updated data from 2005, 2006).

2007). These patterns of vegetation phenological states act in concert with climatic drivers such as temperature and precipitation (Los et al. 2001; Zhang et al. 2004). Accounting for interannual variability of phenology in numerical weather prediction (Chapeaux et al. 2000) and climate modeling (Bounoua et al. 2000; Lu and Shuttleworth 2002) results in differences of up to 0.9 K in air temperature in Mediterranean and Central Europe and can modify precipitation rates up to 9 mm/month in northern latitudes.

Climate change might further alter these vegetation–climate feedback processes. Twentieth-century warming already resulted in earlier springs and generally longer growing seasons over Europe (Menzel 2000; Defila and Clot 2001; Studer et al. 2005; Menzel et al. 2007). These trends are likely linked to increases in temperature and decreases in snow duration in temperate and alpine ecosystems during the last century. Furthermore, higher atmospheric CO_2 levels can possibly lead to enhanced water-use efficiency in plants (e.g., Field et al. 1995). In a recent modeling study, Gedney et al. (2006) suggest for instance that observed positive runoff trends during the 20th century may be due to this effect. Also ground observations of CO_2-enriched trees document such water saving effects, though the absolute response is highly species dependent (Leuzinger et al. 2005).

Reversely, potential future modifications of the hydrological cycle, such as more frequent drought conditions, could also have an impact on the carbon cycle. In their study for the 2003 heat wave, Ciais et al. (2005) performed a thorough analysis of 15 CarboEurope tower sites, and found that 2003 switched Central Europe from a net carbon sink to a net carbon source. Gross Primary Production (GPP, i.e., the carbon uptake of plants during photosynthesis) decreased by 30% as a result of rainfall deficit and heat. Despite the high soil temperatures, the carbon loss through ecosystem respiration also decreased slightly, heterotrophic decomposition being inhibited by the prevailing dry soil conditions. A decreased vegetation carbon sink would ultimately result in a further enhancement of atmospheric CO_2 concentrations and consequent impacts on the climate system. Finally, direct CO_2 effects on carbon assimilation may also be possible (enhanced carbon assimilation in enhanced CO_2 conditions), though these may be more limited than previously assumed (Körner et al. 2005).

6 Conclusions and Outlook

We have seen in this review that the interactions between land and the atmosphere are manifold and can impact climate variability along various paths. We have focused here only on Europe and on seasonal-to-interannual climate variability, but land-atmosphere interactions can impact climate in many other regions (e.g., Koster et al. 2004a) and on a much wider range of temporal scales (Claussen et al. 2004).

While the study of land–atmosphere interactions offers promising perspectives for future research, there are also some open issues impeding progress in this field. The main limitation is the lack of ground observations of key variables such as soil moisture or evapotranspiration. The Global Soil Moisture Data Bank (Robock et al. 2000) provides access to soil moisture observations from several measurement networks around the globe, but data is lacking in many regions and in particular in Europe. The Fluxnet network (Baldocchi et al. 2001) provides flux measurements of energy, water (evapotranspiration) and CO_2 around the world, and especially in Europe (through CarboEurope). Some studies have shown the usefulness of these measurements for the process-based assessment of climate models (Stöckli and Vidale 2005; Teuling et al. 2006). Nonetheless, for certain applications the data set lacks spatial and temporal continuity.

In this light, approaches that allow to obtain indirect estimates of relevant land surface quantities such as soil moisture or evapotranspiration are very promising and could significantly advance research in this field. For instance, combined atmospheric–terrestrial water balance estimates using reanalysis data and runoff observations (Seneviratne et al. 2004; Hirschi et al. 2006a) have been shown to provide useful information on basin-scale variations in terrestrial water storage and have been employed in several applications (e.g., Andersen et al. 2005; Hirschi et al. 2006b, 2007; Jacob et al. 2007; Seneviratne et al. 2006b; van den Hurk et al. 2005). Moreover, several satellite data products show some promising results such as the Gravity Recovery and Climate Experiment Mission (e.g., Tapley et al. 2004; Rodell et al. 2004a), microwave remote sensing products (e.g., Reichle and Koster 2005), and radiometrically derived biophysical vegetation products, e.g., land cover maps, NDVI or LAI (Tucker et al. 1985; Champeaux et al. 2000; Justice et al. 2002; Stöckli and Vidale 2004; Running et al. 2004). Finally, approaches combining observations and model data such as the Global Soil Wetness Project (Dirmeyer et al. 1999, 2002), the Global Land Data Assimilation System (Rodell et al. 2004b), and other land data assimilation products might ultimately help to obtain reliable global estimates of the relevant climate variables.

In the area of ecosystem fluxes and vegetation–climate interactions, improving our process-based understanding at different scales will require the integration of ground and space-based observational networks and numerical modeling initiatives, combined with the complementary collaboration of different research communities (Canadell et al. 2000; Running et al. 1999; Turner et al. 2004). For instance, bottom-up studies from flux towers allow ecosystem researchers to document the seasonal-to-interannual biospheric functioning in response to climate variability. They are valuable both for the development and the validation of empirical and process-based ecosystem models (Running et al. 1999; Stockli and Vidale 2005). These data sets have nonetheless some limitations, in terms of availability and consistency (Houghton 2003), as well as with regard to their applicability to regional-to-global scale processes. Inverse modeling through data assimilation can for instance provide effective means to generate regional flux estimates of the global hydrological and biogeochemical cycle based on a heterogeneous and incomplete distribution of local measurements (Gurney et al. 2002). By linking such top-down measurements with local-scale bottom-up ecosystem measurements, the mentioned scale gap can possibly be overcome (e.g., Denning et al. 2003).

In conclusion, the investigation of land–atmosphere interactions and their role for seasonal to interannual climate variability is a growing interdisciplinary research field offering significant promises for climate research. Land–atmosphere interactions are relevant for climate predictability on several time and spatial scales, and their better understanding could advance many climate applications important for society such as seasonal forecasting and climate-change modeling. Finally, we have shown that they are relevant in many ways for climate in Europe and should therefore be better investigated on this continent, especially in terms of improved soil moisture networks and the integrated analysis of the already existing models and data.

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