Soil Moisture–Atmosphere Interactions during the 2003 European Summer Heat Wave

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ABSTRACT

The role of land surface–related processes and feedbacks during the record-breaking 2003 European summer heat wave is explored with a regional climate model. All simulations are driven by lateral boundary conditions and sea surface temperatures from the ECMWF operational analysis and 40-yr ECMWF Re-Analysis (ERA-40), thereby prescribing the large-scale circulation. In particular, the contribution of soil moisture anomalies and their interactions with the atmosphere through latent and sensible heat fluxes is investigated. Sensitivity experiments are performed by perturbing spring soil moisture in order to determine its influence on the formation of the heat wave. A multiyear regional climate simulation for 1970–2000 using a fixed model setup is used as the reference period.

A large precipitation deficit together with early vegetation green-up and strong positive radiative anomalies in the months preceding the extreme summer event contributed to an early and rapid loss of soil moisture, which exceeded the multiyear average by far. The exceptionally high temperature anomalies, most pronounced in June and August 2003, were initiated by persistent anticyclonic circulation anomalies that enabled a dominance of the local heat balance. In this experiment the hottest phase in early August is realistically simulated despite the absence of an anomaly in total surface net radiation. This indicates an important role of the partitioning of net radiation in latent and sensible heat fluxes, which is to a large extent controlled by soil moisture. The lack of soil moisture strongly reduced latent cooling and thereby amplified the surface temperature anomalies.

The evaluation of the experiments with perturbed spring soil moisture shows that this quantity is an important parameter for the evolution of European heat waves. Simulations indicate that without soil moisture anomalies the summer heat anomalies could have been reduced by around 40% in some regions. Moreover, drought conditions are revealed to influence the tropospheric circulation by producing a surface heat low and enhanced ridging in the midtroposphere. This suggests a positive feedback mechanism between soil moisture, continental-scale circulation, and temperature.

1. Introduction

A record-breaking heat wave affected the European continent in summer 2003. With mean summer [June–August (JJA)] temperatures exceeding the 1961–90 mean by about 3°C over large areas and by over 5°C regionally (Schär et al. 2004, hereafter S04), it was very likely the hottest European summer over the past 500 yr (Luterbacher et al. 2004). In central and southern Europe the socioeconomic impact of this extraordinary heat wave was disastrous. Estimates by the World Health Organization (WHO) and the Earth Policy Institute indicate a statistical excess over the mean mortality of between 22 000 and 35 000 heat-related deaths across Europe (Larsen 2003; Vandentorren et al. 2004; Hémon and Jougla 2004; Koppe and Jendritzky 2004; Schär and Jendritzky 2004). During the maximum heat wave, in the first 2 weeks of August, the mortality rate in France increased by 54% (Hémon and Jougla 2004).

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The financial loss due to crop failure over southern, central, and eastern Europe is estimated at $12.3 billion (Heck et al. 2004). Forest fires in Portugal alone resulted in total damage costing $1.6 billion (Heck et al. 2004). In addition to these socioeconomic impacts, the 2003 heat wave was associated with a mass loss of the Alpine glaciers of 5%–10% (Zemp et al. 2005). The high temperatures led to excessive near-surface ozone concentrations with major impacts upon the continental-scale air quality (Solberg et al. 2005; Ordonez et al. 2005).

Temperature anomalies as observed in JJA 2003 are statistically highly unlikely, even when the observed warming is taken into account (S04). Schönwiese et al. (2004), using a statistical time series analysis on German surface temperature series, reveal an increasing probability of hot summers taking place along with a warming trend observed especially within the recent decades (Klein Tank et al. 2005). Stott et al. (2004) point out that this increasing probability of anomalously warm summers can partly be attributed to past human influence on the climate system. Several model studies suggest that events such as the 2003 summer heat wave will become more frequent, more intense, and longer lasting in the future (S04; Beniston 2004; Meehl and Tebaldi 2004; Vidale et al. 2007). Several studies have suggested that the projected changes in summer climate strongly rely on soil moisture–atmosphere interactions (Seneviratne et al. 2006b; Rowell 2005; Rowell and Jones 2006; Vidale et al. 2007).

Heat waves are generally associated with specific large-scale anticyclonic atmospheric circulation patterns (Black et al. 2004). These patterns are often characterized by quasi-stationary 500-hPa height anomalies that dynamically produce subsidence, clear skies, light winds, warm-air advection, and prolonged hot conditions at the surface (Kunkel et al. 1996; Palecki et al. 2001; Xoplaki et al. 2003; Meehl and Tebaldi 2004). The surface temperature response to such circulation anomalies is amplified by a positive feedback due to suppressed evapotranspiration owing to the lack of soil moisture (S04; Hartmann 1994; Lakshmi et al. 2004; Ferranti and Viterbo 2006; Seneviratne et al. 2006b).

Several studies have analyzed the mechanisms that contributed to the formation of previous summer heat waves. Cassou et al. (2005) demonstrate that anomalous tropical Atlantic heating may significantly favor the formation of Rossby wave trains and atmospheric blocking conditions as in 2003. Vautard et al. (2006) show that rainfall deficiencies in the preceding winter Mediterranean precipitation have a discernable effect on the frequency of heat waves through northward transport of latent heat fluxes. Della-Marta et al. (2006) suggest that preceding winter North Atlantic SSTs and January–May Mediterranean precipitation (and thus spring soil moisture anomalies) may provide predictive skill for summer heat waves. On a multidecadal time scale, the variability of European summer temperatures has been attributed to basin-scale changes in the Atlantic Ocean, associated with the Atlantic multidecadal oscillation (Sutton and Hodson 2005).

In the specific case of the 2003 heat wave, the preceding spring was anomalously warm throughout central and western Europe. This warm anomaly was accompanied by a persistent precipitation deficit (Fink et al. 2004), starting as early as February and lasting until the end of summer 2003. Using satellite imagery and meteorological station data, Zaitchik et al. (2006) demonstrate that the early vegetation green-up due to springtime warmth, together with the lack of precipitation, resulted in an early season soil moisture deficit. Black et al. (2004) and Zaitchik et al. (2006) point out that the first extreme temperature anomaly in June was mainly the result of the persistent anticyclonic circulation anomaly. During the entire month of June a characteristic wave pattern of 500-hPa height anomalies featuring deep troughs over the eastern Atlantic and western Russia, and ridges over Europe and central Russia, was observed (Ferranti and Viterbo 2006). The anomalously clear skies and the extremely strong radiative anomalies in June contributed to further loss of soil moisture. In July, few Atlantic frontal systems penetrated as a result of a blocking pattern over Scandinavia and a pronounced split-flow configuration farther south (Ferranti and Viterbo 2006), while temperature anomalies were somewhat less extreme (+2°C above the climatological mean). Temperature anomalies culminated in a maximum heat wave over France between 2 and 12 August. This episode was associated with an abnormally positive phase of the summer Northern Annular Mode (NAM) index (Ogi et al. 2005) and unusually low upper-level ozone concentrations (Orsolini and Nikulin 2006). Black et al. (2004) suggest that the exceptionally high temperatures in late summer 2003 were mainly the consequence of the atmospheric circulation enabling a dominance of the local heat balance over Europe. The severe late-summer drought contributed to strongly reduced latent heat flux and positive sensible heat anomalies (Black et al. 2004; Zaitchik et al. 2006).

In addition to land surface temperatures, strong mean Mediterranean SST anomalies of about 2.1°C were observed during JJA 2003 (Jung et al. 2006; Grazzini and Viterbo 2003). While SST anomalies exceeded the long-term standard deviation by more than a factor of 3 in the western part, they were far smaller in the southeastern part of the Mediterranean. However, Jung et al. (2006) and Ferranti and Viterbo (2006) suggest that the enhanced SSTs in the Mediterranean had a marginal influence on the midtropospheric dynamical circulation in Europe and rather followed the tropospheric temperature signal. Black and Sutton (2006) challenge these results, suggesting that the SST anomalies in both the Indian Ocean and in the Mediterranean Sea had a significant influence on the 2003 heat wave. They point out that the Mediterranean contributed most strongly to the early part of the heat wave and the Indian Ocean enabled the positive temperature anomalies to persist into August.

The overall goal of this study is to improve our understanding on the land surface–related processes and feedbacks that led to the extreme anomalies during summer 2003 by means of simulations and observational data. We perform a set of regional climate simulations for the year 2003 using the European Centre for Medium-Range Weather Forecasts (ECMWF) operational analysis as lateral boundary conditions. Thereby we ensure that all simulations are driven with a realistic representation of the large-scale atmospheric circulation.

The first specific focus of our study is to establish the extent to which such a modeling framework is able to replicate an extreme event such as summer 2003. To this end, the simulated fields are compared to a 31-yr-long model simulation (1970–2000). The 2003 anomalies are carefully validated with independent observational temperature and precipitation fields as well as diagnostic estimates of terrestrial water storage variations (Seneviratne et al. 2004; Hirschi et al. 2006b) in the months prior to and during the heat wave. Note that the term "heat wave" is often defined as a number of consecutive days with temperatures exceeding some threshold. In this sense, summer 2003 was associated with at least three distinct heat waves. However, in the current study we use the term heat wave in a seasonal sense, while addressing the temporal variations of the signal in considerable detail.

The second focus of this study is to assess the role of soil moisture–atmosphere interactions as a key feedback mechanism for the 2003 summer heat wave. We find that the soil moisture content was relatively high in the beginning of the year 2003 due to anomalously high precipitation in summer and autumn 2002 over central and eastern Europe (e.g., the flooding event in August 2002; see Rudolf et al. 2003; Christensen and Christensen 2003). However, the persistent precipitation deficit and the excess in total net radiation in late winter and spring 2003 contributed to an early and rapid soil drying. To investigate the effect of this anomalous soil moisture depletion upon the subsequent summer drought, we perform a sensitivity experiment, which includes 10 simulations with perturbed spring soil moisture. Soil moisture is artificially increased and decreased on 1 April to identify the sensitivity to different soil moisture conditions given the same large-scale atmospheric circulation. The difference of the control and the wettest simulation, which produces approximately long-term mean late summer soil moisture conditions, provides of a rough estimate on the soil moisture contribution to the heat wave. We analyze the response of temperature, precipitation, and atmospheric circulation to different soil moisture conditions, as well as the related anomalies in the water and surface energy budget.

An earlier study by Ferranti and Viterbo (2006) analyzed the influence of different soil moisture initializations in June 2003 upon the seasonal forecast of the subsequent three summer months. They found a significant atmospheric response to substantial soil moisture perturbations during 2–3 months. Note that in the Ferranti and Viterbo (2006) framework, there is no guarantee that the atmospheric circulation is close to its observed evolution. Indeed, their simulations did not replicate the characteristic 2003 circulation anomalies, whereas our study addresses the soil moisture–atmosphere feedback processes using a best estimate of the observed continental-scale circulation (ECMWF operational analysis). In contrast to Ferranti and Viterbo (2006), we prescribe the circulation at the lateral boundaries and thereby ensure a realistic representation of the observed persistent circulation anomalies. This allows us to isolate the response to different soil moisture conditions given identical lateral boundary conditions. Moreover, in our simulations the effect of the imposed soil moisture perturbation persists substantially longer than in their experiment (see section 4a).

The paper is organized as follows. We start by presenting the numerical model and its application in the experiments, as well as the employed observational datasets in section 2. Section 3 details the validation of the control simulation. The results of the sensitivity experiments are presented in section 4 with emphasis on soil moisture, temperature and precipitation response, circulation anomalies, and surface energy budgets. The main conclusions are discussed in section 5.

2. CHRM climate model and experimental setup

a. *Model details and origin*

The Climate High-Resolution Model (CHRM) version 2.3 (Vidale et al. 2003) is a state-of-the-art regional climate model (RCM) using a regular latitude–longitude grid ($0.5° \times 0.5°$) with a rotated pole and a hybrid

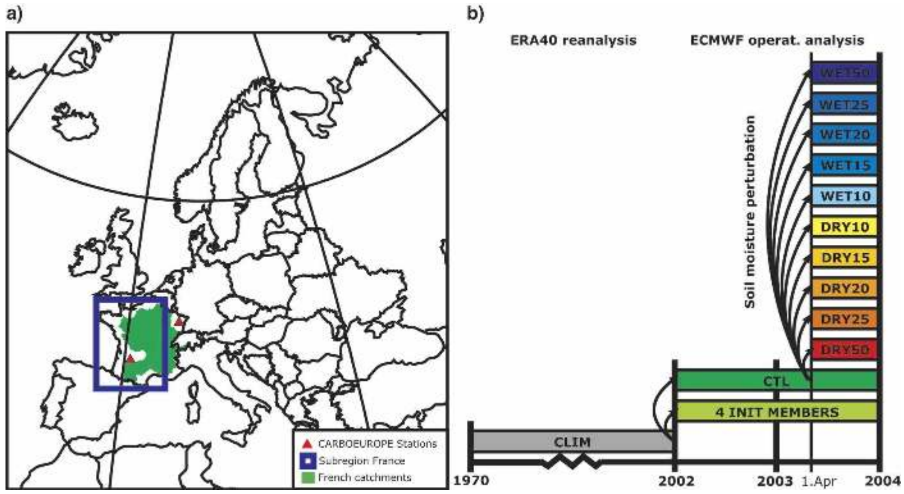

FIG. 1. (a) Map of the model domain used in all simulations. The land portion within the blue square is used for the analysis of area-averaged soil moisture, temperature, precipitation, and the surface energy budget. The red triangles mark the location of the two CARBOEUROPE stations, Le Bray (southwestern France) and Hesse (northeastern France). The French basin (encompassing the Rhone, Loire, Seine, and Garonne catchments) used for the validation of the terrestrial water storage cycle is highlighted in green. (b) Schematics of the sensitivity experiments. The lateral boundary conditions for the different periods are indicated at the top of the panel. On 1 Apr, the soil moisture of the sensitivity experiments is reduced/increased by up to ±50%.

sigma pressure coordinate with 20 vertical levels and 3 active soil layers (total depth of 1.7 m). The CHRM is the climate version of the former mesoscale weather forecasting model of the German and Swiss Meteorological Services known as the High-Resolution Model (HRM) or, formerly, the Europa-Modell (EM; Majewski 1991; Majewski and Schrodin 1994). The model has been modified by Lüthi et al. (1996) for application as a regional climate model. Its physical package includes a mass flux scheme for convection (Tiedtke 1989), Kessler-type microphysics (Kessler 1969; Lin et al. 1983), a land surface scheme (Dickinson 1984), and a soil thermal model (Jacobsen and Heise 1982). The soil moisture storage capacity had to be increased to improve the simulation of land surface balances of heat and water in a realistic and sustainable fashion on time scales longer than in standard NWP (see Vidale et al. 2003). The soil moisture evolves freely after initialization of the RCM and is never corrected or nudged in the course of the simulation. Vertical diffusion and turbulent fluxes in the atmosphere are parameterized by a flux gradient approach of the Beljaars and Viterbo (1998) type in the surface layer and Mellor and Yamada (1974) type in the boundary layer. The CHRM model also includes a radiative transfer scheme developed by Ritter and Geleyn (1992). The CHRM has been validated regarding its ability to represent natural variability on different time scales (Vidale et al. 2003). Note that the model has not been tuned for an optimal representation of the 2003 heat wave. Earlier versions of the model have previously been used in a series of sensitivity and process studies about the role of soil moisture–precipitation feedbacks (Schär et al. 1999) and the influence of vegetation changes on the European climate (Heck et al. 2001). More recently the current model version has been used in climate scenario simulations (S04; Vidale et al. 2007; Seneviratne et al. 2006b).

b. *Experimental design*

The computational domain used in this study covers Europe and the northeastern Atlantic with a horizontal resolution of about 56 km (see map in Fig. 1a). The experiment includes 16 simulations, which are detailed in the following subsections (see also schematic in Fig. 1b).

1) CTL, INIT, AND CLIM

The conducted experiment includes a control simulation (CTL), an ensemble of four simulations with slightly perturbed initial conditions (INIT), and a climatological reference simulation covering the whole 40-yr ECMWF Re-Analysis (ERA-40) period (1958–2001). The CTL and INIT simulations (2002–03) are driven by assimilated lateral boundary conditions and sea surface temperatures from the ECMWF operational analysis (T_L511) using the relaxation boundary technique (Davies 1976). The CTL and INIT simulations are initialized on 5 consecutive dates, on 3 January 2002 (CTL) and on 1, 2, 4, and 5 January 2002 (INIT), respectively, and use virtually identical initial soil conditions.

As a reference period we use the 31-yr period (1970–2000) out of the long-term 44-yr CHRM simulation driven and initialized with ERA-40 boundary conditions. The simulated soil moisture fields at the end of this run (31 December 2001) are used for the soil moisture initialization of the CTL and the INIT simulations. This provides the best estimate for the soil moisture initialization field in January 2002 and ensures that it is within a typical interannual range of the CHRM soil moisture equilibrium.

2) DRY and WET sensitivity experiments

In addition to the CTL, INIT, and CLIM simulations, 10 sensitivity simulations with perturbed spring soil moisture conditions (DRY, WET; see Fig. 1b) are performed in this study. The simulations are initialized on 1 April 2003, with the corresponding fields from the CTL simulation but perturbed soil moisture fields. The initialization fields are altered by increasing or decreasing the soil moisture by 10%, 15%, 20%, 25%, or 50% in the total soil depth at every grid box over the model domain. In the WET experiments, the soil moisture reached the water holding capacity at some grid boxes after the perturbation. In this case the additional soil water went instantaneously into runoff and was removed from the grid boxes. After the perturbation, the soil moisture evolves freely until the simulations end in December 2003. All DRY and WET simulations are driven with the same lateral boundary conditions as the CTL and INIT simulations. The comparison of these perturbed simulations against CTL allows isolating the effect of the soil moisture on the 2003 summer climate over Europe.

Most analyses have been performed for regional averages over nine European subdomains. However, in order to keep this paper concise, we decided to present regionally averaged analyses for the subdomain of France only (outlined in Fig. 1a). This region was selected because of its high anomalies during the peak episode in early August and the availability of terrestrial water storage estimates and observational data for validation.

c. *Observational data*

1) GISTEMP

In this study, the National Aeronautics and Space Administration Goddard Institute for Space Studies (NASA GISS) Surface Temperature Analysis (GISTEMP; Hansen et al. 1999, 2001) is used to validate the simulated 2003 temperature anomalies compared to the climatology. GISTEMP is a global analysis of surface temperature using observational station data as input data. The analysis includes land surface and sea surface temperature ($1° \times 1°$) as anomalies with respect to a user-defined reference period, in our case 1970–2000.

2) GPCC

We use the Full Data Reanalysis Product (Rudolf et al. 1994, 2005) provided by the Global Precipitation Climatology Centre (GPCC) to validate simulated precipitation anomalies. The global dataset (resolution $0.5° \times 0.5°$) covers the period 1951–2004 and is based on quality-controlled in situ observations from a large number of rain gauge stations.

3) BSWB

The diagnostic basin-scale water balance (BSWB) approach (Seneviratne et al. 2004) allows one to assess monthly changes in terrestrial water storage (TWS) for large-scale catchments ($>10^5-10^6$ km^2) using estimates of atmospheric moisture flux convergence and changes in atmospheric moisture content from reanalysis data, and streamflow measurements within the area. TWS encompasses soil moisture, groundwater, snow, and surface water. Here we use estimates derived using moisture flux convergence from ERA-40 for 1970–2000 (Hirschi et al. 2006a) and from the ECMWF operational analysis for 2003 (Hirschi et al. 2006b), respectively. The discharge data are based on runoff measurements retrieved from the Global Runoff Data Centre (available online at http://grdc.bafg.de/). Seneviratne et al. (2004) and Hirschi et al. (2006a) demonstrate successful validation of this method for various river basins.

3. Validation of CTL simulation

Here we present a validation of the CTL simulation for 2003. The CHRM has been extensively validated regarding its ability to represent temperature variations on annual as well as on interannual time scales (Vidale et al. 2003, 2007). However, the representation of an extreme event such as summer 2003 is an additional challenge for a climate model. Here, we focus on the validation of surface temperature (seasonal and daily time scale), precipitation, and soil moisture.

a. *Daily and seasonal temperature*

Figure 2 displays the summer (JJA) 2003 temperature anomaly with respect to the reference period 1970–

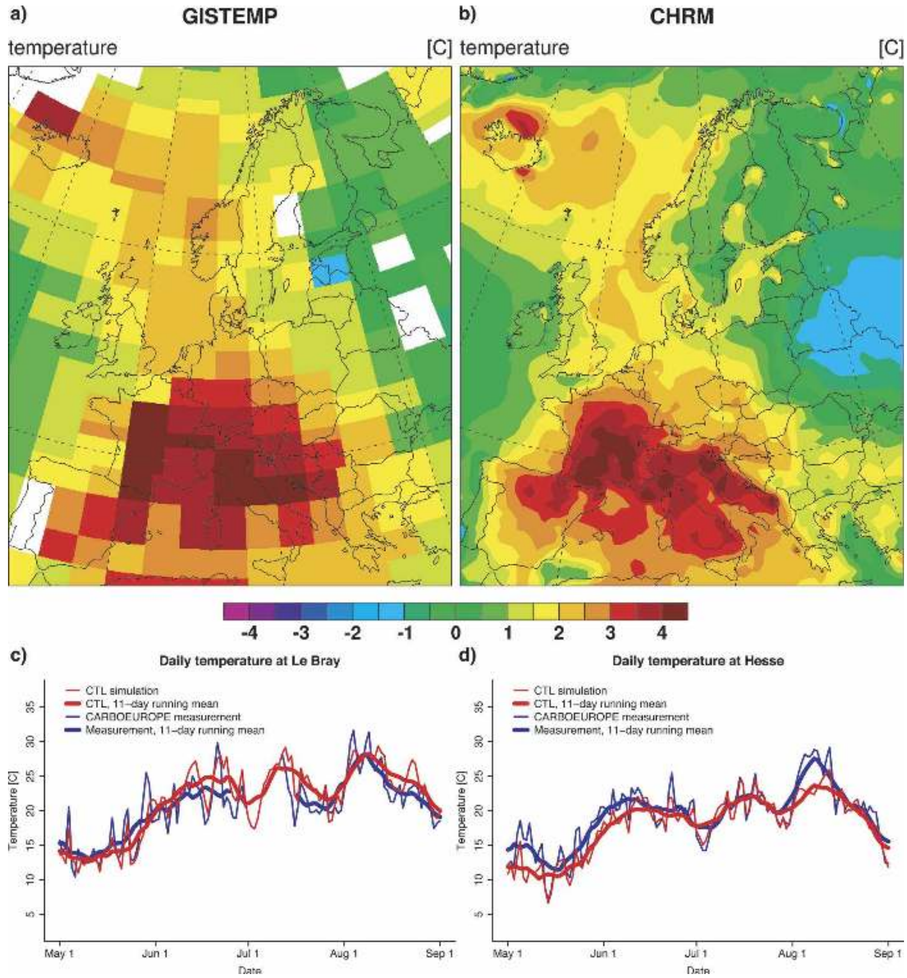

FIG. 2. Validation of surface temperature. Summer (JJA) temperature anomalies 2003 w.r.t. 1970–2000 (K) represented by (a) GISTEMP analysis and (b) the difference between CTL and CLIM simulations. (c),(d) Observed daily mean temperature (thin blue lines) and simulated temperature (CTL) (thin red line) at Le Bray and Hesse, respectively. The thick lines show an 11-day running mean of the corresponding datasets.

2000 represented by the GISTEMP dataset (Fig. 2a) and the CHRM CTL simulation (CTL − CLIM; Fig. 2b). The overall anomaly patterns compare well, despite a weak underestimation of the spatial extent of the warm anomaly over central Europe. The amplitude of the warm anomalies is in reasonable agreement except for the maximum heat anomaly over northern France, where it is slightly underestimated. On a monthly time scale (not shown), the agreement is best in June, whereas in July the strong anomaly over Scandinavia is somewhat underestimated. During the maximum heat wave in August the heat anomalies over France, the Alps, and northern Italy are underestimated by about 1°C in CTL. Apart from the weak underestimation of the extreme July and August temperatures, the monthly anomalies are well captured. Validation of the perturbed simulations reveals that the observed central European temperature extremes in July and August are better captured in the runs with spring soil moisture reduced by 10% (DRY10) and 15% (DRY15). This may indicate that the water stress and the related flux anomalies are somewhat underestimated in the CTL simulation. Good agreement between simulated (CTL − CLIM) and observed (GISTEMP) temperature anomalies is found in all months prior to the summer heat wave, from January to May, with biases similar to or smaller than in June (not shown).

The simulated temperature is also consistent with observations on shorter time scales. Figures 2c,d display simulated daily mean temperatures (CTL; inverse distance weighted average of the four nearest grid boxes) compared against observations at two CARBOEUROPE

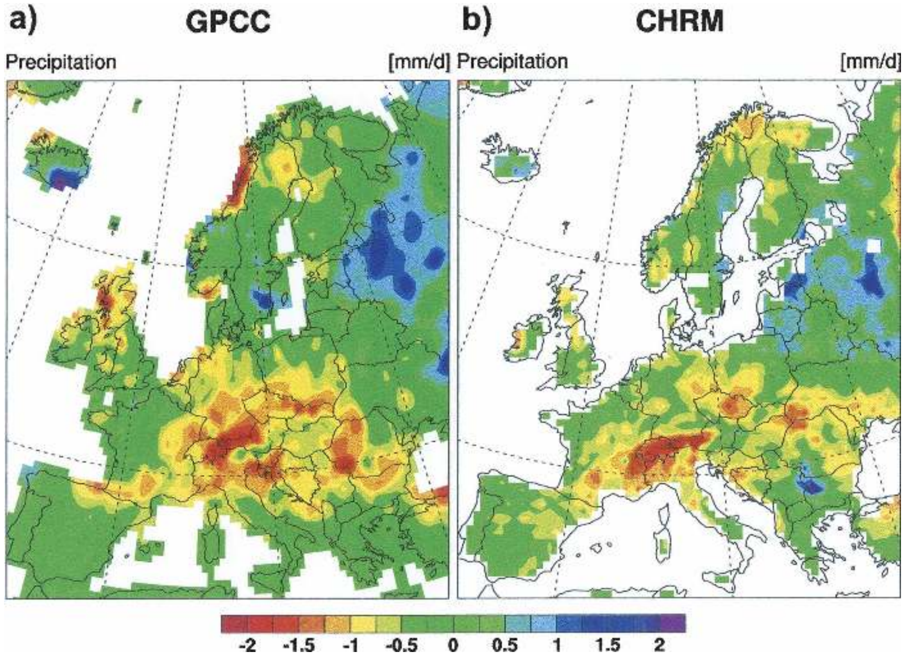

FIG. 3. Summer precipitation anomaly 2003 w.r.t. 1970–2000 represented by (a) the GPCC precipitation analyses and (b) the difference between CTL and CLIM simulation (mm day^{-1}).

stations in Le Bray (western France, coniferous forest) and in Hesse (northeastern France, hardwood forest; see map in Fig. 1a). At both stations, simulated daily temperature variations are in good agreement with the measurements. While the timing compares very well, the amplitude of the short-term variations is somewhat underestimated. This is mainly due to the smoothing effect of the averaging over grid boxes, which reduces the variability with respect to a single measurement station. For Hesse, the 11-day running average (thick line) shows local underestimation (around 3°C) of the maximum heat wave episode in the first half of August. The maximum warm anomalies in Le Bray are well captured.

b. *Precipitation anomaly fields*

European precipitation was below normal during the spring and summer of 2003 (Black et al. 2004; Zaitchik et al. 2006). We validate the model's representation of the pronounced precipitation deficit using GPCC observational data [see section 2c(2)]. Figure 3 displays the 2003 summer (JJA) precipitation anomaly relative to 1970–2000 for GPCC and CTL. There is good agreement on the overall anomaly pattern that shows strongly reduced summer precipitation over central Europe and anomalously wet conditions over northeastern Europe. Over Germany and eastern Europe, the lack of precipitation is somewhat underestimated by the model.

Note that during spring 2003 equally pronounced precipitation deficiencies are observed over France, the Alpine region, and southeastern Europe (not shown). This negative precipitation anomaly pattern is relatively well captured by the CTL simulation. Some parts of the Alpine region experienced persistent negative precipitation anomalies of 2 mm day^{-1} averaged over winter, spring, and summer 2003. The spring precipitation deficit has contributed to a strong soil moisture depletion, which will be addressed in the next section.

c. *Soil moisture cycle and terrestrial water storage variations*

The realistic representation of the seasonal soil moisture evolution is crucial for a simulation of the 2003 summer heat wave, as it strongly affects the partitioning of the surface energy into sensible and latent turbulent fluxes. In observational terms this involves large uncertainties due to the lack of a dense soil moisture observational network allowing for a spatially representative validation. Hence, we compare our simulations with basin-scale water balance [see section 2c(3)] estimates for the French catchments of Rhone, Loire, Seine, and Garonne (total area: 335 450 km^2; outlined in Fig. 1a).

The BSWB provides monthly rates of change of terrestrial water storage ($\Delta TWS/\Delta t$), which are compared against monthly changes of simulated snow and soil moisture aggregates. Figure 4a depicts the monthly TWS variations in 2003 (CTL, red; BSWB, blue) and

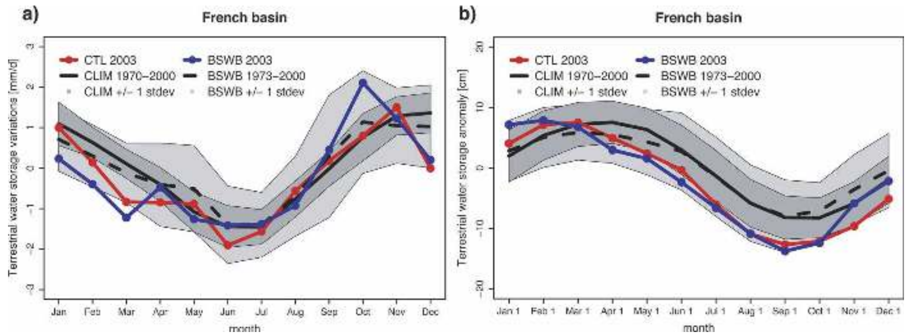

FIG. 4. (a) Monthly terrestrial water storage (TWS) variations averaged over the French basin (encompassing the Rhone, Loire, Seine, and Garonne catchments) diagnosed with the BSWB approach (see section 2c) for 2003 (blue line) for 1973–2000 (dashed black line; ±1 std dev marked; light gray shading). Corresponding values (sum of soil moisture and snow depth) for the CTL simulation for 2003 (red line) and 1970–2000 (CLIM; mean marked by solid black line; dark gray shading), (b) Same as (a), but absolute anomalies of TWS relative to the climatological annual mean.

the climatological mean (CLIM, solid line; BSWB, dashed line). An estimate of the time-integrated TWS anomalies is shown in Fig. 4b. Note that the BSWB provides no exact information on the absolute TWS values in 2003 with respect to the mean climatology but rather on the amplitude and phase of the annual cycle.

The climatological mean TWS variations (Fig. 4a) and time-integrated TWS anomalies (Fig. 4b) compare reasonably well between the simulations and the BSWB data. The TWS increase is somewhat overestimated in winter and underestimated in autumn. This results in a slight time shift and a weak overestimation of the climatological TWS cycle by the model (Fig. 4b). The interannual variability is underestimated by the model. This systematic deficiency has been demonstrated to be a general problem for a wide range of RCMs (Hirschi et al. 2007).

In 2003 the largest differences between the CTL simulation and BSWB are found in the rates of change in January and October. However, both datasets agree on anomalously strong TWS depletion in February and March 2003. As indicated above, the rapid decrease in soil moisture is associated with persistent negative precipitation anomalies (section 3b). This results in strongly reduced TWS from spring to autumn (Fig. 4b). The TWS variations (Fig. 4a) reveal only weak reductions in late spring and around average variations in late summer, both according to BSWB and CTL. Hence, after an anomalous depletion in early spring 2003 the TWS remained at a constant anomaly below the climatological mean until autumn (Fig. 4b). In the following section, we will assess the influence of these soil moisture anomalies in spring 2003 on the evolution of surface variables in the following months.

4. Sensitivity experiment

a. Soil moisture evolution

In this section we analyze the effect of the spring soil moisture perturbations applied in the DRY and WET simulations on the subsequent soil moisture evolution. The soil saturation at the time of the perturbation (1 April 2003) is shown in Fig. 5a in terms of the soil moisture index (SMI; two uppermost soil layers with 10- and 25-cm depths, respectively). The soil saturation is found to be relatively high except for mountainous regions in the Alps and Norway.

The soil moisture evolution before and after the perturbation is shown in Fig. 5b for the example of the subdomain France (FR; outlined in Fig. 1b). The DRY and WET experiments are not symmetric, since at some grid boxes the water holding capacity is reached after the soil moisture increase in the WET experiments. In these cases the soil moisture was instantaneously removed from the grid box through runoff.

The differences due to the slightly differing atmospheric initial conditions between the CTL and the four INIT members are revealed to be very small (maximum JJA range = 2.4 mm) compared to the interannual variability of CLIM (σ_{JJA} = 31 mm). In early 2003, the simulated soil moisture content (CTL and INIT) is relatively high due to anomalous precipitation amounts in summer and autumn 2002.

Independent of the soil moisture content in April, all perturbed and nonperturbed simulations show dramatic reductions of soil moisture from April to late August. This highlights the important role of the strong radiative anomalies, the lack of precipitation, and early activation of vegetation, which resulted in a strong des-

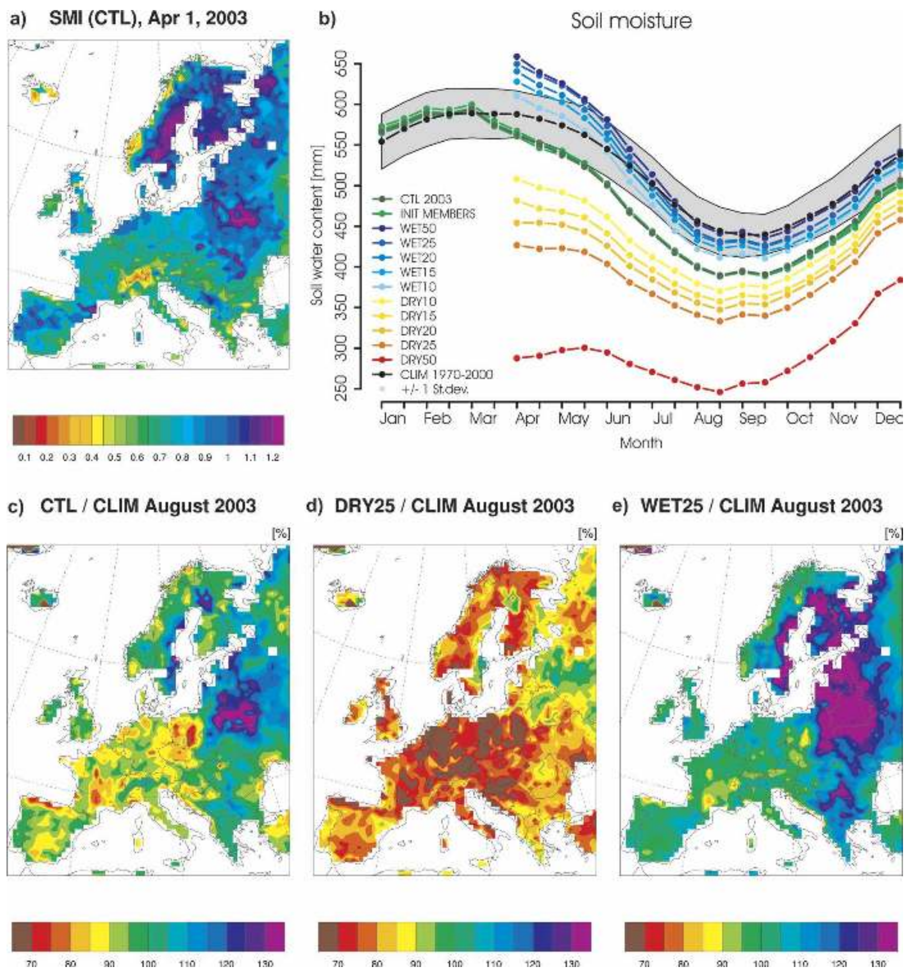

FIG. 5. (a) Soil moisture index (two uppermost soil layers with 10- and 25-cm depth, respectively) in the CTL simulation at the time step of perturbation (1 Apr 2003). (b) Semimonthly 2003 soil moisture depth averaged over France for the CTL simulation (dark green line), the four INIT members (light green lines), and the DRY (yellow to red lines) and the WET (bluish lines) simulations. The black line indicates the mean values for the period 1970–2000 (CLIM) and the gray shading indicates the typical interannual range ($\pm$1 std dev). Each tick mark on the abscissa represents the average over half of a month (e.g., the 1–15 Jan and 16–31 Jan). (c)–(e) Aug 2003 soil moisture content in the CTL, DRY25, and WET25 simulations, respectively, divided by the climatological mean (CLIM, 1970–2000) for the corresponding month.

iccation of the soil. In August 2003 the soils in all regions except for northeastern Europe and Scandinavia are anomalously dry in the CTL simulation (Fig. 5c) and very dry over almost the entire domain in DRY25 (Fig. 5d). Even in WET25 (Fig. 5e) and WET50 (not shown), the soil moisture values over parts of western and central Europe drop slightly below the long-term mean. Hence these two wet simulations represent a hypothetical 2003 climate with summer soil moisture conditions close to the climatological mean over France (Fig. 5b), as well as over central Europe and east toward Poland and Hungary (Fig. 5e). Note that the lateral boundary conditions are identical for CTL, WET, and DRY simulations. The difference between the WET25/WET50 and the CTL simulation provides a rough quantification of the soil moisture effect upon the 2003 summer heat wave for France as well as central Europe.

Over all regions, the imposed soil moisture anomalies in the DRY and WET simulations decrease continuously over time and approach the CTL in a quasi-exponential fashion. The characteristic half-life period varies between 3 and 8 months depending on the hydraulic conductivity of the soil and the saturation at the time of the perturbation. In FR (Fig. 5a), the disturbance is halved within the first 3.0–3.5 months (until

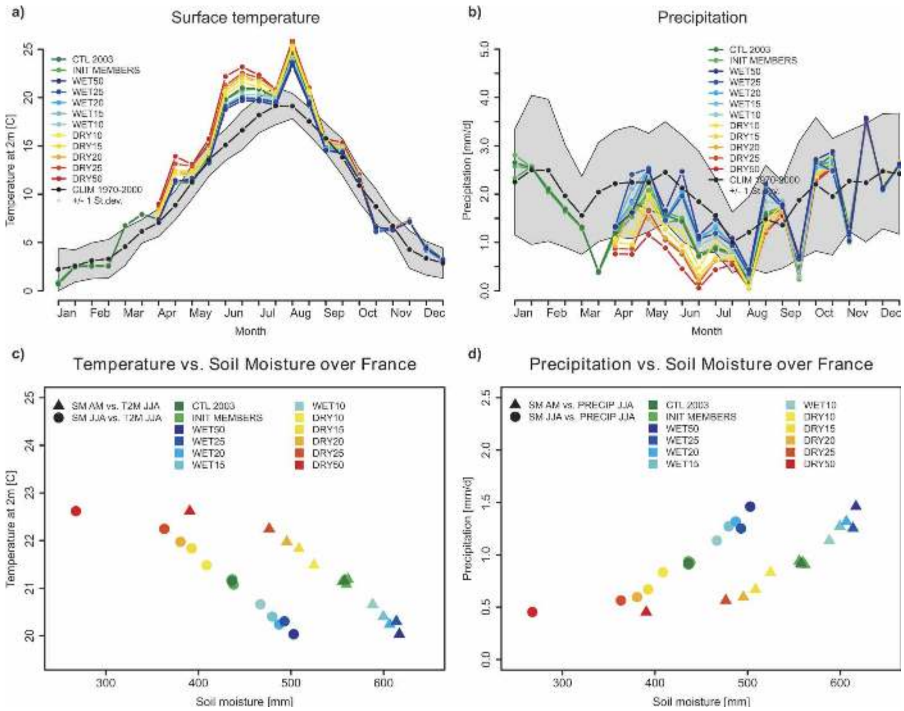

FIG. 6. (a) Semimonthly 2003 temperature at 2 m averaged over France (conventions as in Fig. 5b). (b) Same as (a), but for precipitation in mm day^{-1}. (c) Scatterplot between JJA 2003 surface temperature and soil moisture in the spring (April–May, triangles) and subsequent summer (JJA, circles) for all simulations. (d) Same as (c), but for precipitation and soil moisture.

mid July) of the WET runs and within 4 months (until August) of the DRY runs, respectively. The decay of the imposed perturbation is slowest and soil moisture memory longest in the simulations with small perturbations. This compares well with previous studies (Koster and Suarez 2001; Seneviratne et al. 2006a) suggesting that the sensitivity of runoff with respect to soil moisture increases with higher soil moisture content, and the sensitivity of evapotranspiration with respect to soil moisture increases with lower soil moisture contents. Therefore, the longest soil moisture memory is expected at intermediate soil wetness values (Seneviratne et al. 2006a).

b. Temperature response

The imposed spring soil moisture perturbations substantially affect surface temperature during the subsequent months. To quantify this effect, surface temperatures in the perturbed, unperturbed, and climatological simulations are compared. Figure 6a shows semimonthly temperature averaged over the subdomain FR for 2003 (colored lines) and the mean model climatology (1970–2000). In general all perturbed and unperturbed simulations follow the same path in 2003, showing the characteristic warm anomalies in June and the first half of August. The summer temperatures correlate quasi-linearly (gradient ~ -0.015 K mm^{-1}) with the absolute soil moisture amount in summer or in the preceding spring (Fig. 6c). The range of this effect in our simulations is relatively large and amounts to around 2.5 K, which corresponds to more than 2 standard deviations of interannual summer temperature variability.

Note that the driest simulation (DRY50) is an exception, since the additional warming with respect to the DRY25 simulation is relatively small. Due to the extensive perturbation the soil saturation is close to the wilting point in these two simulations. Hence an additional soil moisture reduction does not further affect the latent cooling and surface temperature.

The spatial temperature anomaly patterns of the DRY25 and WET25 simulations with respect to the unperturbed CTL simulation are displayed in Fig. 7. The reduction of spring soil moisture by 25% results in a strong summer (JJA) warm anomaly of about 2°C over a zonal band covering land regions between the Mediterranean and the North Sea and the Atlantic and the Black Sea. The effect is much weaker over the Iberian Peninsula and northern Europe. The anomaly DRY25 − CTL is caused solely by reduced spring soil

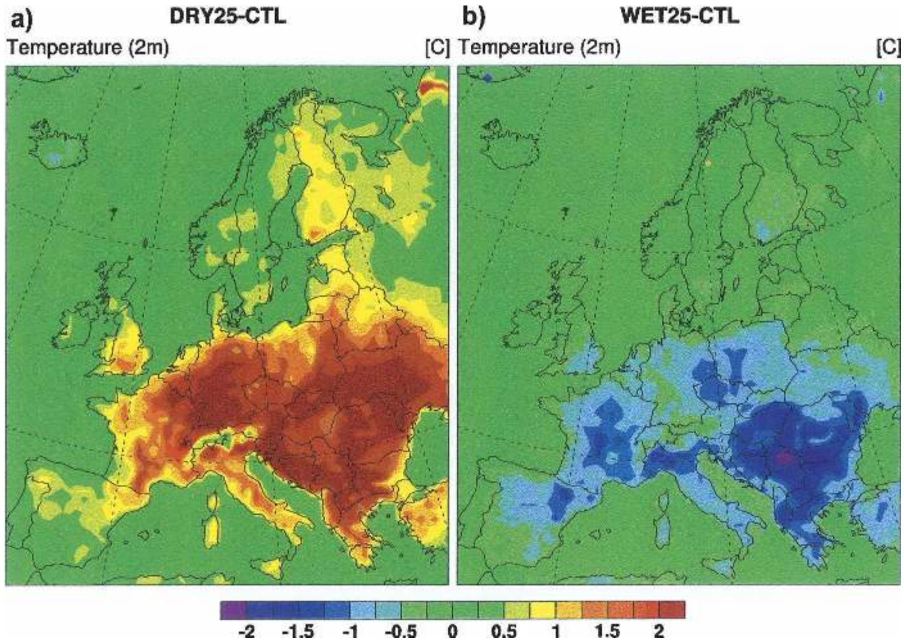

FIG. 7. Summer 2003 temperature anomaly due to spring soil moisture perturbation in (a) DRY25 − CTL and (b) WET25 − CTL.

moisture and the subsequent feedback processes and corresponds to about 30%–50% of the 2003 warm anomalies (CTL − CLIM) over France and large parts of central Europe (Fig. 2b). Hence, if soil moisture had already been at low levels in early 2003, the anomaly DRY25 − CTL would have superimposed upon the CTL − CLIM, leading to a strongly enhanced heat anomaly. The soil moisture effect on surface temperature occurs mainly through reduced latent cooling and to a lesser extent through a positive circulation feedback to be discussed in section 4d.

Increased spring soil moisture (WET25 − CTL) produces an opposite summer temperature signal and results in a reduction of the summer temperature anomaly (Fig. 7b). Again the strongest anomalies are found in a zonal band between the Mediterranean and the North Sea. Over large areas, the spring soil moisture increase in WET25 implies negative temperature departures of around −1.5°C with respect to CTL (WET50 − CTL; even more than −2°C over some regions; not shown). This signal represents a substantial portion of the 2003 summer temperature anomaly in the unperturbed simulation (CTL − CLIM). These findings suggest that 2003 summer temperatures over France and parts of central Europe would have been substantially cooler (by up to 2°C) given the same large-scale circulation pattern but climatological mean summer soil moisture. Note that this finding may be model dependent and that the soil moisture anomaly has only been validated over France. Over northern Europe there are virtually no effects of the soil moisture increases.

These regional differences of the response are mainly related to the different sensitivities of the latent heat flux to changes in soil moisture. To analyze these sensitivities we have calculated the percentage of net radiation going into latent heat flux for different simulations and different regions (not shown). This analysis revealed different regional sensitivities depending on the soil saturation. The sensitivity is low at dry (near wilting point) and at wet (near field capacity) soil moisture contents and strong at the intermediate contents. Over Scandinavia evaporation is not limited due to the high soil saturation. France and central–eastern Europe show high sensitivities for all simulations with intermediate soil moisture values. The two driest simulations (DRY25 and DRY50) show weaker sensitivity to further drying, since the maximum limitation is reached at many grid points (wilting point). Likewise the sensitivity is very weak over the generally dry Iberian Peninsula, explaining the small temperature response to soil moisture perturbations.

In a recent study, Ferranti and Viterbo (2006) analyzed the effect of soil moisture initial conditions on 2003 summer temperatures. They initialized the ECMWF atmospheric model with different May soil moisture conditions and performed seasonal forecasts for summer 2003. They did not prescribe the large-scale circulation but nevertheless found a significant temperature response, which shows the same overall anomaly pattern as in our simulations, albeit shifted somewhat to the south with maximum amplitude over the Mediterranean and the Iberian Peninsula.

c. *Precipitation response*

Previous studies (e.g., Rowntree and Bolton 1983; Schär et al. 1999) have shown that European summer precipitation is sensitive to soil moisture conditions. Schär et al. (1999) have identified in numerical experiments that the effect of soil moisture conditions on precipitation is in relative terms strongest in a belt stretching from the Atlantic to the Black Sea to the north of the Mediterranean.

Here, we present a brief analysis (confined to the subdomain FR) of the soil moisture–precipitation feedback. The spring soil moisture perturbations are revealed to have a substantial effect on precipitation during the subsequent months until the end of summer over France (Fig. 6b) as well as over the rest of the domain except for northeastern Europe and northern Scandinavia (not shown). Generally lower spring soil moisture leads to lower summer precipitation.

The precipitation response to the soil moisture perturbation is less continuous than for temperature. Over France the soil moisture–precipitation relationship appears relatively nonlinear, particularly in the DRY simulations (Fig. 6d). In the two driest simulations evaporation is strongly limited, and a further drying has only a minor effect on precipitation. A similar relationship applies to all other analyzed western, central European regions and the Iberian Peninsula (not shown). Over northern Europe, soil moisture is close to saturation so that evaporation and precipitation are not affected (see also Rowell and Jones 2006).

The soil moisture effect on precipitation mainly occurs through a limitation of summer evaporation, which reduces the convective activity. Schär et al. (1999) demonstrated by means of detailed budget analyses that the feedback cannot be interpreted in terms of precipitation recycling, since recycling is far too inefficient on the spatial scales that matter in the European context. They suggested that the surplus of precipitation over wet soils is derived primarily from atmospheric advection and is triggered by more efficient convective precipitation processes over wet soils. Here we do not perform budget analyses to distinguish between advected and recycled precipitation as well as local and nonlocal soil moisture effects. However, we expect the same mechanisms to apply as described in Schär et al. (1999).

d. *Circulation response*

The persistent strong anticyclonic circulation anomaly over central Europe played a crucial role during the 2003 summer heat wave (Black et al. 2004; Ogi et al. 2005). The 500-hPa height anomaly (Fig. 8a) is well captured by the CTL simulation and compares well with the National Centers for Environmental Prediction (NCEP) reanalysis and ECMWF operational analysis (Grazzini et al. 2003). The 500-hPa anomaly fields (CTL–CLIM) show a characteristic wave pattern with negative anomalies over the eastern North Atlantic and western Russia and pronounced positive anomalies northwest of Scandinavia and central Europe. The 1000-hPa height anomaly (Fig. 8d) is much weaker. In particular, over central Europe the positive anomaly is hardly detectable at the surface ($<$10 m). This is in good agreement with ECMWF operational analysis and NCEP reanalysis (not shown). We suggest that the relatively weak 1000-hPa height anomaly is partly due to the strong surface heating causing a weak surface heat low as described below.

The analysis of the sensitivity experiment reveals that soil moisture affects geopotential height from the surface to the upper troposphere. Figure 8 shows the 500- (Figs. 8b,c) and the 1000-hPa (Figs. 8e,f) geopotential height anomaly caused by the spring soil moisture reduction (DRY25–CTL; Figs. 8b,e) and increase (WET25 − CTL; Figs. 8c,f). In DRY25 the 1000-hPa height is found to be substantially reduced by up to 12 m. The shape of the anomaly corresponds to the associated surface warm anomaly depicted in Fig. 7a, pointing at the presence of a heat low mechanism. The same effect as in DRY25, but opposite in sign, causes a positive 1000-hPa height departure in the WET25 simulation (Fig. 8f).

The effect of soil moisture on 500-hPa geopotential height is reverse and of similar amplitude. The dry anomaly in DRY25 forces a positive 500-hPa height anomaly aloft, situated roughly above the same region covered by the surface heat low. The amplitude of this positive height departure (DRY25 − CTL), induced by the soil moisture perturbations, corresponds to about 10%–15% of the total CTL − CLIM anomaly.

Qualitatively consistent results are found for the other sensitivity experiments. These findings suggest that a moderate soil moisture reduction may enhance a positive height anomaly at upper levels and thus make it more persistent. Through this mechanism a dry soil anomaly should positively feed back on itself by forcing a strengthening of the anticyclonic circulation and an associated slight northward shift of the storm tracks.

The height anomaly DRY25 − CTL has its maximum

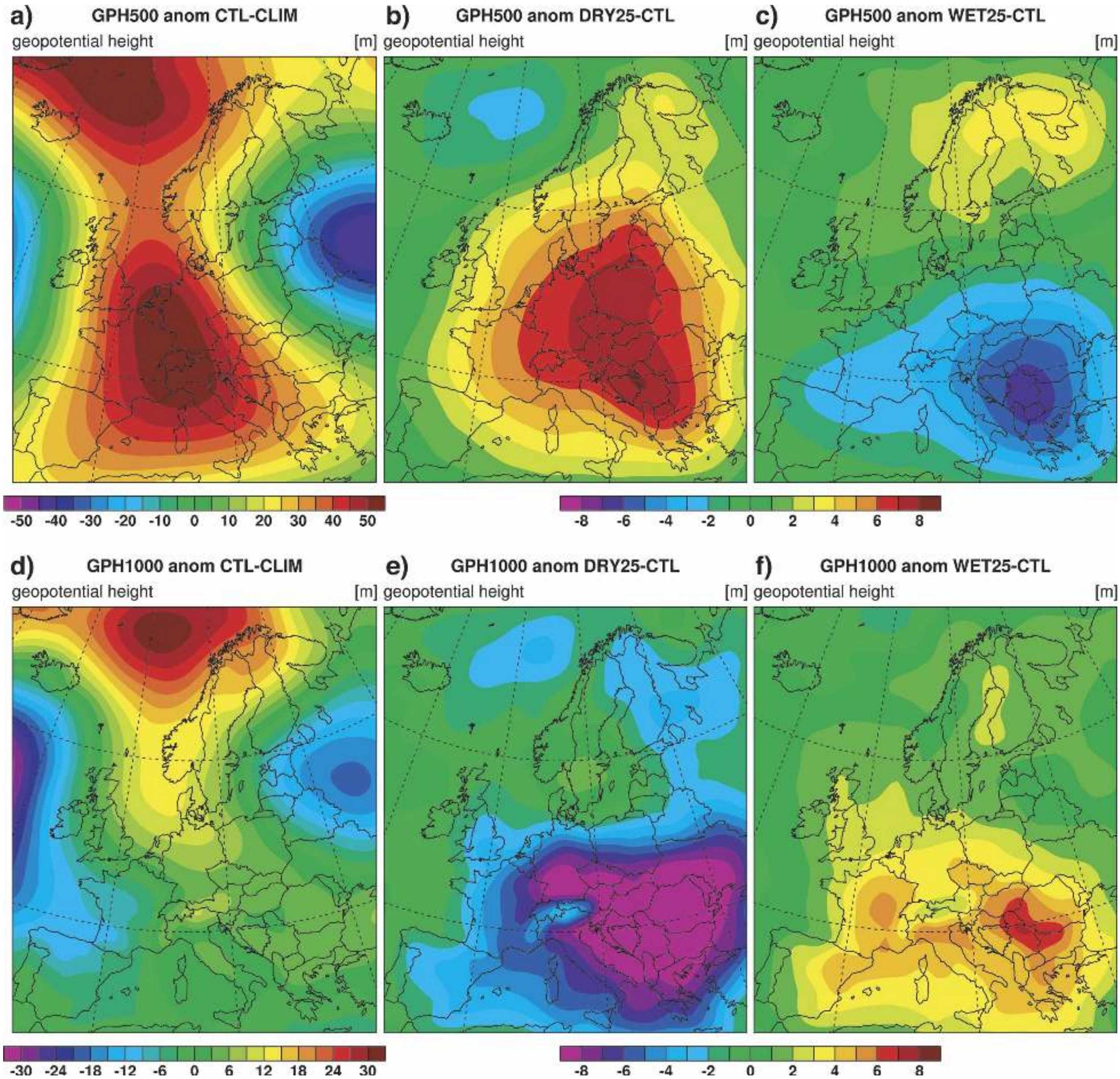

FIG. 8. Summer 2003 geopotential height anomalies CTL − CLIM at (a) 500 and (d) 1000 hPa, respectively, w.r.t. 1970–2000 mean. (b), (e) Same as (a), (d), but for DRY25 − CTL spring soil moisture perturbation. (c), (f) Same as (b), (e), but for WET25 − CTL spring soil moisture perturbation. Note that the scales are different in (a), (d) and (b)–(f).

slightly east of the CTL − CLIM anomaly. Hence, the dry anomaly adds a slight eastward component and may imply a weak change in advection to generally warmer air masses originating from a more eastern sector (i.e., the Mediterranean and continental northern Africa). The vertical height anomaly profile (not shown) over FR shows that the positive height anomaly is increasing in the upper troposphere, reaching a maximum anomaly of 20 m at 250 hPa. This corresponds to the level of the strongest 2003 height anomalies (NCEP and ECMWF; not shown). A comparable response of the 500- and 1000-hPa geopotential height to dry soils has been observed in the study Ferranti and Viterbo (2006). Previously, Oglesby and Erickson (1989) and Pal and Eltahir (2003) performed soil moisture sensitivity experiments over North America using different climate models and different experimental setups. Both studies found a heat low at the surface and an enhanced positive height anomaly aloft due to substantially reduced soil moisture. The increased 1000–500-hPa thickness is directly linked to higher tropospheric air temperature and an expanded atmospheric column in this layer. The

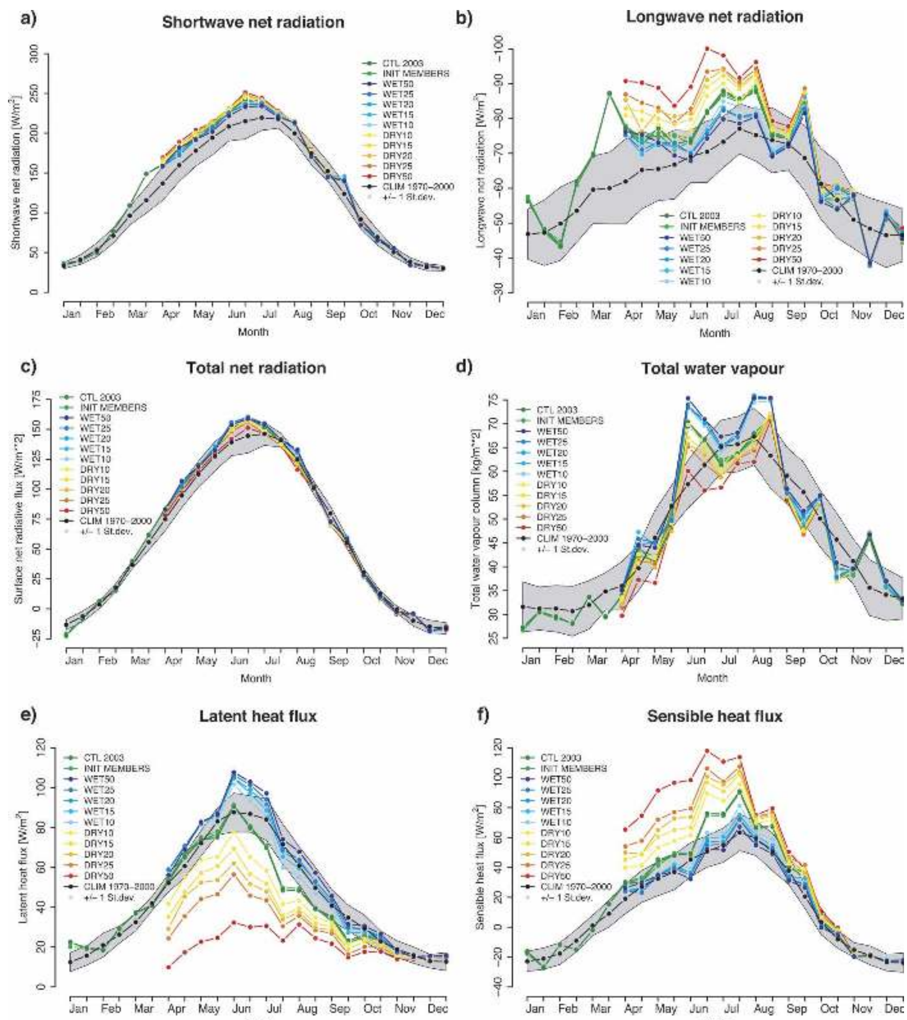

FIG. 9. Surface energy balance averaged over the land portion of domain FR (see Fig. 1a, conventions as in Fig. 5b): (a) Semimonthly 2003 surface shortwave net radiation, (b) longwave net radiation, (c) total net radiation, (d) total water vapor column, (e) latent heat flux, and (f) sensible heat flux.

tropospheric temperature increase due to drought conditions is strongly influenced by the enhanced sensible heat flux (reduced latent cooling). This will be discussed in the following subsection. Moreover, the reduced latent heat flux in the DRY simulation implies less diabatic (condensational) heating at midtropospheric levels. It is possible that this enhances subsidence through the mechanism described in Xue (1996).

e. Surface energy balance

Black et al. (2004) have shown that during the maximum heat wave from 6–12 August, air was essentially trapped within a strong anticyclonic block over northern France. Their analysis showed that air parcels traveled only very short distances, and they concluded that the regional heat budget must have enabled the extremely high temperatures to occur. To investigate the role of the surface energy budget we compare the simulated radiative and turbulent fluxes in 2003 (CTL) against the model climatology (CLIM). In a second step the effect of the soil moisture perturbations on the surface energy budget is evaluated by analyzing the DRY and WET simulations.

In the case of the sensible heat flux, diurnal fluxes are discussed separately; since daily averages are found to represent sums of reverse day and night signals, thereby obscuring important processes. Figures 9a–f display, respectively, semimonthly shortwave, longwave, and total net radiation, vertically integrated absolute water vapor as well as latent and sensible heat fluxes averaged over

FR for CTL, INIT, DRY, and WET simulations and the model climatology (CLIM). Note that for simplification positive-signed sensible and latent heat fluxes are defined to be directed upward.

1) Radiative and turbulent flux anomalies in 2003

In good agreement with Black et al. (2004) and Ferranti and Viterbo (2006), surface shortwave net radiation in the CTL simulation is found to be anomalously positive throughout the period of mid-February to mid-August 2003. On average the climatological mean (CLIM) is exceeded by 17 W m^{-2} during these 6 months. Maximum anomalies are reached in late March and June (around +30 W m^{-2}). Low integrated liquid cloud water contents (not shown) during these two periods point to anomalous clear-sky conditions. Ferranti and Viterbo (2006) show that cloudiness in the ECMWF analysis was reduced during the whole period between spring and early summer 2003. The hottest period over FR (in early August), however, was associated with somewhat smaller shortwave net radiation departures than the June and late March periods.

Outgoing longwave radiation over FR was anomalously strong from mid-February to August and again in late September due to the high surface temperatures. The average departure (February–August) of the CTL simulation from climatology (CLIM) amounts to 11 W m^{-2}. This is in good agreement with Moderate Resolution Imaging Spectroradiometer (MODIS) satellite measurements, which recorded an average summer anomaly of 12.6 W m^{-2} between April and September over France (Zaitchik et al. 2006). Interestingly, the strongest longwave net radiation anomalies are not found during the warmest periods in June and early August, but in late March. This is due to the compensating effect of anomalously strong longwave downward radiation (not shown), with values of 20 W m^{-2} (CTL − CLIM) during June and early August. This increase of downward radiation is caused by high absolute atmospheric water vapor content (Fig. 9d) and high tropospheric temperatures. Zaitchik et al. (2006) suggest that the reduction of longwave downward fluxes due to reduced cloudiness (not shown) was overpowered by the increase due to higher humidity and temperature.

The total net radiation (Fig. 9c) over FR was anomalously high (CTL − CLIM) between March and early July, when the shortwave downward dominated over the longwave upward radiation. We suggest that this extremely persistent total net radiation excess during all spring and early summer months strongly contributed to the depletion of soil moisture. Interestingly, however, simulated total net radiation returned to climatological mean conditions in late July and during the hottest phase in early August. This finding is in agreement with in situ radiation measurements at the Baseline Surface Radiation Network (BSRN) station in Payerne, Switzerland (Ohmura et al. 1998; not shown), which is the only BSRN station that is available for the period and region under consideration. A comparison with net radiation at three CARBOEUROPE stations in France reveals large local differences. While the daily net radiation during the first half of August 2003 was anomalously high in Le Bray, it was around the long-term average in Puechabon, France, and below the climatological mean in Hesse (not shown).

Ferranti and Viterbo (2006) and Black et al. (2004) identified positive total net radiation anomalies in the ECMWF operational analysis during August. They discuss some uncertainties due to the representation of low-level clouds (Ferranti and Viterbo 2006). Similarly, it is possible that our simulation underestimates the shortwave radiation at the surface due to the use of an old aerosol climatology that overestimates clear sky shortwave absorption (Hohenegger and Vidale 2005).

While the exact net radiation anomalies during early August involve uncertainties, it is interesting that in our simulations the hottest phase during the first two weeks of August is well represented despite the absence of an anomaly in total net radiation (Fig. 9c). This counterintuitive result indicates that the extreme August temperatures may be the result of the partitioning of the net radiation in latent and sensible heat flux, and not directly inflicted by net radiation anomalies.

The latent heat flux (Fig. 9e) in late summer 2003 was strongly constrained by the lack of soil moisture. In agreement with the analysis by Ferranti and Viterbo (2006), the CTL simulation shows latent heat flux values around the long-term average until mid-June. However, in the following months until October, the lack of soil moisture led to a dramatic reduction of latent heat flux. Averaged over JJA, the simulated latent flux anomaly (CTL − CLIM) over FR amounts to 11 W m^{-2}. This anomaly is somewhat stronger than the larger-scale European average anomaly derived from the ECMWF analysis (Black et al. 2004). However, the anomaly varies strongly in space and is highly dependent on the representation of the soil moisture content. Note that in the CHRM the vegetation cover is prescribed by a seasonally varying climatological mean leaf area index (LAI). Thus the vegetation cover is not directly affected by the dry conditions. However, through the dependence of the stomatal resistance on the soil moisture content and temperature, the transpiration rate may still be strongly sensitive to perturbations.

The simulated deficit of the latent heat flux in late summer 2003 is compensated by increased sensible heat flux (Fig. 9f). The respective anomaly amounts to 17 W m^{-2} in JJA (CTL − CLIM). This signal is found to be the sum of two opposite effects during day and night (not shown). During daytime the upward sensible heat flux is strongly enhanced, whereas during nighttime the downward directed sensible heat flux is enhanced. Analysis of the vertical temperature profile indicates that a nighttime temperature inversion develops in the boundary layer, resulting in downward sensible heat flux during the night and reduced upward sensible heat flux in the early morning. The simulated temperature inversion develops due to nighttime radiative cooling. In our simulation this phenomenon may be overestimated during the first half of August, when it contributes to unrealistically small daily mean sensible heat flux departures in contrast to previous studies by Black et al. (2004) and Zaitchik et al. (2006).

2) Effect of soil moisture perturbations on the surface energy balance

Here we evaluate the sensitivity of the different fluxes to soil moisture perturbations by comparing the DRY and WET simulations (red to blue lines in Fig. 9). Shortwave net radiation in the DRY simulations is slightly increased (Fig. 9a) mainly through reduced liquid cloud water content. The shortwave response corresponds to roughly half of the soil moisture effect on longwave net radiation (Fig. 9b), which is directly related to the surface temperature. Hence, the longwave effect dominates over the shortwave net radiation effect, which means that total net radiation in summer is generally lower in the DRY than in the WET simulations. Longwave downward radiation on the other hand seems to be indifferent to soil moisture perturbation (not shown). We expect this to be the consequence of two compensating effects: reduction due to lower atmospheric water vapor in the DRY simulations, which is compensated for by the increase due to the higher tropospheric temperatures.

The soil moisture effect on the turbulent fluxes is very pronounced. Both the latent and sensible heat fluxes are highly sensitive to the soil moisture conditions. A spring soil moisture reduction of 10% (DRY10) produces latent heat flux departures from the CTL simulation of 10 W m^{-2}. Over FR, this roughly corresponds to the anomaly CTL − CLIM. This indicates that a moderate additional spring soil moisture reduction may substantially enhance the latent heat anomaly in the subsequent summer. Note that the DRY10 simulation produced more realistic maximum temperature anomalies than the CTL simulation (section 3a). The reverse effect due to wet perturbations may produce latent heat flux values around the climatological mean despite the presence of strong anticyclonic forcing. This strong sensitivity of latent heat fluxes to soil moisture is the main factor causing the strong surface temperature differences between the perturbed simulations and substantially contributed to the temperature extremes in late summer 2003.

5. Conclusions

Key processes and feedback mechanisms contributing to the intensity and persistence of the 2003 European summer heat wave have been analyzed by means of regional climate simulations. Through careful validation these simulations are found to provide a credible and coherent picture of the 2003 summer heat wave. From our analysis, it is evident that both the anomalous atmospheric circulation during the summer itself as well as the anomalously dry continental-scale soils have played important roles.

The dry soil moisture conditions resulted from a combination of factors. Soil drying began early in the year with a persistent precipitation deficit. Later, the loss of soil moisture accelerated in response to early vegetation activation in the months preceding the extreme summer event. Moreover, the persistent excess of shortwave and total net radiation due to exceptional clear sky conditions further amplified the drying through evapotranspiration. After the strong reduction, soil moisture remained at exceptionally dry conditions throughout summer and autumn 2003.

The prime objective of our study was to assess the role of spring soil moisture anomalies in the subsequent extreme summer. We find a string of evidence suggesting that the dry soil conditions and ensuing soil moisture dynamics were a key in the sequence of events that led to the record-breaking summer of 2003.

- In our control integration, the hottest phase of the summer in early August took place despite the absence of an anomaly in total net radiation. Thus, the partitioning of net radiation in latent and sensible heat fluxes, which are to a large extent controlled by soil moisture, has strongly contributed to the extreme August temperatures. The lack of soil moisture resulted in strongly reduced evapotranspiration and latent cooling, which was compensated for by enhanced sensible heat flux. This mechanism amplified the surface temperature anomalies, particularly during August 2003.
- Comparison against our 30-yr (reanalysis driven) model climatology demonstrates that the soil mois-

ture depletion over central Europe and France in early spring 2003 exceeded the multiyear average by far. Subsequently, soil moisture remained at extremely low levels until the end of the summer.

- Sensitivity experiments using perturbed spring soil moisture suggest that given climatological mean soil moisture conditions in summer and similar continental-scale circulation, the 2003 JJA surface temperature anomalies would have been reduced by around 40%. Thus in absence of soil moisture feedbacks, summer 2003 would still have been warm, but it would not have been such a devastating event as it turned out to be.
- In the hypothetical case of dry conditions present already in winter 2003, central Europe might have experienced even substantially warmer conditions. Our analysis clearly shows that this effect occurs through reduced latent cooling, which is compensated for by enhanced sensible heat flux.

It is important to stress that the spring soil moisture conditions and the summer atmospheric circulation may largely be seen as independent forcing factors. Nevertheless, our results suggest that soil moisture perturbations can affect continental-scale circulation and that there is a positive feedback between the two. The response of the atmospheric circulation to the (soil moisture induced) low-level heat source is that of a heat low at lower levels and an anticyclonic ridging in the upper troposphere. In summer 2003, this has likely amplified the (preexisting) anticyclonic circulation, which in turn should positively feed back on surface temperature and soil moisture conditions. According to our sensitivity experiments, the strength of this feedback may explain up to about 10 m in 500-hPa geopotential height. However, the strength of this estimate may be somewhat constrained by the lateral boundaries.

It should not be overlooked that several of the processes that participate in the soil moisture feedbacks are of small scale and, thus, must be parameterized at the employed numerical resolution. The parameterization of these processes must be considered uncertain. As a result, the strength of the simulated feedback mechanisms may be model dependent and should be interpreted with caution. Despite these uncertainties, the simulated seasonal and daily temperatures and terrestrial water storage variation show notable agreement with the validation datasets. This provides confidence that the simulations represent a realistic and consistent picture of the 2003 summer heat wave.

We have demonstrated that soil moisture may strongly amplify European temperature anomalies in an extreme summer such as in 2003. This result raises a series of questions. First, it would be interesting to analyze the role of the discussed feedbacks in a more typical summer, and more generally, the contribution of soil moisture dynamics to the seasonal cycle and the interannual variability. Second, it would be desirable to study the soil moisture effect in other previous extreme events, such as the 1976 heat wave over Great Britain and northern France. This may add to our understanding of the processes governing heat waves, which—in response to climate change—are expected to become more frequent, more intense, and longer lasting in the future.

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