The role of the land surface in the climate system

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Abstract

The role of the land surface in the climate system is illustrated, with a focus on recent experience at the European Centre for Medium-Range Weather Forecasts (ECMWF). Global energy and water budgets are discussed and compared with their counterparts over the ocean, highlighting physical mechanisms responsible for their differences. Time scales associated with the global hydrological budget are presented. Using field data and model results, soil moisture is shown to be responsible for modulating the surface-atmosphere interaction at a continental scale, on time scales ranging from the diurnal to the seasonal. After a brief review of the impact of land surface on weather, three ECMWF case studies are presented where a more realistic representation of land surface was crucial for the performance of the forecast system. They correspond, respectively, to the role of soil moisture in determining the position and intensity of the precipitation maximum in an extreme event of mid-latitudes summer, the role of albedo of the snow in the presence of forests in spring and the effect of soil water freezing as a thermal regulator of the surface in cold climates. Finally, the evolution of the systematic errors in the ECMWF forecasts of near surface temperature and humidity is presented over the last ten years; a clear signature of changes to the representation of land surface processes (and other physical processes affecting the energy and water fluxes at the surface) can be found on that record.

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1. Introduction

The climate system is influenced by the land surface at a variety of time and spatial scales. *Firstly*, the atmosphere is in direct contact with the surface, which acts as a source or sink for the atmospheric enthalpy and moisture (and as a sink for the atmospheric momentum), via the surface sensible heat flux and evaporation (and the surface stress). *Secondly*, the surface conditions act as a regulator for important feedback cycles in the climate system. *Thirdly*, the partitioning of net radiation at the surface into sensible and latent heat fluxes determines the soil wetness evolution, which acts as one of the forcings – or, at the very least, a modulator - of low frequency variability. In fact, after the sea surface temperature, soil wetness and snow mass are the most important "memory" mechanisms for time scales ranging from weeks to seasons. *Finally*, the surface energy fluxes determine to a large extent the surface weather variables, such as screen-level temperature, humidity, and wind speed and, to a lesser extent, low-level cloudiness and precipitation. Humankind lives in the lowest two-metres of the atmosphere and is directly affected by the atmospheric conditions at the near surface.

This paper illustrates the impact of the land surface in the climate system, with a focus on examples based on recent experience at the European Centre for Medium-Range Weather Forecasts (ECMWF). The primary purpose of the paper is to highlight typical weather regimes/ecosystems where the land surface is of relevance to the evolution of atmospheric variables, rather than presenting a systematic review of the multifaceted research on surface-atmosphere interactions.

The importance of land surface in numerical weather prediction (NWP) stems from a blend of practical considerations and basic physical principles. Those are not very different from the general arguments presented in the opening paragraph, but two key issues are of paramount importance for NWP. First and foremost, accurate forecasts of near surface weather parameters are requested by the NWP users' community. The quality of such products as the diurnal cycle of near surface air temperature and humidity, winds, low level cloudiness and precipitation is to a large extent determined by the physical realism of the model representation of the surface-atmosphere interactions. Secondly, remote sensing observations of the atmosphere have increased importance to characterise the state of the atmosphere over land. The sensors that are sensitive to the lower troposphere can only be used effectively in the presence of a good quality background field of skin temperature.

It is important to emphasise that, despite the fact that deterministic forecasts of NWP do not extend beyond two weeks, the model land surface controls much longer time scales. The soil variables in the model are initialised every 6 hours based on indirect, imperfect and very sparse observations (Douville *et al.* 2000). Data assimilation combines that information with a short-term forecast to get an optimal estimate of the land surface initial conditions. Cycling the creation of initial conditions is tantamount to an extended model integration nudged by observations and extends the memory of the surface variables beyond the forecast duration. The realism of the representation of land surface processes is crucial for handling correctly the *memory* properties, represented by soil moisture in subtropics and mid-latitudes spring and summer and by snow mass in high latitudes and mountainous areas.

Although the importance of the land surface in the climate system has been established with, among others, the pioneering work of Namias (1958), Budyko (1963; 1974 and references therein), Manabe (1969), Charney (1975) –see review by Garratt (1993) - the impact of land surface on weather was only recently recognised, as can be seen by the scarce literature available. In Section 2 some general remarks on the surface energy and water budget will be presented, while Section 3 will elaborate on the role of the land surface, with emphasis on the long time scales characteristic of soil water. In Section 4 some surface-atmosphere interaction mechanisms that are relevant for the atmospheric simulation at the regional to continental scales are reviewed, based on numerical or observational studies that are relevant for NWP. Section 5 will illustrate when and where the land surface can affect the weather, based on examples from recent experience at the European Centre for Medium-Range Weather Forecasts (ECMWF). A short summary is presented in Section 6.

2. Surface energy and water budget

Fig. 1 represents in a schematic form the energy and water balance at the land surface. The surface albedo, controlling the fraction of the incident shortwave radiation absorbed by the surface, depends on the soil and vegetation type and state and on the amount of snow present. The net longwave radiation, *LW*, depends also on properties of the land surface, namely the surface emissivity and the surface skin temperature. Since the net radiation flux (the sum of solar and longwave radiation) is downward, and because the land surface has a small thermal inertia, the sum of latent and sensible heat fluxes must be an upwards flux. Note that the surface latent heat flux, *LE*, in the energy budget (left panel) is equal to the latent heat, *L*, times the evaporation flux, *E*, in the water budget (right panel), indicating that the water is available at the surface in a condensed phase and is passed to the atmosphere in the vapour phase. In that process, the surface undergoes evaporative cooling.

As mentioned in the introduction, the partitioning of the energy available at the surface into latent and sensible heat depends crucially on the soil moisture. Vegetated covered surfaces have the ability to draw water from a depth of order 1 m (the root layer), while for bare ground only the water in the top few cm of soil contributes for evaporation. The latent and sensible heat fluxes (*LE* and *H*, respectively) play a different role for the atmosphere. Sensible heat at the bottom means energy immediately available to the atmosphere, and contributes to the heating and/or deepening of the planetary boundary layer (BL), that shallow portion of the atmosphere directly affected by the surface. The surface evaporation flux does not directly heat the atmosphere, but provides moisture to the BL or, in the case of deep convection, to the whole troposphere. In favourable conditions, that contributes to precipitation generation mechanisms, with the associated release of latent heat into the whole troposphere. It is clear that, when compared to the sensible heat flux, evaporation can indirectly lead to a very efficient transfer of energy affecting a much deeper atmospheric layer. For an entire atmospheric column, the net radiative cooling is balanced by energy involved in phase changes inside the column (condensation of water vapour and evaporation of rain) and sensible heat flux at the surface. Land surface processes affect directly or indirectly this balance.

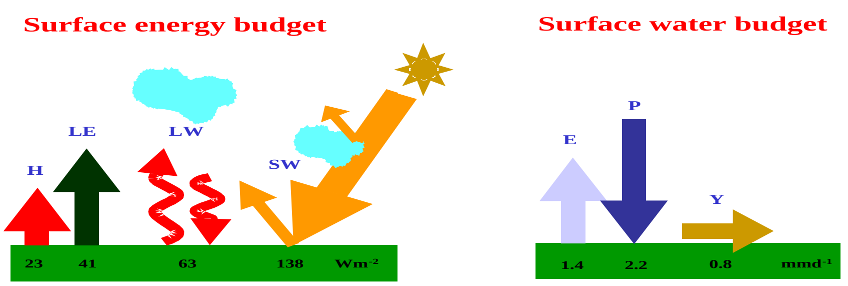


Figure 1. Schematics of surface energy and water fluxes. *H*, *LE*, *LW* and *SW* stand for surface sensible and latent heat flux, surface longwave and shortwave radiation flux, respectively; *E*, *P*, and *Y* stand for evaporation, precipitation and runoff. Numbers at the bottom represent averaged values over all land points for the ERA15 reanalysis (1979–1993).

Table 1. Mean surface energy fluxes in the ERA15 atmospheric reanalysis

	SW	*LW*	*H*	*LE*	G_0	*Bo=H/LE*
Land	138	-63	-23	-41	0	0.6
Sea	163	-51	-10	-104	-2	0.1

Table 1 summarises the surface annual mean fluxes for the 1979-1993 period covered by the 15 year European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis (ERA15, Gibson *et al.* 1997). Values presented are global averages over land and sea separately, in W m^{-2}, and downward fluxes are positive. The net heat flux, G_0, is the sum of all the surface fluxes. The contrasts between land and sea are clear. Even for such a large time period, the net flux is non-zero over sea, emphasising the larger thermal inertia of the oceans. The continents have a fast responsive surface and adjust their surface temperature in order to maintain a zero-heat flux at the surface, while the oceans have a much larger thermal inertia, with relatively small variations in surface temperature and flux imbalances allowed in much longer time scales. The last column, the Bowen ratio, *Bo*, is the ratio of sensible and latent surface heat fluxes. The larger values over land are indicative of the relatively difficulty of accessing the water at the surface. Over vegetated surfaces, this corresponds to the physiological mechanisms controlling transpiration while over bare ground the water directly accessible for evaporation is limited to the top few soil cm.

Surface energy fluxes for an entire season (not shown in the table above) still balance out, indicating that the energy associated to the seasonal changes in soil temperature are negligible when compared to the individual surface energy fluxes. However, for the surface water balance on a seasonal time scale, the storage term, or change in total soil water content, can be of similar magnitude to the precipitation or evaporation, which is self-evident in any extended drought period. In the next section, we will discuss the different time scales regulating the surface and regulated by the surface.

3. TIME SCALES AND THE ROLE OF SOIL MOISTURE

3.1 Global time scales

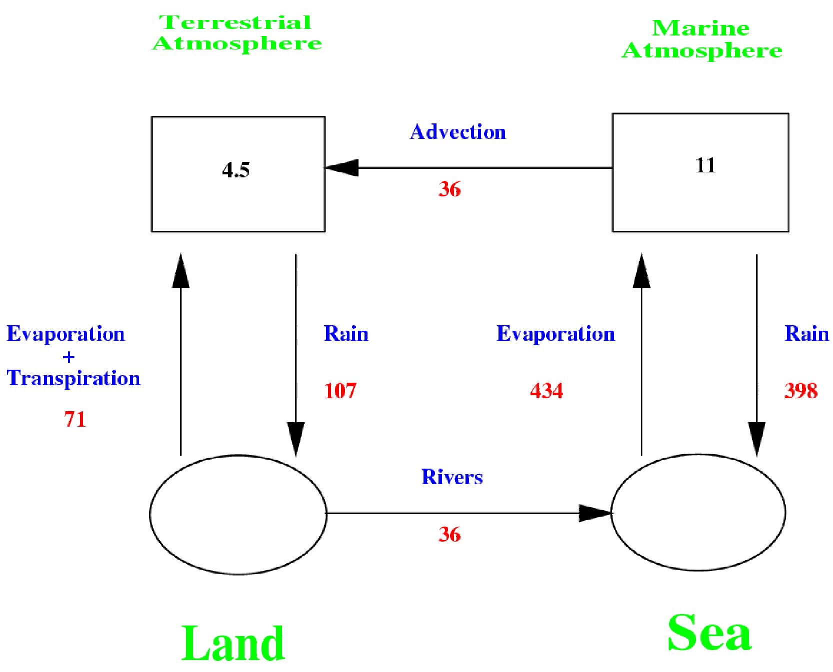


Figure 2. The global water cycle (from Chahine 1992). Units of water in reservoirs: 10^{15} kg; units of water in fluxes: 10^{15} kg yr^{-1}.

Fig. 2 represents the size of the moisture reservoirs of the terrestrial atmosphere and the marine atmosphere (rectangles in the figure), the exchanges of moisture between them, and between the atmosphere and the surface below (arrows in the picture). The sea surface evaporates at the potential rate, while over land there are additional mechanisms that reduce the evapotranspiration rate: dryness of the soil or, over vegetated areas, physiological mechanisms that can reduce or shut transpiration from the plant leaves and trunks making the water from the root zone effectively unavailable for the atmosphere above. Precipitation over land is about a quarter of that over sea. Note that precipitation exceeds evaporation over land, while over sea the reverse is true. In order to have a closed budget for the terrestrial atmosphere, advection of moisture across a vertical wall projecting over the continent boundaries has to match the difference precipitation minus evaporation. Advection is roughly half of the water evaporated over land (see Peixoto and Oort 1992 for estimates based on radiosonde observations), suggesting an annual recirculation ratio (ratio of the rainfall coming from local evaporation over total rainfall rate) of 67% (71/107). To close the hydrological cycle, the advection has to be matched by the river runoff: the global averaged influx of fresh water into the ocean is estimated in this way as 36 x 10^{15} kg yr^{-1}. For continental areas, annual runoff, evaporation and precipitation are approximately in the ratio 1:2:3.

Rigorous formulations of the atmospheric branch of the hydrological cycle can be found in Peixoto 1973 and Peixoto and Oort 1983, 1992. So-called "aerological estimates of runoff", based on measurements of atmospheric water vapour transport and a closure at the surface, have been applied successfully to basins with areas larger than 10^6 km^2 (Rasmusson 1967, 1968, 1971; Peixoto 1973).

The rainfall rate and the size of the reservoir can be combined to give a time scale 4.5/107 = 0.042 years = 15 days: the terrestrial atmosphere would be emptied by rainfall in 15 days. In a similar way the reservoir would be replenished by surface evapotranspiration in 23 days (4.5/71 years)[1]. The time scales associated to marine rainfall are only 7.5 days, and the corresponding value for evaporation is 6.8 days. This suggests (a) a more vigorous hydrological cycle over the ocean; (b) a land surface control over large time scales (weeks to months), through the evapotranspiration flux of water at the surface. The implication of the above on the extended predictability of the atmosphere due to exchanges of water with the land surface has been discussed by many authors (see, e.g., Namias 1958, Mintz 1984, Dümenil and Bengtsson 1993, Dirmeyer and Shukla 1993).

The accuracy of the numbers shown in Fig. 2 varies widely: see Chahine 1992, for the sources used to produce these particular estimates. Any literature review shows a very large dispersion in those numbers (see, e.g., Viterbo 1996): global estimates of the total column water vapour can vary by as much as 34%, while runoff estimates differ by 45%. Independent estimates of precipitation have smaller ranges of uncertainty (notwithstanding the extensive areas of the planet where observations are very scarce), but direct or indirect estimates of evaporation are subject to very large uncertainty.

3.2 The role of soil moisture

In order to illustrate the role of soil moisture in shaping the interaction surface-atmosphere, we will use model data over the Red-Arkansas River basin, a sub-basin of the Mississippi River basin. Data is based on the ECMWF reanalysis ERA15 (Gibson *et al.* 1997) used also in Section 2. Betts *et al.* (1998, 1999) have studied the ECMWF model energy and water budget over the Mississippi River basin, as described by short-term forecasts. Basin-average data was used for nine years, 1985–93 and analysed at different temporal scales; we will concentrate in this sectio on summertime 5-day average ECMWF data.

1. The global view on the atmospheric water budget presented in the text can be regionalised, allowing, under certain restrictive assumptions, an estimate of how much moisture that precipitates comes out from local evaporation versus horizontal transport. For a discussion on the "intensity of the hydrological cycle", i.e., the time scales associated to the emptying and repleneshing of the atmospheric water reservoir at different locations on the globe, see Trenberth (1999).

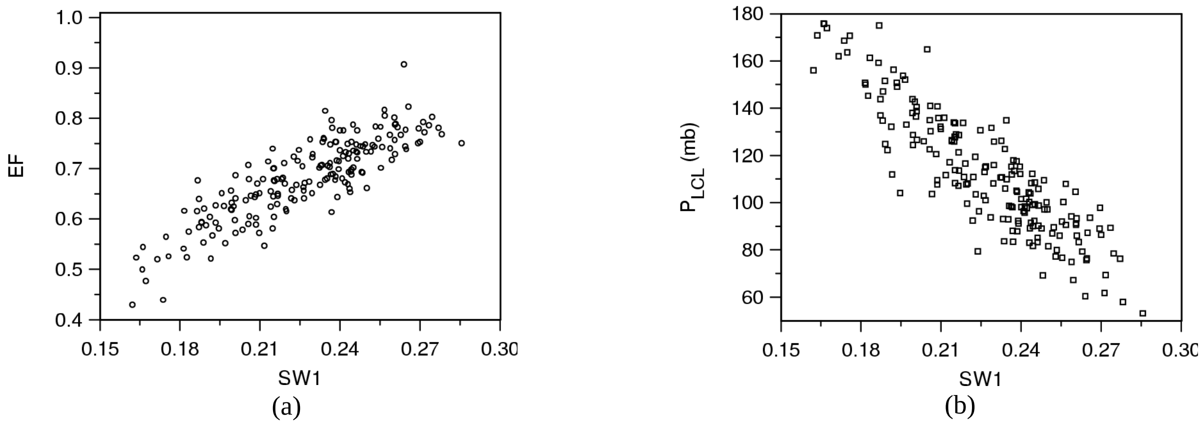


Figure 3. (a) Scatterplot of 5-day average evaporative fraction over warm soils (0-7 cm layer soil temperature > 296 K) against 0-7 cm soil water (SW1). (b) As Fig 3(a), but for pressure height to the lifting condensation level (P_{LCL}) (from Betts *et al.* 1998).

We will illustrate here that the ECMWF data shows a similar coupling in the midsummer on the 5-day timescale between soil water, evaporation and low-level thermodynamics as the First International Satellite Land Surface Climatology Project Field Experiment (FIFE) data[2] (Betts and Ball 1998) show on the diurnal time scale. We define a 5-day mean evaporative fraction as

$$EF = \frac{LE}{H + LE}.$$

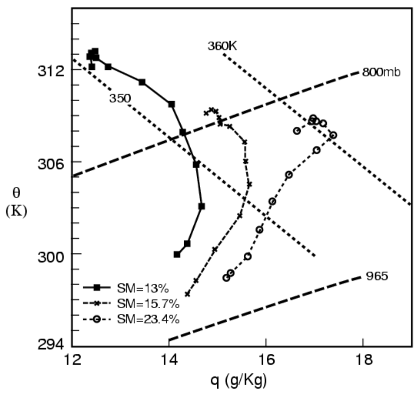
where H and LE are the 5-day averages of the surface sensible and latent heat fluxes. Fig. 3 (a) shows the strong coupling between EF and the top layer model soil water, SW1. The data plotted are all the 5-day values (1985-1993) for which the 0-7 cm mean soil temperatures are > 296 K, representative of the warm months.

Fig. 3 (b) shows a similar coupling of P_{LCL} (the pressure height of the lifting condensation level, determining cloud base) to SW1 for the same data. The model resistance to evaporation between the saturated interior of a "leaf" and the surrounding air is dependent on soil water, and this vegetation resistance is therefore one key factor in determining the equilibrium saturation level difference, P_{LCL}, in the saturation pressure budget of the BL (Betts and Ball 1998).

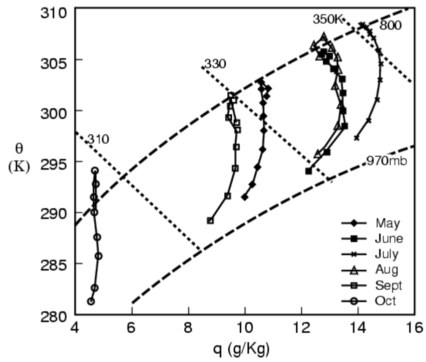
We have shown 5-day averages, but the patterns and slopes in Figs. 3 (a) and (b) are similar (but shifted slightly to higher P_{LCL} and lower EF) if the 12-h daytime average are used instead. Note that the lower limit in Fig. 3 (b) (corresponding to very wet soils) is near the oceanic equilibrium of $P_{\mathrm{LCL}} \approx 60$ hPa (e.g., Betts and Ridgway 1989). The oceanic surface boundary is saturated, and has no additional resistance of evaporation corresponding to the vegetative resistance over land. Figs. 3 (a) and (b) are particularly significant because neither of these relationships of EF and P_{LCL} on soil water depend strongly on soil temperature at this warm temperatures.

2. FIFE was an experiment measuring the different components of the surface energy and water budget of a 15 x 15 km^2 of natural grassland located in Kansas, US, and heavily instrumented over the period 1987-1989.

3.3 The interplay between the diurnal and the seasonal time scales



(a)



(b)

Figure 4. (a) Daytime diurnal cycle of potential temperature (q) and mixing ratio (q) at 2 m from 1145 to 2345 UTC for monthly dry-day composites (FIFE averages for 1987); (b) (q, q) plot of surface data for selected 28 days from July and August 1987, composited by soil moisture, showing the dependence of mean diurnal cycle on surface evaporation (from Betts *et al.* 1996).

Fig. 4 (a) (from Betts *et al.* 1996) shows the daytime diurnal cycle of the FIFE 2-m thermodynamic data for the predominantly sunny and dry days from May to October 1987. The axes are potential temperature (θ) and mixing ratio (q). This (θ, q) plot can be regarded as the heat and moisture budget on orthogonal axes (Betts 1992). There are 19, 21, 25, 22, 23, and 22 dates in each average from May to October. The selection criteria were near-noon surface net radiation above a threshold (which was 450 W m^{-2} in midsummer, falling to 300 W m^{-2} in October) and no significant daytime rainfall. Here we can see the diurnal and seasonal cycle together. The points are plotted hourly, starting at 1145 UTC, shortly after sunrise in midsummer. The seasonal rise and fall of mean temperature and mixing ratio can be seen: July is the warmest month. October is noticeably drier, after the vegetation has diedand evaporation is low. Saturation pressure lines of 970 and 800 hPa are shown dashed. The surface pressure is near 970 hPa. It can be seen that at the morning minimum temperature, the 2 m air is about 30 hPa from saturation, except in October, when it is more unsaturated. The diurnal range of mixing ratio q is relatively small in all months. There is generally a rise of q in the morning, when the BL is shallow and capped by relatively moist air from the BL of the preceding day, and a fall in the afternoon, as the growing BL entrains drier air from higher levels. May shows no afternoon fall of q, probably because of the higher soil moisture and evaporation. May and June do not reach as low afternoon saturation pressures as the later months of July, August, September, and October. This means a lower mean lifting condensation level (see also previous section) or cloud base in spring. Probably this reflects a seasonal drying of the surface, although changes in upper air thermodynamic structure may be involved. It is clear that the afternoon maximum of equivalent potential temperature θ_e is controlled mostly by the seasonal shift. The isopleths of θ_e =310, 330, 350 K are shown dotted. The rise of θ_e from morning minimum to afternoon maximum is around 14 K in all months.

The sum of surface sensible and latent heat fluxes is a surface source for increasing θ_e (e.g., Betts and Ball 1998). This surface θ_e is proportional to the sum of H+LE, and it is not affected by the Bowen ratio. It is entrainment of low θ_e air from above the BL, together with the deepening of the BL, that reduce the BL θ_e rise, and so feed back on both the shallow and even more importantly on precipitating convection. Thus one of the important aspects of the BL evolution over land is how large entrainment at BL top is. The daytime BL over land is primarily thermally generated (in strong winds, shear plays a role), and thus linked to the surface virtual heat flux (which over land is usually dominated by the sensible heat flux). Hence if the surface Bowen ratio is large, although the surface θ_e flux may be unchanged, the large H flux drives more entrainment, produces a deeper BL, and the diurnal rise of θ_e is reduced. Fig. 4 (b) shows how this diurnal cycle over land depends on soil moisture and, as a consequence, the surface evaporation.

A total of 28 days from July and August 1987 during FIFE, which were affected little by precipitation or cold air advection, were composited by soil moisture (*SM*: measured gravimetrically in the top 10 cm). The points are hourly values from 1115 UTC (near sunrise) to 2315 UTC. The dry soil composite (*SM* = 13%, for which the measured mean surface Bowen ratio at noon was 0.8) reaches a higher afternoon maximum $\theta_e \approx 352$ K (the θ_e isopleths are shown dotted). In contrast, the wet soil composite (SM = 23.4%), for which Bowen ratio at noon was 0.4), reaches a much cooler afternoon θ maximum, but a much wetter q value, so that the afternoon $\theta_e \approx 361$ K. Some of this shift of θ_e is associated with the shift of the entire diurnal path of higher q with higher soil moisture, but about half is the result of reduced entrainment of dry low θ_e into the BL. Over wet soils, H is much reduced and the BL deepens less rapidly. For all the three composites, the surface available fluxes (net radiation minus ground heat flux) were nearly identical, so that the surface θ_e fluxes were similar. This local feedback between soil moisture, evaporation and afternoon θ_e equilibrium probably produces on large spatial scales a positive feedback between soil moisture and precipitation, which has been the subject of much research (see, for example, Brubaker *et al.* 1993, Trenberth 1999). The analogy over the tropical oceans is the link between BL and sea surface temperature (SST), which influences the prevalence of deep convection over warm water. Over land variations in soil moisture can lead to as large variations in BL θ_e as larger as those produced by several degrees of change in SST. On continental scales, higher soil moisture and higher evaporation over land would lead to a higher afternoon θ_e maximum relative to the surrounding ocean and shift more of the global precipitation over the continents (Betts *et al.* 1996). This feedback has been seen in global models (Mintz 1984). On the regional scale, the Mississippi flood case study, presented in Section 5.1 below, suggested that the multiple BLs over the Midwestern United States controlled the *location* of precipitation, rather than this mechanism.

3.4 A schematic view of the role of land surface

To conclude this section, a schematic description of the interactions between the surface and the atmosphere will be presented. Inspired by an early, much more complex diagram by Horton (1931), Dooge (1992) (see also Kuhnel *et al.* (1991)) summarised the interaction between the land surface and the atmosphere in the picture reproduced (with adaptations) in Fig. 5 . The diagram illustrates the behaviour of the soil and the atmosphere within a complete cycle composed of a wet period followed by a dry period. Let us start just after a long episode of rainfall, point A in the figure. The soil water is available in abundance in the root layer[3] and its evolution is going to be determined by evaporation. While the soil has plenty of water, the rate of evaporation is controlled by the near-surface atmospheric moisture: the regime is controlled by the atmosphere and the evaporation is at the potential rate. Below a certain level of soil moisture (point B in the picture), physiological mechanisms will limit the supply of water from the root layer into the atmosphere, and the evaporation will drop below its maximum value (potential evaporation, E_{pot}). The regime is under a vegetation (soil) control. When precipitation starts (points C) it will meet a soil dry enough during the initial stages, so that infiltration (I_f, that part of water that falls as precipitation and is effectively collected by the soil for future use) will equal precipitation. The evolution of water in the soil is once more atmospheric controlled, via the rate of precipitation. Beyond a certain value (point D), the soil does not have the ability to soak all precipitation, some of it goes into runoff. This last phase is again soil controlled; the state of the soil determines the rate of infiltration.

Land surface parametrizations have to represent correctly the surface fluxes and the evolution of soil moisture in

3. For the purpose of this discussion, field capacity may be considered as the threshold beyond which there is a minimum canopy resistance to evaporation.

all four phases of the cycle, and to switch from the atmospheric control into the soil control regime. The evolution of soil moisture will determine when point D will occur, and the evaporation formulation will determine point B. The crucial areas, from the point of view of the atmosphere, are the quadrants BC and CD. During spring and summer (where the atmospheric demand can be very large), the system remains much longer in the half-circumference BCD than in the opposite part of the cycle.

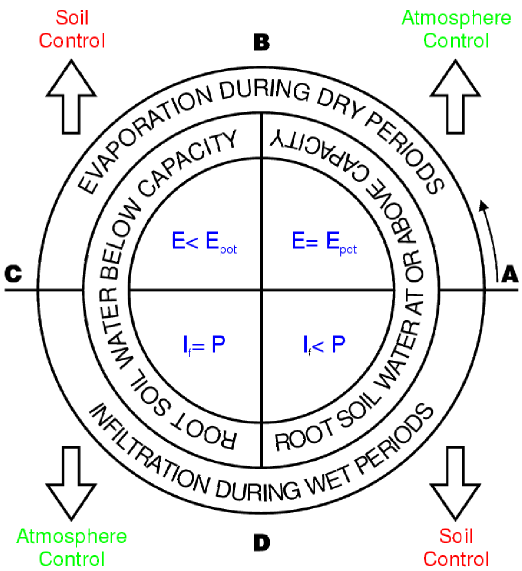


Figure 5. The hydrological rosette: Schematic depiction of the interaction between the soil hydrology and the atmosphere (adapted from Dooge (1992)). E_{pot} and I_f represent potential evaporation and infiltration, respectively.

4. Impact of land surface on weather: A brief literature survey

Betts *et al.* (1996) review the impact of land surface in the context of global numerical weather prediction: Diurnal and seasonal feedback loops are discussed as well as feedback loops controlling the BL evolution. We will highlight here typical mechanisms controlling the interaction between land surface and the atmosphere, over the US, Europe, and the tropics.

Early modelling efforts (Benjamin and Carlson 1986; Lanicci *et al.* 1987) have shown the sensitivity of precipitation in the US Great Plains to evaporation upstream, in the Mexican Plateau. The characteristic storm environment, leading to heavy precipitation over the Midwest, involves the breakdown of a capping inversion formed by an overlying pre-existing boundary layer from the Mexican plateau, which overlies the cool moist BL originating in the Gulf of Mexico. This complex pattern of differential advection impacts on the strength of the capping inversion, and the strength of the inversion is controlled by evaporation upstream of the precipitation area. Lower values of evaporation lead to a stronger capping inversion, and the low level flow from the Gulf will not break through the inversion until much further north. The location and extent of the heavy precipitation associated with the July 1993 US floods was found to be highly sensitive to the correct representation of these mechanisms in the ECMWF model (Beljaars *et al.* 1996; Viterbo and Betts 1999a; Section 5.1).

Spatial gradients of soil moisture can also enhance the differential heating maintaining and reinforcing the surface front in a pre-storm environment and intensifying the thermally direct (ageostrophic) circulation (Chang and Wetzel 1991; Fast and McCorcle 1991). A similar mechanism is active in a cold front associated with a severe squall line developing explosively (Koch *et al.* 1997).

The seasonality of leaf area index impacts on the systematic errors of US Midwest lower tropospheric temperature in summer (Xue *et al.* 1996; Yang *et al.* 1994). At a smaller scale, there are many observational and modelling studies demonstrating the importance of mesoscale fluxes and smaller-scale heterogeneity, specially at low wind speeds.

Over Europe, Rowntree and Bolton (1983) demonstrated the role of local and non-local response of medium-range rainfall forecasts to anomalies in the initial soil. The mechanisms relevant to this soil moisture-precipitation feedback were scrutinised in Schär *et al.* (1999) in a study demonstrating the impact of idealised (and large) anomalies of soil water on the European summer circulation. Unlike the desert-albedo feedback hypothesis (Charney 1975), the European soil-precipitation feedback is not of a large-scale dynamical nature, i.e., it is not associated to changes in large-scale flow. Precipitation recycling has also a small role in Europe. Three main feedback loops have been identified. Wet soils, associated with low Bowen ratios, lead to the build-up of a shallow BL, concentrating moist entropy at low levels and giving higher values of convective available potential energy (CAPE). Additionally, lower Bowen ratios lead to higher relative humidity, lowering the level of free convection. Finally, a positive feedback of radiative origin, with increased cloud cover, but larger net radiative flux, leads to larger moist entropy and convective instability.

All the examples above refer to extratropics spring and summer examples, in snow free situations. Section 5.2 below presents spring examples in the presence of snow. There is little or no impact of land surface on the atmospheric circulation in winter (see Giorgi (1990) and Section 5.3 below).

Despite an extensive list of publications on the role of land-surface in the tropical climate, there is scanty evidence of its impact for short- and medium-range forecasts. A notable exception is the work of Walker and Rowntree (1977) who demonstrated the role of enhanced soil moisture gradients on the short-range (1-2 days ahead) forecast of the generation of easterly waves, using a simplified model over West Africa.

5. Examples from ECMWF recent experience

5.1 Soil moisture

July 1993 showed anomalously high precipitation over the Central USA, with exceptional flooding of the Mississippi (Changnon 1996). During this month, the new version of the ECMWF model (CY48) and the then operational version (CY47)[4], were running in parallel at full resolution (spectral truncation T213, grid-point spacing ~ 60 km), including data assimilation. Beljaars *et al.* (1996) compared the performance of the two schemes, looking at the average of all one-, two- and three-day forecasts verifying between 9 and 25 July. While the day one precipitation of the two systems were very similar, and similar to the observed precipitation, the forecasts at day 3 were markedly different. In the new system, the location and intensity of the maximum precipitation was similar to the observations (40 N, 95 W), while the old system had less then half the precipitation amount in the area of the observed maximum and had a spurious maximum of precipitation displaced 800 km NE. In the old system, there was a gradual reduction in precipitation from day 1 to day 3, while the new system was able to better maintain the intensity. However, evaporation at the area of maximum precipitation was similar for the old and new system, and in both systems there was no evidence of forecast spin down, strongly suggesting that the local evaporation was not re-

4. The main differences between CY47 and CY48 rely on the surface and boundary layer processes parametrization. The soil model in CY48 has 4 layers, with no heat flux and free drainage bottom boundary condition, with soil properties (heat and water conductivities and diffusivities) dependent in a non-linear way on soil moisture (Clapp and Hornberger 1978). CY47 has 2 layers plus 1 climate layer underneath, with constant water soil properties. CY48 has a smaller roughness length for heat than for momentum, while in CY47 they are identical. For more details see Viterbo and Beljaars (1995).

sponsible for the differences in precipitation. It turns out that the maximum of the evaporation difference was located over the Mexican Plateau, 1000 km SW of the precipitation maximum, two to three days upstream as suggested by backwards trajectories ending up at 750 hPa, 40 N, 95 W. The mean thermodynamic profiles, similar for day 1 forecasts, were very different for day 3 forecasts. The old model showed a too strong capping inversion above the BL, with air too warm and too dry and much lower values of CAPE. It is clear that the differential advection mechanism characteristic of the US Monsoon was responsible for the differences in precipitation. When compared to the new model, the soil on the Mexican Plateau had much lower values of soil moisture in CY47, giving a much reduced evaporation, which in turn produced a warm and dry air mass that capped the BL downstream, inhibiting convection. In CY47, the soil model values were strongly forced to an erroneous, too dry, climatology: In such a data dense area, atmospheric profiles were initiated to correct values, but during the forecast they slowly felt the influence of the erroneous soil moisture values. In CY48 the soil moisture values were initialised to field capacity at the beginning of July, consistent with values of June precipitation in the area much above normal. There was no forcing to climatology in CY48 (Viterbo and Beljaars 1995) and the model was capable of maintaining high values of moisture throughout July. Monthly integrations performed with CY47 and CY48 suggested the importance of the memory associated to idealised soil moisture anomalies in the initial conditions (Beljaars *et al.* 1996). The monthly precipitation fields with CY48 compared much better to observations than those of CY47.

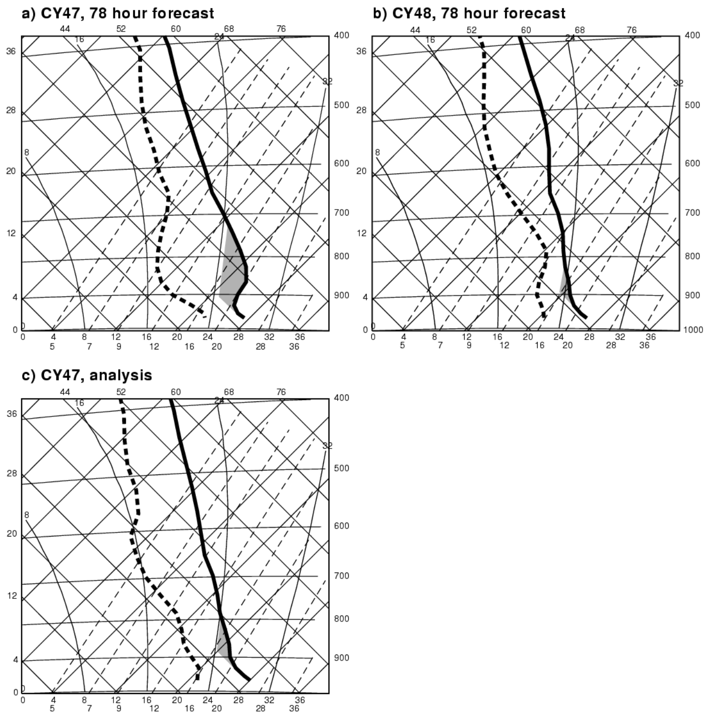


Figure 6. Profiles (so-called tefigrams) of temperature (solid line) and dewpoint (dashed line) with CY47 (a) and CY48 (b) of the averages verifying from 9 to 25 July at the forecast range 78 hours. The model location is 40 N 95 W. (c) shows the average verifying analysis. The shading indicates the area where a parcel lifted from the surface has a lower temperature than the surrounding air, i.e., the shaded surface area is a measure for the stability a parcel has to overcome before convection can occur (from Beljaars and Viterbo 1999).

CY48 surface model ran with predicted soil moisture throughout the first half of 1994. It was clear that a very large near-surface warm and dry bias developed over the NH continental areas at the end of spring and beginning of summer. A scheme to initialise soil water based on the short-term forecast errors of near-surface atmospheric humidity was developed to overcome that problem (Viterbo 1996). In order to test the new scheme, three complete data assimilation-forecast experiments were ran at T213 for the month of June 1994: (a) Control (CY48, no assimilation of soil moisture); (b) As control, but using the initialised soil moisture values, and; (c) As in (b), but using a prognostic cloud scheme with much more realistic cloud cover over land (Tiedtke 1993). The near-surface warm and dry bias, reduced from Control to the initialised soil water experiment, in response to a wetter soil. A lower tropospheric warm bias developed in the Control model and was greatly reduced when initialisation of soil water is used. Both experiments had too little cloud cover over land with too large surface shortwave radiative fluxes, but the wetter soil conditions of experiment (b) managed to maintain evaporation in the face of excessive net radiation at the surface. The third simulation displayed even smaller biases, associated with a larger, more realistic cloud cover and smaller radiative biases.

The algorithm to initialise soil water is successful in controlling model drifts but dampens the seasonal cycle and interannual variations of evaporation and soil moisture. Viterbo and Betts (1999a) revisited the July 1993 simulation, using the initial soil water algorithm and the prognostic cloud scheme. The new system gives poorer results for precipitation than Beljaars *et al.* (1996). Although much better than CY47, there is a suggestion of northward displacement and reduction in the precipitation maximum in day 2 forecasts. It appears that the initialisation of soil moisture at field capacity at the beginning of July in Beljaars et al. was crucial to obtain a good simulation of the excessive rainfall events.

5.2 Boreal forests

Surface albedo is the prime regulator of the net energy available at the surface. The albedo of snow-free land surfaces ranges from values of 0.1 in forests to values of 0.35 over deserts. For areas seasonally covered with snow, that range can extend up to 0.85. Betts and Ball (1997) analysed the annual cycle of albedo in the BOREAS experiment, performed in Canada during 1994, focussing on the snow season, comparing several measurement sites located over grass, aspen and coniferous. Representative values for daily averaged albedo of snow-covered grass sites are 0.75, while corresponding values for the aspen and conifer sites are 0.21 and 0.13, respectively, with values as high as 0.4, one to two days after snowfall. The lower albedo of the boreal forests in the presence of snow corroborates data from other observational studies and the few attempts of making a hemispheric-satellite based estimate of albedo.

The ECMWF model version of 1994, at the time of the BOREAS experiment, treated the albedo of snow covered areas with no regard to the land cover: beyond a critical value for snow depth, the albedo of snow covered areas was rarely outside the 0.7-0.8 range. As a result, when compared to experimental results, net radiation was too low and near-surface air temperatures were too cold. A modification to the scheme was designed such as the albedo of snow-covered surfaces tended to the asymptotic value of 0.2 in the presence of forests and 0.7 otherwise (Viterbo and Betts 1999b). The modified scheme has much reduced biases in temperature and radiation in the high latitudes. The cold bias in temperature in the control scheme extended in the vertical to the whole troposphere, increasing with forecast range and affecting most continental areas. The bottom panel of Fig. 7 shows the day 5 forecast error of 850 hPa temperature, averaged for March and April in 1996. The very high albedo induces cooling errors exceeding –3 K in North America and –7 K in Asia. The top panel shows the corresponding figure for 1997. In spring 1997, with the new snow albedo scheme, the cold bias over the boreal forest has been almost eliminated. The new scheme also improved the quality of the medium-range forecasts, as evidenced by better scores of 500 hPa geopotential fields.

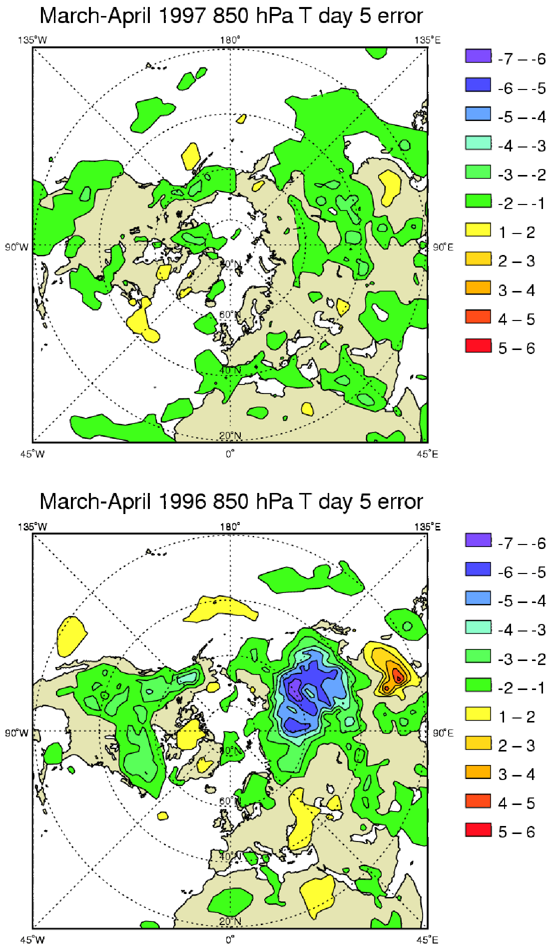


Figure 7. Comparison of the average 5-day forecast temperature errors at 850 hPa, for the ECMWF operational model during March-April 1996 (bottom) and 1997 (top) (from Viterbo and Betts 1999b).

The forecast results above corroborate the study of Thomas and Rowntree (1992) on the role of the boreal forests in conditioning the climate at high latitudes. The spring months of two five-year experiments, the first with a (realistic) snow albedo and the second with the high latitude forests removed are compared. The latter experiment is colder than the former in the continental areas north of 50 N. Pielke and Vidale (1995) suggested that the boundary between tundra and boreal forests is a region of enhanced horizontal temperature gradients, acting as a pre-conditioner for baroclinic instability and "locking" the climatological position of the polar front. In analysing further refinements to the ECMWF snow model, van den Hurk *et al.* (2000) show that (a) simulating the boreal forest control on evaporation in spring (reduced transpiration from frozen soils) and (b) increasing the runoff over frozen soils, improves the agreement of model results with observations.

5.3 Soil water freezing: A regulator of cold climates

The operational ECMWF forecasts for the winters of 1993-94 through to 1995-96 showed a tendency to produce a cold bias over continental areas in winter. This error was particularly severe during the winter of 1995-96 for Scandinavia, a year characterised by reduced snow amounts and, consequently, larger thermal coupling between the soil and the atmosphere above. Viterbo *et al.* (1999) diagnosed two main problems contributing to that error. Firstly, the energy involved in phase changes in the soil was not taken into account. When positive temperatures approach 0 C, a substantial amount of the external cooling demand (i.e., infrared cooling) is used to freeze the soil water, thereby decreasing the rate of soil cooling; a similar effect occurs in melting. Soil water phase transitions act as a thermal barrier, increasing the soil inertia at temperatures close to 0 C. Secondly, the downward sensible heat flux, prevailing in winter conditions, was too small, leading the model into a positive feedback loop where cooling reduced the heat flux, and made the soil even colder. Model changes were designed to incorporate the missing physical mechanisms. Separate seasonal integrations with both model schemes revealed a greatly alleviated soil and near-surface atmospheric cooling drift: Screen-level temperatures reduced from –10 C to close to zero. Despite the considerable warming in the model soil and near-surface winter climate in continental areas, there was negligible impact on the free atmosphere temperature and the atmospheric flow. In winter, stable situations, the atmosphere is decoupled from the surface and changes at the surface do not propagate upwards, unless they affect the momentum budget.

6. Conclusions

The examples presented above have shown that land surface can have a significant impact on the atmosphere at the synoptic/continental scale when it affects the partitioning of the surface energy into sensible/latent heat, via the soil water. This effect can be local or non-local. On the other hand, land surface has a significant impact on the atmosphere at the synoptic/continental scale when it affects the net energy at the surface, e.g., change of the albedo in snow-covered areas in spring. However, in winter, statically stable, conditions the land surface is decoupled from the atmosphere: Large variations in surface temperature affect only the lowest hundred metres of the atmosphere and do not have a significant impact on the circulation.

Recent experience at ECMWF, summarised above, shows that forecasts systems are sensitive to misrepresentations of longer timescales in the land-surface/atmosphere interaction. Progress can only be made with sustained efforts on validation of the model components. Data assimilation/forecast systems are ideal tools to validate parameterisations because of their constant confrontation with observations (in a “perfect synoptics” scenario) and a very large community of users requiring accurate diurnal cycle of weather parameters. In this context, GEWEX (Global Energy and Water Experiment) Continental Scale Experiments and the Coordinated Enhanced Observation Period planned for 2002 will play a prominent role. If the surface and upper-air observations arrive at the NWP centres they can be assimilated. Analysis and short term forecasts from those centres are the best hope of having a coherent picture of surface and atmosphere variables and of the surface fluxes.

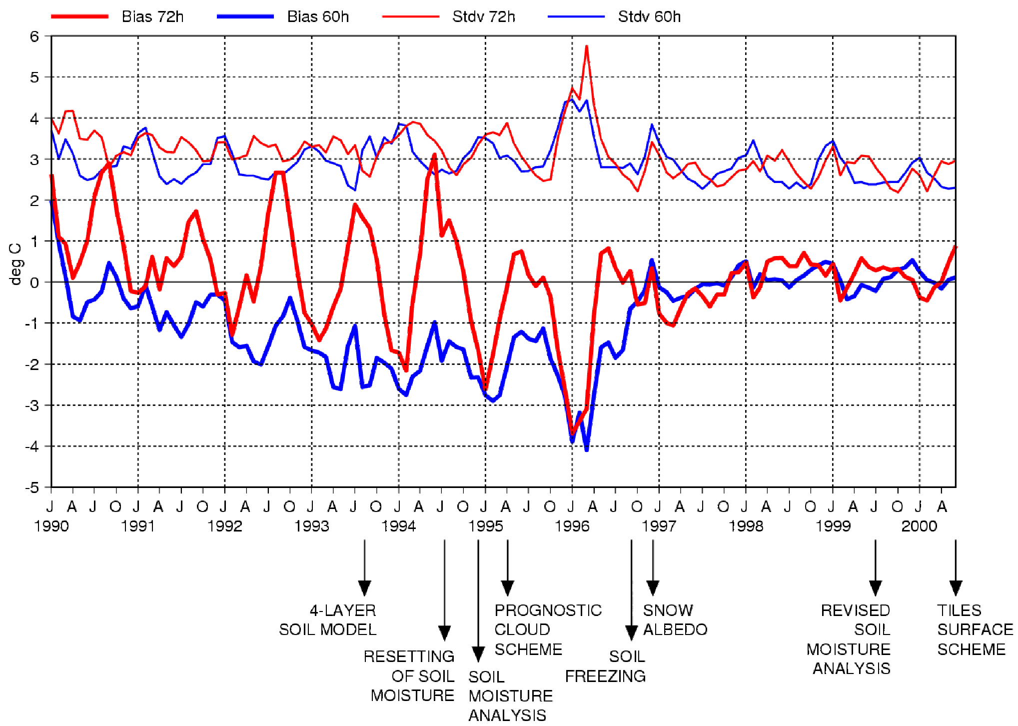


Figure 8. History of monthly biases (thick lines) and standard deviations (thin lines) with respect to observations of the daytime (72-hour: red line) and night-time (60 hour: blue line) operational two-metre temperature forecasts, averaged for all available surface stations in the European area (30º N to 72º N and 22º W to 42º E).

An important feedback loop in the surface-atmosphere interaction is the link between the anomalies of soil moisture and precipitation. Its correct representation in the model relies on the interplay of several parameterisations: evaporation and soil moisture, boundary layer processes, clouds and convection. Most NWP models have their poorest scores on precipitation in the tropics and spring/summer extratropics and improvements on the understanding of the soil moisture-convection interaction will most likely greatly alleviate that deficiency. Validation of model results against results from the Large Scale Biosphere-Atmosphere Experiment in Amazonia (LBA) (see forthcoming Special Issue of *J. Geophys.Res.* 2001) observations of the diurnal cycle of precipitation and atmospheric thermal and humidity profiles over the Amazon River Basin will probably be a key for progress in that area.

Probably the best summary of the impact of land surface on weather can be shown on Fig. 8 , displaying the history of ECMWF operational short-range forecast errors of 2 m temperature over Europe as a time-series of monthly averages. These errors show a large annual cycle, are different for night and day (72 and 60 hour forecasts verifying at 12 and 00 UTC, respectively), and have a rich history of the many model changes that were made over the years. We will discuss only the model changes made from 1993 onwards.

In August 1993, a surface scheme with a climatological deep-soil boundary condition for temperature and moisture was replaced by the free-running four-layer scheme (Viterbo and Beljaars 1995), but the impact is not very obvious. The summer daytime bias of August 1993 was smaller than that of the previous year but, at that time, the soil scheme has been running freely for only two-months (including the July parallel test described earlier). The next summer showed a pronounced warm bias related to a gradual drying out of the soil which was reduced in July 1994 by resetting the soil moisture to field capacity over vegetated areas. A simple soil moisture analysis scheme was introduced in December 1994 (Viterbo 1996) with a clear beneficial impact on the daytime bias for summer 1995. The night-time temperatures have been biased cold for many years, related to an overly large amplitude of the diurnal cycle. The winter of 1995/1996 was particularly bad, mainly because the European area was blocked for most of the winter with easterly winds and very cold temperatures, although changes to the cloud scheme or the orographic drag might have had a negative impact on night-time temperatures. It is interesting that the reduction of the daytime bias actually increased the night-time bias by displacing the entire diurnal cycle to colder temperatures. Soil freezing and increased boundary layer diffusion in stable layers, introduced in September 1996, improved the monthly error statistics considerably. The wintertime bias was largely eliminated and the amplitude of the diurnal cycle was down to a reasonable level. The snow albedo reduction described in Section Boreal forests was introduced in December 1996, but its impact is not clear over Europe, due to the relatively small area covered by snow and the overall magnitude of the errors linked to the excessive soil cooling, corrected three months earlier. Finally, a much more selective way of initialising soil moisture (Douville *et al*. 2000), introduced in April 1999, might be responsible for a slight reduction of the standard deviation of temperature errors in that year. A new surface scheme was introduced in June 2000 (van den Hurk *et al*. 2000), but it is too early to assess its operational performance. It is fair to point out that the statistics presented in Figure 8 are averaged over a month and over a large area. The errors on a day-to-day basis can still be large, but are less systematic and are often related to errors in the forecasted clouds or the presence of snow.

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