Tellus (1999), 51A, 674–685
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TELLUS
ISSN 0280–6495

Relaxation of soil variables in a regional climate model

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(Manuscript received 1 October 1998; in final form 12 March 1999)

ABSTRACT

A regional climate model, HIRHAM4, has been used with perfect boundary conditions from the ECMWF reanalysis project with an integration area covering the Mediterranean area. The model has been applied iteratively to one particular year, 1982, in the sense that boundary conditions corresponding to that year were applied for several "years" (iterations); soil temperature and moisture have been allowed to evolve through the simulation, whereas all atmospheric variables have been reinitialized for each iteration. This way the soil variables, in particular temperature, approach a stationary state with a vanishing net energy and water exchange with the atmosphere. This approach over most of the integration area follows an exponential function after a transient period of not more than 1 year. This has the consequence that the asymptotic soil temperature can be calculated after just 4 iterations of the model, resulting in an estimate much closer to equilibrium than the value after 4 years. Soil water is much more noisy and reaches approximate equilibrium after a few iterations; the extrapolation procedure for this quantity is only recommended for arid regions. In general, the present experiment constitutes a method to distinguish between the transient behavior of slowly varying fields and the inter-annual variability. The asymptotic state is independent of the initial soil state before the first iteration and is therefore a property only of the model and of the forcing through the integration period. This way of obtaining a "canonical" soil state may, for instance, have uses for model inter-comparisons, as the arbitrary influence from initial transients has been removed.

1. Introduction

The general initialization problem for numerical weather prediction and climate models has received great attention. While the initialization of the atmosphere is very important for the quality of weather predictions, climate simulations are more affected by variables with a large inertia like the state of the sea and of soil variables. In general circulation models (GCMs) and regional climate models (RCMs) the soil initialization, which is the subject of the present work, has important consequences for the energy and water balance of the atmospheric model.

Attention has been paid to the feasibility of initializing soil moisture from observations (Mahfouf, 1991). In this work the soil variable was initialized through a fit to results from a measuring campaign in a particular region. While this method seems reasonable for obtaining initial soil variables for weather prediction models in areas with observations, it is not possible to apply this procedure when the goal is climate sensitivity investigations. In such studies the same procedure has to be employed for both control and scenario experiments. In this case the main objective will be to avoid influence from transients as much as possible. This may be approached by the method presented here, where stationary values of soil temperatures are obtained by simulating successive instances of the same "typical" year.

This method is somewhat similar to the multi-year spin-up of the ocean part of coupled ocean–

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atmosphere GCMs (CGCMs) with climatological atmospheric fields. When simulations are performed with coupled ocean–atmosphere models, great care is normally taken to ensure that the ocean is in equilibrium with the average atmosphere. For a treatment of this issue see, e.g., Hansen et al. (1985). Equilibration of ocean models is performed by imposing climatological atmospheric fields as upper boundary conditions and integrating the ocean model for a long time (centuries). Several problems need to be taken care of in the design of a CGCM study. During a coupled experiment, flux corrections normally have to be employed to avoid a drift of the mean climate away from a realistic state. In transient climate change experiments there are important problems (the cold start problem, Hasselmann et al. (1993)) arising from the fact that, e.g., the rise in CO_2 concentrations have a shorter characteristic time than the heating of the oceans; this means that it is not the correct starting point to assume present climate with an ocean in equilibrium with the atmosphere.

The problem of initializing soil variables is not as hard as the ocean relaxation problem when looking at long-term climatology, the main reason being that the heat capacity of the relevant soil column is orders of magnitude smaller than that of any model ocean, resulting in much shorter characteristic time scales. Furthermore, we are not dealing with lateral fluxes in the description of the soil. The soil–atmosphere interaction is taken care of column-wise, normally with local conservation of energy. Contrarily for CGCMs, errors in the lateral gradients of temperature and salinity will give rise to erroneous flow that may change the mean global patterns of ocean–atmosphere energy exchange.

Initialization of soil variables is important, though, at an annual to decadal time scale. Due to the one-dimensional nature of the parameterization and the diffusive equations governing heat transport in the soil, the soil variables will tend to drift towards a quasi-stationary state with a vanishing long-term energy exchange with the atmosphere. It has been seen (Christensen et al., 1997) that errors in the soil state does influence the model climate. But the time scale of a soil state drift is expected to be of annual time scale and thus resolvable by numerical experiments as demonstrated in the present work.

Previous investigations of the relaxation of soil variables interacting with the atmosphere are quite scarce. In the PILPS project (Project for Intercomparison of Land-surface Parameterization Schemes (Chen et al., 1997)) a comparison of different land-surface parameterizations was made by off-line simulations at the Cabauw measurement site in the Netherlands. These models were allowed to spin up by repeating the forcing as many times as needed to get below a certain iteration-to-iteration change of the relevant fields. In an off-line simulation there is, of course, no feedback with the atmosphere, but the relaxation time needed (typically 3–5 years) is expected to be comparable to relaxation times of full runs with an atmospheric model. It is, however, a question that has to be looked into, how relaxation times do depend on feedback to the atmosphere.

It is a difficult and computationally expensive task to make very long spin-up runs of a model in order to avoid sensitivity of the result to the initial values of soil temperature and water content. One important aim of this paper is to set up an experiment where the effects of transients can be examined without the simultaneous variation due to natural interannual variability. This will be done by a sequence of 1-year simulations (iterations) with an RCM with boundary conditions relating to one particular year. At the beginning of each iteration, the initial atmospheric state will be reinserted, but the soil variables will be allowed to evolve throughout the experiment. It will be shown that the soil temperature actually has a physically meaningful equilibrium value that can be accessed by a spin-up run of a limited number of years and a subsequent extrapolation of these results. Thus, it may be possible on these grounds to develop a method for the initialization of fields with a multi-year relaxation time like the soil temperature.

For most fields without a very long relaxation time, a 1-year spin-up period will be quite sufficient, and it will not matter whether the preceding N years or N iterations of one preceding year are used as spin-up algorithm. As it will be illustrated below, however, there are several reasons to apply the latter method for fields with a long memory, e.g., the soil temperature. Firstly, the interannual variability of these fields will be rather small, and secondly, it should be feasible to perform an extrapolation of the field value to its stationary value in order to eliminate transients. Ideally, for a numerical experiment simulating actual climate, a very long simulation of the years preceding the period of interest would be made before its start. This way, e.g., a cold period in the years prior to the experiment would be reflected in low initial soil temperature values. This is most important for experiments where present climate is investigated. For climate simulation purposes it seems important to obtain initial values of soil variables that are determined by equilibrium properties of the model itself, such that results are not determined by a more or less arbitrary choice of initialization. It will be important, though, that the period used for an iterative initialization is representative for the climate, i.e., not too anomalous; this will in reality not be possible for the whole region of interest, of course. In the case of present-climate simulations, where the year prior to the actual experiment is not typical of the preceding period in general, hybrids of the present algorithm might be considered; e.g., a few iterations of a period of more than 1 year preceding the time of interest.

The present study has branched off from an RCM investigation of possible feedback mechanisms which could have maintained the much less dry Mediterranean climate of Roman times, a part of the EU-project "Land-surface processes and climate response". The advantage of applying a regional climate model in the present work is that the atmospheric internal variability is limited at the same time as a physical feedback is working between atmosphere and soil. By repeating boundary conditions from one particular year several times, we will drastically reduce interannual atmospheric variability; it is important to realize that some internal variability does remain, though, even if the atmosphere will be fixed at the boundaries of the integration region. The internal variability of regional climate models are discussed further in Polcher (1999). Regional models have a long history (Giorgi, 1990) and are known to be able in general to reproduce the large-scale climate of the boundary conditions while improving the climate at smaller length scales; on the other hand, errors in the large-scale forcing cannot be mended by the regional model (Noguer et al., 1998; Christensen et al., 1997; Christensen et al., 1998) The local atmosphere–soil interaction is described at least as realistically as in GCMs; therefore it is expected that the relaxation mechanism will be similar in GCMs and RCMs.

In this paper we will study the relaxation of soil temperature and moisture during an iterative experiment with a regional climate model as described above. We will investigate a region covering continental Europe and Northern Africa with the help of a regional climate model in order to look at regional variations of relaxation times for soil variables and the possibility of extrapolations to the stationary values.

We are using so-called perfect boundary conditions, reanalysis fields from the European Centre for Medium Range Weather Forecasts (ECMWF). The boundaries reflect one particular year, in this case 1982. The experiment runs for several years, with the boundary conditions of 1982 repeating themselves in a cyclic way. The model atmosphere is reset to values corresponding to the beginning of 1982 at the start of each iteration, but the soil variables are allowed to evolve.

While ignoring the variability of the soil variables due to the interannual climate variability, this approach allows an extrapolation of soil temperature and moisture towards a stable state, which is in balance during the particular year of the experiment. For variables like the soil water content, which has quite short variational time scales, the value at the end of an iterative spin-up run should be a reasonable choice; without discernible relaxations of multi-year time scale, a treatment of the year prior to the iteration year should not be necessary.

2. The numerical model

The regional climate model HIRHAM4 employed in this study is based on the HIRLAM* short range weather prediction model (Källén, 1996). In order to make a model that is suitable for long climate integrations, the more advanced physical parameterization of the Max-Planck Institute climate model ECHAM4 (Roeckner et al., 1996) has been incorporated into the model. A detailed description of the combined model,

* High-resolution limited area model; developed by the national meteorological institutes in Denmark, Finland, Holland, Iceland, Ireland, Norway, and Sweden, later also in cooperation with France and Spain.

called HIRHAM4, can be found in Christensen et al. (1996), but see also Christensen et al. (1998). Some of the important characteristics of the model are summarized below.

HIRHAM4 is a standard primitive-equation Eulerian staggered grid point model with a prognostic cloud water equation. The model has a variable number of vertical hybrid levels, at present 19. We operate with a lateral boundary relaxation zone, currently 8 points wide, following Kållberg and Gibson (1977) with a quasi-exponential relaxation function for wind, temperature, and surface pressure. Moisture and cloud water, however, are relaxed according to a so-called inflow/outflow scheme where only the value on the edge of the area is modified: if the flow is directed out of the integration domain, a value extrapolated from upstream quantities is applied at the model levels, otherwise the boundary value is taken from the coarser-resolution field. The inflow/outflow relaxation strategy, originally documented in Giorgi et al. (1993) and widely used, yields a better consistency of cloud cover between the relaxation zone and the model interior.

An advanced parameterization for precipitation processes is adopted. Convection is described in the Tiedtke (1989) mass-flux formulation with modifications to the formulation of deep convection (Nordeng, 1994). Liquid water in stratified clouds is a prognostic variable and treated according to the Sundqvist (1988) scheme with modifications according to Roeckner et al. (1992).

Land surface parameterization uses 5 temperature layers and one (bucket) moisture layer. A zero-flux lower boundary condition is applied for temperature. The thickness of the 5 layers increase in order to include time scales between 3 h and one half year. Runoff is calculated within the Arno scheme (Dümenil and Todini, 1988), which softens the onset of runoff by introducing a distribution of soil water holding capacities in the grid box. There are no vegetation types, only a vegetated fraction; vegetation differences are described directly through maps of climatological albedo, soil water maximum, roughness length, and leaf area index.

The global data set of fields of land surface parameters used in ECHAM4 (Claussen et al., 1994) is constructed from the major ecosystem complexes of Olson et al. (1983). Surface mean orography and variances are obtained from the US Navy database, which has a spatial resolution of $1/6^\circ \times 1/6^\circ$. Other surface fields are based on Claussen et al. (1994). All parameters are constant in time, i.e., not depending on the month. Initial soil temperature is based on a 5° monthly surface air temperature climatology database, where time lags and amplitudes in the various soil layers are introduced in order to reproduce the correct annual cycle.

Soil moisture is started from empty reservoirs in this experiment in order to enhance the relaxation amplitude and thus make the relaxation easier to study. In regular spin-up runs it is of course more sensible to use a more realistic value, like the climatological value normally used with HIRHAM (Roeckner et al., 1996; Källén, 1996).

3. Experimental set-up

The integration area, shown in Fig. 1, comprises Central and Southern Europe, the Mediterranean Sea, and a large part of Northern Africa. With a horizontal resolution of 0.5° or about 55 km in a rotated latitude–longitude system, there are 102×61 integration points. The time step applied is 5 min.

The experiment has been performed with boundaries taken from the ECMWF reanalysis project (ERA) from the year 1982 and 5 days from the end of 1981. The resolution of ERA is T106, or approximately 150 km. The model was spun up for 5 days corresponding to 27–31 December 1981. Soil temperature was initialized as described in the previous section. The initial soil temperature thus comes from a low-resolution climatology, and the soil reservoirs are initially empty. The atmosphere and soil states were then stored, and 7 iterations of the whole year 1982 were performed. At the end of each iteration, the atmospheric state of 1 January 1982 was substituted in order to be consistent with the boundary forcing, but the soil variables were allowed to evolve.

4 Results and discussion

As a simple model approach to the relaxation of the variable T, representing temperature or moisture, possibly temporally and spatially averaged, towards equilibrium, let us study consecutive

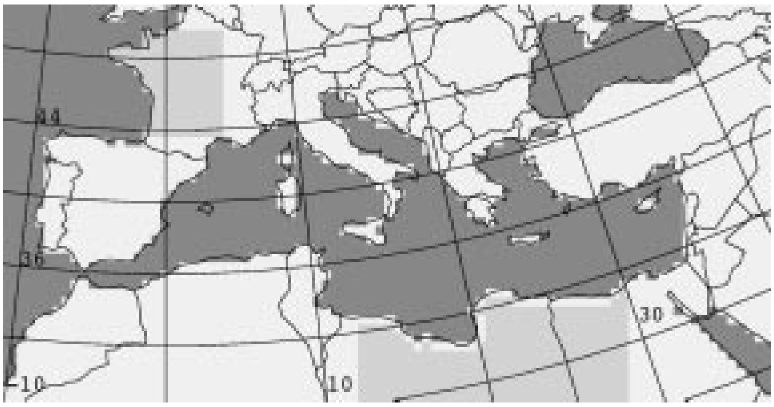

Fig. 1. Land–sea mask of the Mediterranean integration area. The grey land areas are the two areas "France" and "Libya", over which horizontal averaging has been performed for Figs. 2, 3.

annual averages T_i. Assuming for simplicity that the speed of relaxation is a constant in time, but that the "instantaneous asymptote" T_i^0 depends on the year, a discretized equation for the time evolution of T might look like

$$\frac{T_i - T_{i-1}}{\delta t} = -a\left(\frac{T_i + T_{i-1}}{2} - T_i^0\right). \tag{1}$$

This is a trapezoidal implicit formulation of a linear forced difference equation, see e.g., Mesinger and Arakawa (1976). Other forms of this equation are of course possible, but other implicit or explicit forms of eq. (1) do not change relevant aspects of the solution significantly.

The solution to this equation has the form

$$T_j = \Delta T^{\text{transient}} + T_j^{\text{inf}}$$
$$= K\,\mathrm{e}^{-jc} + \frac{a\,\delta t}{1 + a\,\delta t/2}\sum_{k=0}^{\infty} \mathrm{e}^{-kc}T_{j-k}^0, \tag{2}$$

where $c = \log((1 + a\,\delta t/2)/(1 - a\,\delta t/2))$, a is the relaxation constant, and δt is the time step (in this connection 1 year). This applies for $a\,\delta t < 2$, meaning that fields with too short relaxation times of less than 2 years cannot be treated sensibly with an equation like eq. (1). Here T_j^{inf} is the asymptotic solution at year j as the time of initialization of T approaches minus infinity; K is an integration constant, which allows the expression to be rewritten as

$$T_j = (T_{j_0} - T_{j_0}^{\text{inf}})\,\mathrm{e}^{-(j-j_0)c} + T_j^{\text{inf}} = A_j + B\,\mathrm{e}^{-cj}, \tag{3}$$

where the variable at a particular time j_0 is specified to be T_{j_0}. In the present iterative experiment, the asymptote will be a constant in the iteration number j, so that we will have $T_j = A + B\,\mathrm{e}^{-cj}$, A being the same as the constant value T_j^0.

Assuming that the asymptotes T_j^0 are uncorrelated and normally distributed with mean T^0 and variance σ^2, eq. (2) has the consequence that values of T_j^{inf} are normally distributed with mean T^0 and variance $\sigma^2 a\,\delta t/2$. Thus, small values of $a\,\delta t$ and therefore long relaxation times $\tau = 1/c \simeq 1/(a\,\delta t)$ are synonymous with small sensitivities to chaotic noise of the atmospheric forcing. In other words, variables of a larger inertia, like total soil temperature, are more likely to be of use for this approach than variables with a smaller inertia. This happens for two reasons: The relaxation takes so long time that it can be separated from the intra-annual variations of the field in question, and the inter-annual variability of atmospheric fields are suppressed due to the inertia of the field as just mentioned. The possibility of distinguishing the relaxation in an actual numerical experiment depends on the relaxation time as mentioned here, but also on the relative amplitudes of the relaxing transient $\Delta T^{\text{transient}}$ and the variability of T_j^{inf}.

The present experiments exhibit some noise in the forcing fields, e.g., precipitation, but this noise only arises from internal variability of the model (Polcher (1999) for a fuller treatment of this quantity) and not from the full atmospheric climate variability, as would be the case for an atmospheric model running for several years. Thus, we have a chance to be able to observe transient behavior through the experiment.

Here we have chosen to focus on the temperature of the total soil layer, which in ECHAM4 has a thickness of 9.8 m, and on the bucket soil moisture reservoir. The three parameters needed in eq. (3) are fitted from the time series of annual averages of soil temperature and moisture, T_i, w_i, $i = 1, \ldots, 7$, respectively.

In Figs. 2, 3 are shown area averages of temperature and water content for two land areas as defined in Fig. 1. "France" as representative of a non-arid and "Libya" as representative of a dry region. Apart from T and w (upper panels) we also show the surface forcings in the lower panels. It is obvious from the latter that the first iteration does not belong in a regime where there exists an approximately linear relationship between field and forcing as assumed in eq. (1) (note that some of the initial forcing points are outside the plotting range).

In the upper panels of these two figures are shown as full lines an exponential fit to the points as in eq. (3). The degree of fit of these curves indicate that temperature, as well as soil moisture in Libya do indeed follow quasi-exponential curves, and that noise is eliminated by the area averages. In the case of soil moisture in France, the least-squares fitting does not converge, and no curve has been drawn. The anomalous point in iteration 6 shows no indication of being an artifact by closer inspection of the data.

Both with respect to temperature and water content the two areas here chosen have qualitat-

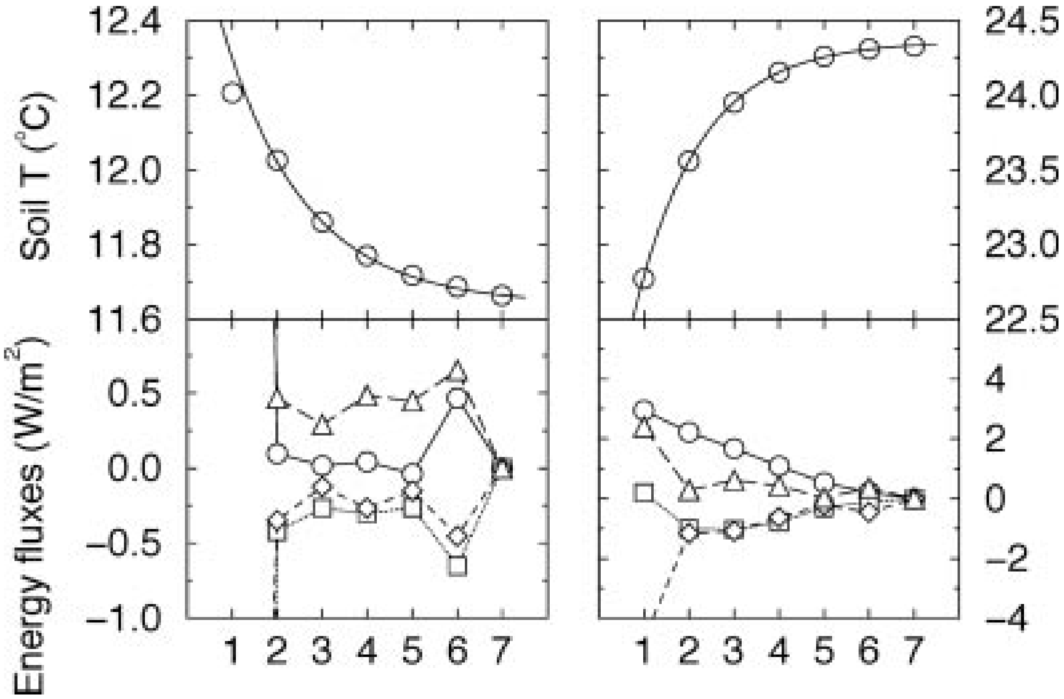

Fig. 2. Area averages of total soil temperature in degrees Celsius (top row) and surface downward energy fluxes in W/m^2 (bottom row) for France (left column) and Libya (right column): see Fig. 1 for definition. The energy fluxes are: circles: latent heat flux; squares: sensible heat flux; diamonds: long wave radiation; triangles: solar radiation. NOTE that fluxes are plotted relative to the value of iteration 7 in order to show the time evolution in one plot. These values are: -49.3 W/m^2 downward latent heat flux, -9.7 W/m^2 downward sensible heat flux, -50.1 W/m^2 downward long-wave radiation, and 110.9 W/m^2 solar radiation. Solid lines in the upper row correspond to fits of (A, B, τ) of eq. (3) to the two area-averaged time series, omitting the first iteration; fitted parameters are (11.64°C, 1.14°C, 1.83 y) for France, (24.36°C, -3.10°C, 1.48 y) for Libya.

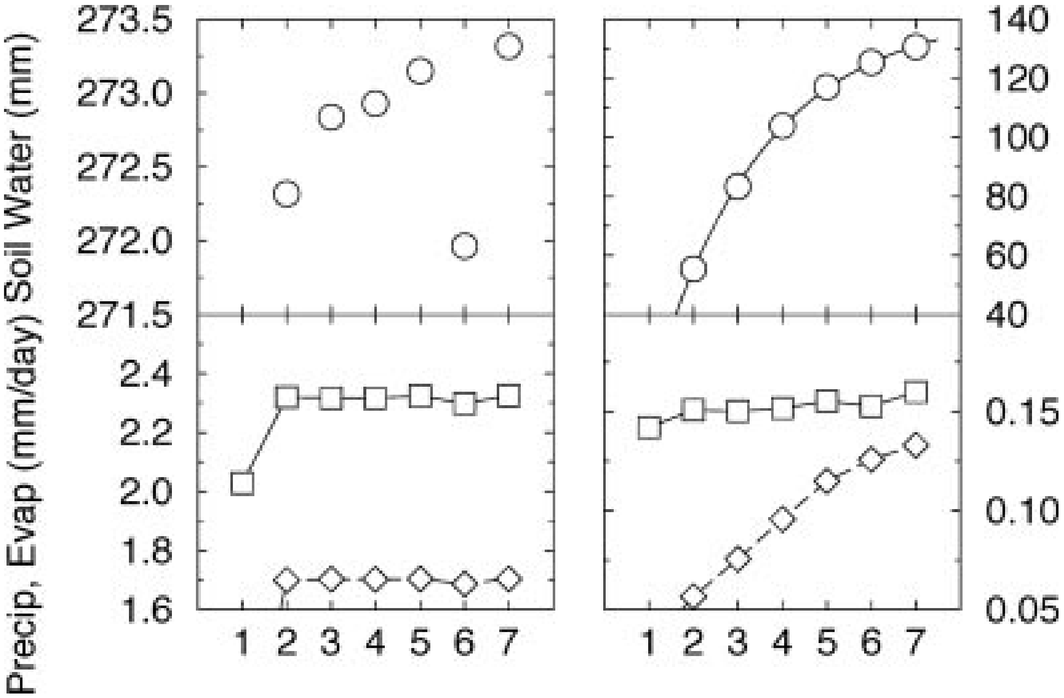

Fig. 3. Area averages of soil moisture (top row) as well as precipitation (lower row, squares) and evaporation (lower row, diamonds) in mm/day for France (left column) and Libya (right column): see Fig. 1 for definition. Solid lines in the upper right panel corresponds to an exponential fit and give (142.8 mm, −196.1 mm, 2.49 y) for Libya, omitting iteration 1. For France the least-squares exponential fit diverges to infinite relaxation time.

ively different evolutions of the fluxes. In the arid desert region, evaporation increases through the entire simulation, as water reservoirs fill up, and thus supplies a restoring force, both with regard to the temperature and the soil water evolution. In France, the forcing is rather noisy after the first iteration, and particularly soil water reflects this noise. The size of the relaxation is much smaller in both cases, and there is not a simple relation between forcing and signal. On this background it is surprising that the best-fit relaxation times for the two different areas are quite similar, a few years; only the relaxation amplitudes differ markedly. In the arid case it is demonstrated that a short relaxation time does not by itself preclude a possible usefulness of an extrapolation procedure for soil fields.

The fits result in three fields of parameters, i.e., A, B, and c of eq. (3). We have performed different sets of fits based on varying choice of which iterations to use. In the following we will compare two different choices: In the first we use 3 out of the first 4 iterations, ignoring the very first iteration in order to allow for short transients to decay; this results in parameter fields with index 34 (parameter fields A_{34}, B_{34}, and c_{34}) and the last 6 of all 7 iterations, index 67. A comparison of results of these fits is found in Figs. 4–9. The parameters for soil water will be called A^w, B^w, and c^w, respectively. In all figures the white color signifies areas where the exponential fit is not a good approximation, i.e., areas with so small relaxation times or amplitudes that the time series is dominated by noise. In the case of only 3 points being used for the fit the following algorithm is applied: A decaying exponential function can be drawn through the 3 points provided that $0 < (T_4 - T_3)/(T_3 - T_2) < 1$. Otherwise we substitute the last point T_4 for A and an out-of-range value for c. When there are more than 3 points we perform an iterative gradient-expansion minimization. For w, there is a maximum water hold-

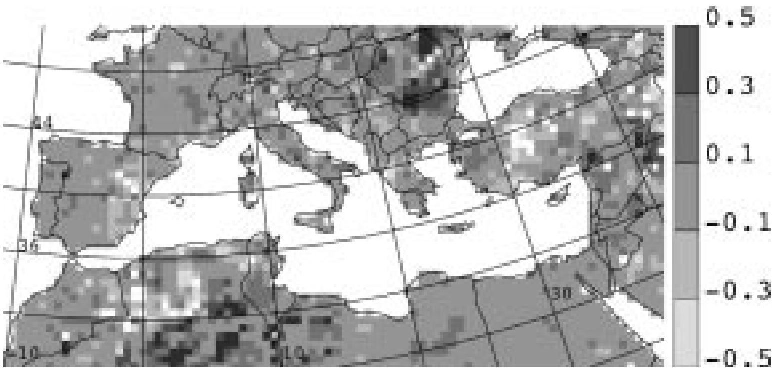

Fig. 4. Soil temperature annual average (°C): difference between asymptotes based on points 2–4 and 2–7, $A_{34} - A_{67}$.

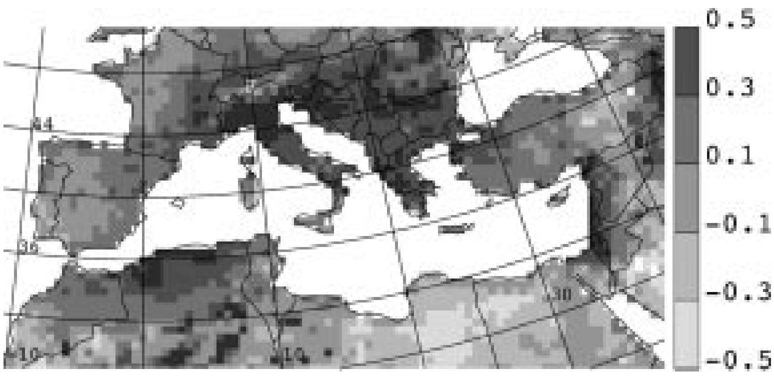

Fig. 5. Soil temperature annual average (°C): difference between iteration 4 and the asymptote based on all iterations (2–7), $T_4 - A_{67}$.

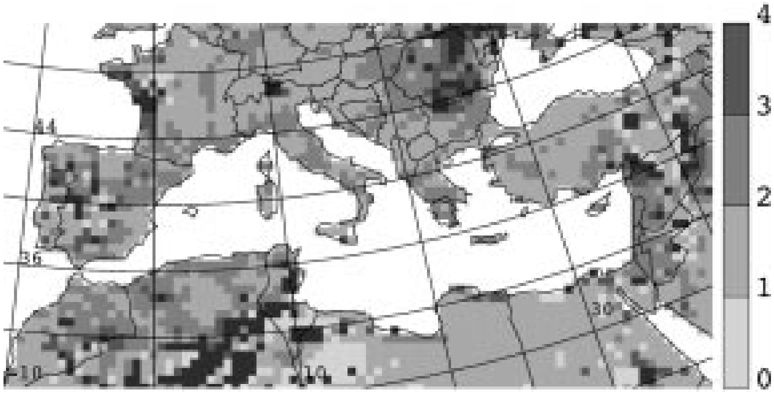

Fig. 6. Soil temperature exponential relaxation time $\tau(y)$.

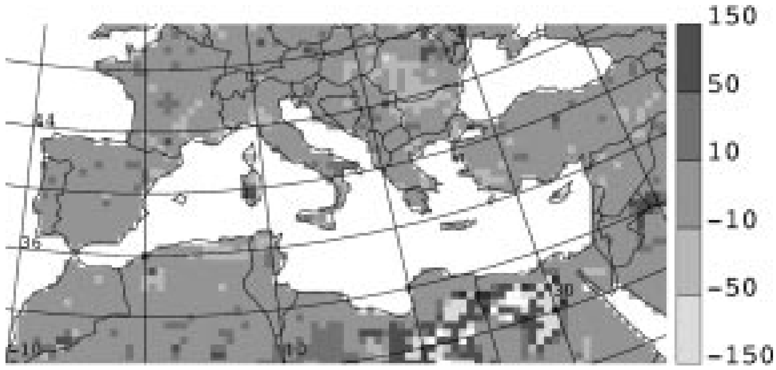

Fig. 7. Soil water annual average (mm): difference between asymptotes based on points 2–4 and 2–7, $A^w_{34} - A^w_{67}$.

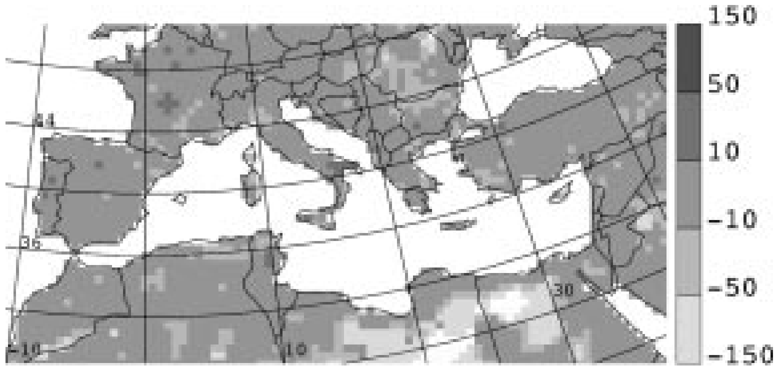

Fig. 8. Soil water annual average (mm): difference between iteration 4 and the asymptote based on all iterations (2–7), $w_4 - A^w_{67}$.

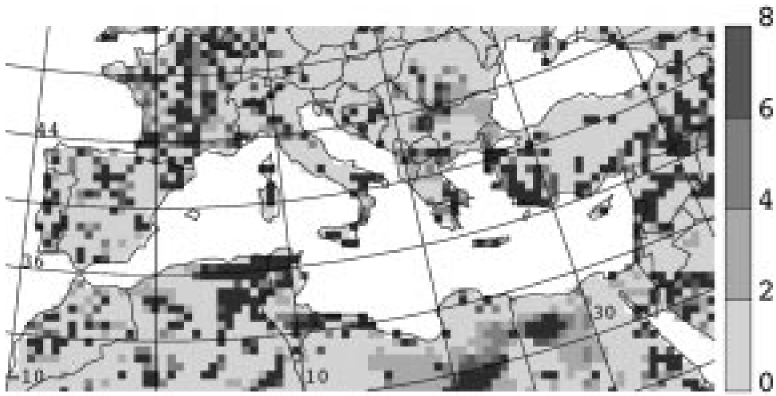

Fig. 9. Soil water exponential relaxation time $\tau^w(y)$.

ing capacity w^{max}. In the case $A^w > w^{max}$ we set $A^w = w^{max}$ and do a two-parameter fit of B and c to the data set.

Let us now consider the maps Figs. 4–9, first of temperature and subsequently of soil water content.

A comparison of Figs. 4, 5 shows that the exponential fit is good over most of the integration area. After 7 iterations the temperature is within 0.1° of the asymptote over most of the area (not shown). The asymptote estimated with the first 4 iterations does not differ significantly from the best estimate available (Fig. 4). In contrast, the actual temperature after 4 iterations is much farther away from the asymptote (Fig. 5), consistent with the order of magnitude of the relaxation time being a few years. Typical deviations are 0.5°. Corresponding differences in surface fluxes can be estimated to up to several W/m^2 from Fig. 2.

In the case of soil water the situation is somewhat different. Fig. 8 shows differences from w_4 from the best asymptote. The corresponding quantity $w_7 - A^w_{67}$ (not shown) is of comparable magnitude. Thus, over most of the area, equilibrium has been reached to within the noise level, the arid region around Libya being a notable exception. Because of noise, the asymptote based on 4 iterations is not better than the value at iteration 4, since the fitted parameters are very sensitive to the noise in w. Noise is important as well in Libya for the same reason. Estimates of the relaxation time depend critically on the noise of the time series, as can be seen in Fig. 9, where values vary significantly; the values of τ do not seem trustworthy. An area average, as performed in Fig. 3, removes some of the noise and would be a necessary approximation for practical purposes. Tests with horizontal smoothing of the field (not shown) with a smoothing width of 5 grid boxes show, however, that deviations are still large, and the 4-point asymptote is still not significantly better than the smoothed value of w_4 compared to A^w_{67}. Thus, this point needs further investigation.

5. Conclusion

The present work is an investigation of the possibility of obtaining a soil state for initialization of regional climate model simulations by running the model to soil equilibrium, investigating regional differences in an area comprising Europe and Northern Africa. The feasibility of extrapolating annual mean values towards the equilibrium value on a point-by-point basis has been investigated for total soil temperature and soil water by repetition of 1 year of boundary conditions for the regional model.

The aim with this procedure was to establish a technique to obtain starting conditions where transients are minimized and less random than what has been the case previously. Effects of transients have to be suppressed as much as possible in modern numerical climate experiments, as climate sensitivity experiments operate with quite small signals. The method proposed here will be useful for this purpose, for GCMs as well as for RCMs, though the computational effort of doing multi-year spin-up runs and thus the potential gain by performing an extrapolation is largest for the latter. To the author's knowledge such an investigation has not been done before, probably because of the large computational requirements for the iterative procedure. This also means that much remains to be done; a systematic investigation of the seasonality as well as the interannual variations of the relaxation procedure needs to be performed, as well as a development of optimally robust fitting algorithms, before the present conclusions can be accepted as general.

The extrapolation procedure was far more successful in the case of soil temperature than in the case of soil water. Relaxation times were of the order of a few years, which is longer than internal soil relaxation times and longer than typical spin-up simulation lengths; in other words, the extrapolation procedure is well suited to save some spin-up time. Since internal temperature transients have a characteristic time of up to half a year, annual averages do indeed seem to be sensible time averages with which to do the time series analysis.

In contrast to the temperature case, the extrapolation procedure did not improve the soil water field compared to simply using the last value obtained during the iteration procedure. Only in Northeastern Africa were there signs of the soil water not having reached its equilibrium value early on; and in this area noise was very large. It is relevant to do a study similar to the present, but with filled water reservoirs as initial condition for the iterations. Anyway, the precipitation in the area under discussion is so low, around 0.15 mm/day (Fig. 3), that no measurable effect is expected of the timing of evaporation of this amount of water.

It remains to be investigated whether the present exercise might as well be done by choosing a number of representative points and then doing a one-column iteration series for each point. In other words, to determine the amount of feedback. It is obvious, e.g., from the small precipitation values at the first iteration in Fig. 3, that stand-alone simulations with the surface scheme alone is not sufficient. The possible success of a one-column experiment with regard to obtaining a suitable extrapolation depends, however, on a good choice of representative points; this seems to require full-blown calculations like the present one to obtain this kind of regionalization of the relevant area. It may also be that representativeness depends on weather type and therefore would have to be redone for each experiment anyway.

This paper does not compare the noise related to interannual climate variability to the feasibility of obtaining equilibrium values from a time series. It has been pointed out that long relaxation times are synonymous with a low sensitivity to interannual variability, but the present work has shown that the amplitude of the transient field is just as important a parameter as the relaxation time for determining the suitability of applying the extrapolation procedure to a particular region. Thus, an experiment with iterations of more than 1 year's duration could be useful.

6. Acknowledgments

This research was supported by the EC Environment and Climate Research Programme, contract ENV4-CT95-0112-PL950189, and by the Danish Ministry of Traffic. Invaluable help from Jan Polcher (Laboratoire de Météorologie Dynamique) and Jens Hesselbjerg Christensen (DMI) is gratefully acknowledged.

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