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The representation of soil moisture freezing and its impact on the stable boundary layer

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Abstract

The 1993 to 1996 version of the ECMWF model had a pronounced near surface cold bias in winter over continental areas. The problem is illustrated in detail with help of tower observations. It is shown that a positive feedback exists in the land surface boundary layer coupling, that has the potential to amplify model biases. If the surface is cooled too much, the boundary layer becomes too stable reducing the downward heat flux and making the surface even colder. This positive feedback is believed to be stronger in the model than in the real atmosphere resulting in diurnal temperature cycles that are too large, and excessive soil cooling on a seasonal time scale in winter.

An important contributor to the excessive winter cooling turns out to be the lack of soil moisture freezing in the model. The importance of this process is obvious from soil temperature observations. The seasonal soil temperature evolution shows a clear "barrier" at $0^o\,C$ due to the thermal inertia of freezing and thawing. A more quantitative illustration is the result of a simple calculation. This shows that the amount of energy necessary to freeze/thaw $1\,m^3$ of wet soil, would cool/warm this soil by about $50\,K$ if the phase transition was not taken into account.

To reduce the winter cold bias in the model, three model changes are described: (i) introduction of the process of soil moisture freezing, (ii) revised stability functions to increase the turbulent diffusion of heat in stable situations, and (iii) increase of the skin layer conductivity. The effect of these changes on the seasonal evolution of soil and $2\,m$ temperatures is investigated with long runs that have relaxation terms towards the operational analysis above the boundary layer. In this way the impact can be studied on the temperature forecasts for the winter of 1995/1996, during which the operational model showed considerable soil temperature drift over Europe. Also short periods of data assimilation (including 10 day forecasts) are carried out to study the diurnal time scales and the impact on model performance.

The model changes eliminate to a large extent the systematic $2\,m$ temperature biases for the winter of 1995/1996 over Europe and make the soil temperature evolution much more realistic. The soil moisture freezing in particular plays a crucial role by introducing thermal inertia near the freezing point and therefore reducing the annual temperature

cycle in the soil. The process of soil moisture freezing leads to a considerable warming of the model's near surface winter climate over continental areas.

1 Introduction

The new land surface scheme introduced in the ECMWF model in 1993 (Viterbo and Beljaars 1995) does not have deep soil climatological boundary conditions for soil moisture and temperature. This has the advantage that the model has the capability of describing anomalies in temperature and soil wetness, but at the same time there is the danger of drift. With free running soil variables, the seasonal evolution of soil moisture and temperature is determined by the energy and hydrological flux forcing provided by the atmospheric model in short range forecasts (so-called 6 hour first guess) during the atmospheric data assimilation cycle. Implicitly it is assumed that the atmospheric data assimilation keeps the soil variables under control. This is reasonable if the coupling between atmosphere and land surface is dominated by negative feedbacks. It is known however, that this does not work very well for soil moisture because a rather strong positive feedback exists. For instance, if soil moisture is low, evaporation is low and therefore precipitation is low and the soil moisture reservoir will not be restored sufficiently, leading to soil moisture drift. To control such drift ECMWF introduced in 1994 a soil moisture nudging scheme based on short range forecast errors of humidity at SYNOP stations (Viterbo 1996), which is a poor man's version of the optimum interpolation method developed by Meteo-France (Mahfouf 1991).

For temperature, one would expect that the turbulent exchange between surface and atmosphere provides sufficient negative feedback to keep the free running soil temperature under control. However, operational experience has shown that considerable drift can occur in winter when the turbulent exchange with the atmosphere is dominated by stably stratified flow with rather weak turbulent exchange e.g. at low wind speeds or high Richardson numbers (Beljaars 1995). Again a positive feedback is playing a role: if the surface becomes cold the boundary layer becomes more stable and therefore turbulent downward heat flux becomes less efficient resulting in less control of the drift (Derbyshire 1998). The lack of soil freezing in the model may be the reason for the soil temperatures becoming too cold, which is then further amplified by very weak boundary layer diffusion.

The purpose of this paper is to describe a series of model changes that prevent the ECMWF land surface temperatures from drifting cold in winter and to show the impact of these changes. It will be demonstrated that soil moisture freezing is important in providing a thermal barrier at the freezing point. Also the role of the vertical diffusion scheme will be illustrated.

It is well known that soil moisture freezing and thawing plays an important role in the thermal budget of large continental areas (e.g. Lunardini 1981; Williams and Smith 1989; Rouse 1984). However, it is still the exception rather than the rule that General Circulation Models (GCM's) and Numerical Weather prediction Models (NWP's) have this mechanism as part of the land surface scheme (for examples of models with soil water freezing see Verseghy 1991 and Slater et al. 1998).

To illustrate the nature of temperature drift problems in the 1993-1995 version of the ECMWF model (without soil moisture freezing), results of a diagnostic study will be presented in which short range forecasts are compared with tower observations (section 2). This study also shows the relevance of soil moisture freezing from the observational point of view. Section 3 and section 4 describe the implementation of soil moisture freezing and changes to the vertical diffusion scheme respectively. In order to study the impact on the seasonal time scale, long integrations at low resolution are performed. To have the same "realization" of the atmosphere in every integration, the atmosphere above the boundary layer is "nudged" towards the analysis of winter 1995/1996 when the soil temperature drift was particularly large in the operational model (we will refer to these long integrations as "relaxation" runs). This procedure also allows for a direct comparison of model results with observations from SYNOP stations and from soil temperature observations in Germany (section 5). Finally short range forecasts are performed in order to study the impact at operational resolution (section 6).

2 Comparison with observations

To gain more insight into the characteristics of the diurnal temperature cycle, the surface energy balance and near surface temperatures observations from the 200 m tower in Cabauw in The Netherlands are used. The period of November 1993 was selected, because it was a situation with flow predominantly from the east with a few cases of a very stable boundary layer. Comparisons of different components of the surface energy balance, of temperatures and wind below 200 m are shown in the Figs. 1-3. The surface energy parameters of the operational model are 6 hour averages of the forecasts up to day 1 and are compared with a time series of observed 6 hour averages (Fig. 1). First of all it is noticed that cooling by thermal radiation in the model is rather realistic on clear nights at a level of about $-80\ W m^{-2}$ (e.g. day 320-323), but that in cloudy situations (e.g. day 325,326,331,332) the model largely overestimates the long wave cooling. The short wave radiation is underestimated on clear days and overestimated on cloudy days, but on the time scale of a few weeks the cloud effects dominate and the total amount of solar heating at the surface is overestimated. The sensible heat flux looks realistic in comparison with observations. However, the downward latent heat fluxes (positive numbers) are rather low in the model. It is not clear how realistic the observed dew deposition fluxes are since they have been determined as a residual of the surface energy balance and have the accumulated errors from the other components. The downward "observed" latent heat fluxes are actually larger than can be expected from theoretical considerations (see Garratt and Segal 1988). This suggests an inconsistency in the observations, so it is rather difficult to draw firm conclusions.

The diurnal evolution of the temperatures up to 200 m are shown in Fig. 2. The skin temperature, the $2m$ temperature and the temperature at the lowest model level follow each other closely as they do in the observations. The main difference between model and observations is the amplitude of the diurnal cycle near the surface and the amount of decoupling between the two lowest model levels. They correspond to level 30 and 31 in the ECMWF model with heights of about 150 and 32 m above the surface respectively. The night time temperature drop is too large in the model and the coupling between model level 30 and 31 is too weak. The latter is particularly true for the period between day 326 and 332, where the lowest model level is completely decoupled from the layer aloft.

The observed near surface soil temperatures (2 cm deep) are shown in Fig. 3a together with the model soil temperatures (layer 1: 0-7 cm). The observations clearly show the "thermal barrier" at about $0^o\,C$ whereas the model soil temperatures keep dropping continuously.

The last term in the surface energy budget is the ground heat flux which is compared to observations in Fig. 3b. The observed ground heat flux is typically $-20\ W\,m^{-2}$, with a decrease to nearly 0 on sunny days. The ground heat fluxes in the model are much higher, typically of the order of $-50\ W\,m^{-2}$ during the night. The high ground heat flux is believed to be the result of too low temperatures near the surface rather than the cause of the temperature bias. A reduction of the ground heat flux, e.g. through reduction of soil thermal conductivity, would probably increase the temperature bias.

Although the momentum boundary layer is beyond the scope of this paper, it is still important to consider wind speed and wind shear because it drives the mixing in the stable boundary layer (e.g. Smedman et al. 1995 and Kim and Mahrt 1992). Beljaars and Betts (1993) compare model wind profiles with Cabauw observations and conclude that the momentum boundary layers are too thick and that the model profiles show too little shear. However, the main feature in this intercomparison is the large "unexplained variability" in the observations. Such variability, which is not represented by the model, may be relevant to the mixing processes (see e.g. Derbyshire, 1995a,b).

It is clear that for the Cabauw location, the night time cooling as well as the seasonal cooling are overestimated by the model and that the lack of soil moisture freezing in the model is playing a role. However, in spite of the detailed observations, it is difficult to draw firm conclusions about the performance of the stable boundary layer scheme. The main difficulty is the lack of closure in the observed surface energy balance, which is typical for most studies particularly at low wind speeds (Mahrt 1998). The fact that the near surface temperature drifts must mean that either the downward turbulent heat flux is too large, or that the cooling through sensible heat is distributed over too shallow a layer. The latter must be the case because there is no way the sensible heat flux can be reduced in the model: the ground heat flux is already too large, the net radiation is reasonable and the dew deposition is limited by theoretical considerations. The only conclusion that can be drawn is that the thermal boundary layer in the model is too shallow.

3 New model formulation: Soil moisture freezing

In high and mid-latitudes the phase changes of water in the soil have an important effect on the water and energy transfer in the soil. In common with most current GCM's, the surface model introduced operationally at ECMWF in August 1993 (Viterbo and Beljaars 1995) did not consider this effect. A proper consideration of the solid phase of soil water would introduce modifications including, in order of importance: a) The thermal effects related to the latent heat of fusion/freezing (e.g. Rouse 1984); b) Changes in the soil thermal conductivity due to the presence of ice (e.g. Penner 1970); c) Suppression of transpiration in the presence of frozen ground (e.g. Betts 1998); and d) Soil water transfer dependent on a soil water potential including the effects of frozen water (e.g. Lundin 1989). In the modifications described here, only the latent heat effects will be considered. The main impact will be to delay the soil cooling, in the beginning of the cold period, and to delay the soil warming in spring, although the latter effect is less important because it occurs when the solar forcing is significant. Both effects make the soil temperatures less responsive to the atmospheric forcing and damp the amplitude of the annual soil temperature cycle.

The soil energy equation in the presence of soil water phase changes can be written as

$$(\rho C)_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial t}\left(\lambda_T \frac{\partial T}{\partial z}\right) + L_f \rho_w \frac{\partial \theta_I}{\partial t}\,, \tag{1}$$

where $(\rho C)_s$ is the volumetric soil heat capacity, T is the soil temperature, t is time, z is the vertical coordinate, λ_T is the thermal conductivity, L_f the latent heat of fusion, ρ_w water density, and θ_I the volumetric ice water content. Without loss of generality, for the grid squares characteristic of NWP models it can be assumed that

$$\theta_I = \theta_I(\theta, T) = f(T)\theta\,, \tag{2}$$

where θ is the total soil water content (liquid + ice), and

$$\begin{aligned} f(T) = 0 &\quad T > T_1\,, \\ 0 < f(T) < 1 &\quad T_2 \le T \le T_1\,, \\ f(T) = 1 &\quad T < T_2\,, \end{aligned} \tag{3}$$

where T_1 and T_2 are characteristic temperatures limiting the phase change regime. In reality, the values of T_1 and T_2 and the function $f(T)$ are a complicated function of soil texture and composition (see e.g. Williams and Smith 1989), but here they will be approximated in a simple way. For an idealized homogeneous, one-component soil, $f(T)$ would be a step-function. The physical reasons for having an interval over which melting/freezing is active, rather than a threshold temperature, include (Williams and Smith 1989): a) Adsorption, resulting from forces between the mineral parts of the soil and the water; b) Capillarity, related to the fact that the water free surface is not plane; c) Depression of the freezing point due to the effect of dissolved salts; and d) Soil heterogeneity.

To avoid an undesirable coupling between the temperature and water equations in the soil, Eq. (2) is simplified to

$$\theta_I = f(T)\theta_f\,, \tag{4}$$

where θ_f is a constant, representing the amount of soil water that can be frozen (thawed). For simplicity θ_f is set to $C_v\theta_{cap}$, with C_v and θ_{cap} the vegetation fraction and the soil water at field capacity respectively (the scaling with C_v was the simplest way of distinguishing between dry and wet areas). Combining Eq. (4) with Eq. (1) results in

$$\left[(\rho C)_s - L_f\rho_w\theta_f\frac{\partial f}{\partial T}\right]\frac{\partial T}{\partial t} = \frac{\partial}{\partial t}\left(\lambda_T\frac{\partial T}{\partial z}\right)\,, \tag{5}$$

showing that the effect of freezing can be interpreted as an additional soil heat capacity, sometimes referred in the literature as the "heat capacity barrier" around freezing. Comparison of the two terms at the left hand side of Eq. (5) leads to another interesting physical interpretation which quantifies the effect of soil water phase transitions. The energy necessary to freeze or melt all the water in one unit of volume is $L_f\rho_w\theta_f$. If this energy is used to heat or cool the same volume without considering freezing or melting, the soil temperature will change by $L_f\rho_w\theta_f/(\rho C)_s$ which is about 50 K for the operational parameters of the ECMWF model with $C_v = 1$. So, not considering the process of soil water freezing/melting can lead to very large artificial temperature changes that do not occur in nature if sufficient soil water is available.

Function $f(T)$, as tested in the ECMWF model, is

$$\begin{array}{lcl} f(T) = & 0 & T > T_1\,, \\ f(T) = & 0.5\left[1 - sin\left(\frac{\pi(T-0.5T_1-0.5T_2)}{T_2-T_1}\right)\right] & T_2 \le T \le T_1\,, \\ f(T) = & 1 & T < T_2\,, \end{array} \tag{6}$$

with $T_1 = T_0 + 1\ K$, $T_2 = T_0 - 3\ K$, and T_o the freezing point of water. Fig. 4 displays Eq. (6) (left axis) together with the ratio of the apparent heat capacity (the sum of the two terms in brackets in Eq. (5)) divided by the soil heat capacity (the first term only) (right axis).

4 Interaction of the surface with the stable boundary layer

The temperature drift problem as occurring in the ECMWF model is very much related to the coupling between the surface and the stable boundary layer parametrization. The surface layer is not considered here because the diagnostics presented in section 2 suggest that the problem is above the surface layer.

In the ECMWF model, turbulent diffusion above the surface layer is parametrized using the eddy diffusivity concept, where for stable situations the eddy diffusion coefficient is expressed in resolved parameters as

$$K_{m,h} = l_{m,h}^2 |d\vec{V}/dz| F_{m,h}(Ri) \,, \tag{7}$$

$$l_{m,h}^{-1} = (kz)^{-1} + \lambda_{m,h}^{-1} \,, \tag{8}$$

where $|d\vec{V}/dz|$ is the wind shear, $l_{m,h}$ are the turbulent length scales for momentum and heat, z is the height above the surface, k the Von Karman constant (0.4), $\lambda_{m,h}$ are the asymptotic length scales and $F_{m,h}$ are the prescribed stability functions of Richardson number $Ri = (g/\theta_v)(d\theta_v/dz)|d\vec{V}/dz|^{-2}$, with θ_v for virtual potential temperature. The asymptotic length scales $\lambda_{m,h}$ are 150 and 450 m for momentum and heat respectively in the ECMWF model, but these length scales are of secondary importance over the lowest few hundred metres of the atmosphere where the length scales $l_{m,h}$ are dominated by the height above the surface rather than by $\lambda_{m,h}$.

Two different forms of the stability functions are shown in Fig. 5. The first one (Eqs. 9-10) was operational in the ECMWF model between 1982 and 1995 (Louis, Tiedtke and Geleyn 1982, hereafter referred to as the LTG scheme) and is the model version that shows the near surface temperature drift in combination with the new land surface scheme introduced in 1993 (Viterbo and Beljaars 1995):

$$F_m = \frac{1}{1 + 2bRi(1 + dRi)^{-1/2}} \,, \tag{9}$$

$$F_h = \frac{1}{1 + 3bRi(1 + dRi)^{1/2}} \,, \tag{10}$$

where $b = 5$ and $d = 5$. An important feature of this closure is the strong momentum diffusion combined with weak heat diffusion at high Ri numbers. These functions are not compatible with observed Monin Obukhov stability functions as e.g. reviewed by Högström (1988). Observationally based stability functions drop off more quickly and have very low diffusion coefficients above say $Ri = 0.5$ (see Beljaars and Viterbo, 1998 for simulations with MO based functions). On the other hand it has to be emphasized that the existing observations are mainly in fully turbulent conditions over flat homogeneous terrain and become increasingly uncertain above $Ri = 0.2$. Effects of heterogeneity, intermittency and any sub-grid variability (i.e. due to katabatic flow or surface heterogeneity) are not represented by the MO based stability functions.

The second closure (Eqs. 11-12) is a modified Louis et al. formulation (hereafter called "revised LTG scheme") with less discrepancy between momentum and heat diffusion:

$$F_m = \frac{1}{1 + 2bRi(1 + dRi)^{-1/2}} \,, \tag{11}$$

$$F_h = \frac{1}{1 + 2bRi(1 + dRi)^{1/2}} \,, \tag{12}$$

with $b = 5$ and $d = 1$. The resulting Prandtl numbers (Pr) are more in line with the recent compilation of observations by Kim and Mahrt (1992) but the diffusion of momentum and heat is still considerably stronger than can be supported by observations (see Kim and Mahrt, 1992). This model change, which was proposed to cure part of the temperature drift problems, is a very pragmatic one and inspired by the following considerations: (i) the revised formulation has very little impact on the surface momentum fluxes over land and therefore hardly affects the model's large scale performance, (ii) the ratio of momentum and heat diffusion is reduced which is believed to be necessary in view of the material presented in section 2 (i.e. a too thick momentum boundary layer and too weak heat diffusion and also more consistent with data published by Kim and Mahrt 1992) , and (iii) the heat diffusion towards the surface is increased which increases the coupling between the atmosphere and the surface and reduces the amplitude of the diurnal temperature cycle.

To illustrate the characteristics of the boundary layer scheme, idealized one column simulations are used with the two different closure functions. A layer with uniform potential temperature and a constant geostrophic wind is cooled at the surface with a constant heat flux for a period of 9 hours. The time evolution of the wind and temperature profiles provides information about the bulk characteristics of the scheme i.e. how much stress a scheme produces and how it distributes the cooling in the vertical. Geostrophic drag laws have already been derived from the ECMWF parametrization (Beljaars 1995); this paper concentrates on the heat transfer aspects of the scheme.

Using a heat flux condition at the surface is not very common for single column stable boundary layer simulations. Very often, a cooling rate is prescribed near the surface as a boundary condition (e.g. Delage 1996), but in such experiments temperature drift can not occur, because the near surface temperature is prescribed. In the current study the heat flux is prescribed, which highlights the feedback on the temperature drift.

A series of idealized single column simulations has been made with the two forms of the stable boundary layer parametrization (Eqs. 9-10 and 11-12 respectively). The single column model performs a time integration for the Ekman layer with a Coriolis term, a pressure gradient term and a turbulent diffusion term for a period of 9 hours. The Coriolis parameter is set to $10^{-4}\ s^{-1}$ and a constant geostrophic wind of 10 $m\,s^{-1}$ (in the x-direction) is used. The initial conditions are: a uniform wind profile equal to the geostrophic wind and a uniform potential temperature profile of $20^o\ C$. Profiles of potential temperature and kinematic fluxes are shown in Fig. 6 for an integration with a constant downward heat flux of 25 $W\,m^{-2}$ applied as boundary condition at the surface. The operational vertical resolution with model levels at 32, 150, 350, 600 ... m above the surface is used for these experiments. The LTG scheme and the revised LTG scheme are very similar in the wind profiles and the momentum flux profiles; both schemes produce a boundary layer which is about 500 m deep. This is believed to be too deep (200 m would be more realistic; see e.g. Zilintikevich 1972 and Nieuwstadt 1981), but schemes with shallow stable boundary layers tend to have less drag at the surface which is detrimental to performance of the current ECMWF model (see Beljaars 1995).

The potential temperature profiles as simulated with the two schemes are rather different (Fig. 6c). The profile with the LTG scheme shows a shallow temperature boundary layer whereas the revised LTG scheme distributes the cooling over a much deeper layer and therefore reduces the temperature drop at the surface. The effect is even more clear in a simulation with 50 $W\,m^{-2}$ downward heat flux at the surface as shown in Fig. 6d. Comparison of Figs. 6c and 6d shows another interesting aspect of the stable boundary layer namely the positive feedback on the near surface cooling. A doubling of the heat flux at the surface increases the temperature drop at the lowest model level from 4.5 to 14 degrees with the LTG scheme and from 3 to 8 with the revised Louis scheme. This nonlinear behaviour is due to a decrease of mixing with increased stability. The more cooling that is applied to the stable boundary the less heat flux it can sustain leading to a strong drop in temperatures in a very shallow layer. This "run away" character of temperatures is related to the limited heat flux the stable boundary layer can sustain and is a well known feature of the stable boundary layer, although its quantitative characteristics are not well documented (Derbyshire 1994, 1995a,b). When the boundary layer becomes extremely stable (e.g. at low winds), turbulence cannot maintain the heat flux any longer and the temperature at the surface will be determined by a balance between radiative cooling and ground heat flux. This is a situation that is believed to occur too often in the current operational model leading to an excessive cooling of the soil and an accumulation of temperature errors in the soil.

The positive feedback as described above can also amplify other model deficiencies. An example of such a deficiency is the missing freezing of soil moisture, which allows the soil temperatures to drop below zero too quickly. This model deficiency makes the stable boundary layer too stable resulting in less downward sensible heat flux and therefore even more soil cooling.

Another aspect of the atmosphere land coupling in the ECMWF model is the skin layer conductivity. The skin layer temperature T_{sk} (radiative surface temperature) is linked to the temperature of the top soil layer T_{s1} by the following relation (Viterbo and Beljaars 1995)

$$G_0 = \lambda_{sk}(T_{sk} - T_{s1}), \tag{13}$$

where G_0 is the ground heat flux and λ_{sk} is the skin layer conductivity parameter. The value of λ_{sk} proposed by Viterbo and Beljaars (1995) is 7 $W\,K^{-1}m^{-2}$. The skin layer acts as an insulating layer on top of the soil and represents the effect of a vegetation or litter layer. It was introduced to limit the ground heat flux and to allow for an instantaneous response of the skin temperature to the radiative forcing (Betts et al. 1993). A recent analysis of data from the EFEDA experiment by Van Den Hurk and Beljaars (1996) suggests that the current operational value of λ_{sk} is too small for the EFEDA area with sparse vegetation. Also theoretical considerations for bare soil suggest that a value of 15 $W\,K^{-1}m^{-2}$ is more appropriate. The increase of λ_{sk} to 15 $W\,K^{-1}m^{-2}$ is included in the package of changes mainly to reduce the amplitude of the diurnal cycle. The skin layer conductivity is a highly empirical parameter which obviously depends on vegetation type soil type etc. A more advanced formulation of λ_{sk} will be considered in future work.

5 Seasonal integrations

Since the soil freezing and the excessive soil cooling act on a seasonal time scale, it would be preferable to test model changes with data assimilation over the seasons where freezing and melting occur (e.g. from October to May). In practice however, such testing is too demanding on computer resources. Therefore, as an alternative to data assimilation, long runs at T63 resolution are used with relaxation towards the operational analysis. The relaxation is applied on U,V,T and q from level 28 upwards (i.e. from about 600 m height upwards) with an additional term that is equal to the difference between the prognostic model field and the corresponding reference field divided by a time constant of 12 hours. The reference fields for U,V,T and q are obtained by linear interpolation in time between operational analyses every 24 hours. The 500 hPa fields from the relaxation runs have errors (compared to the analysis) that are typical for day 1 forecasts. No relaxation is applied at the lowest 3 model levels in order to test the impact of changes to the stable boundary layer and the land surface scheme.

First of all a control experiment has been performed to see whether the near surface temperature problems were reproduced by this model configuration. The experiments started from 1 October 1995 for a period of 4 months with postprocessing of the 12 UTC fields every 24 hours. It turns out that the short range 2 m temperature errors of the operational model are well reproduced by the relaxation integrations. Also the soil temperature evolution is very similar to that in the operational model. These results are confirmed on a global scale by plots of monthly mean differences (not shown) between relaxation runs and operational analyses.

The impact of the revised LTG scheme without and with soil moisture freezing are shown as time series in Fig. 7a for the German area, where soil temperature observations are available. During the cold spells with the severe cold bias in the control experiment, both the revised LTG scheme and soil moisture freezing contribute equally to the improvement of the 2 m temperature. The improvement on the soil temperature evolution (Fig. 8a,b) comes predominantly from the freezing scheme. Time series for Russia (Fig. 7b) show a similar improvement for temperature at 2 m. Fig. 9 shows the impact of the two schemes on the 2 m temperature as monthly mean differences (12 UTC) for January 1996. The revised LTG scheme increases monthly mean temperatures by 1 to 3 degrees over large continental areas of the Northern Hemisphere and over a considerable part of Antarctica. The effect of freezing leads to an additional warming of 2 to 3 degrees over cold continental areas except over Antarctica where the freezing has no effect.

The effect of the increased skin layer conductivity has also been tested with a long relaxation run, but no results are shown, because the influence is only on the amplitude of the diurnal cycle and no impact is seen on the seasonal evolution of the soil and near surface temperatures.

In order to see the evolution of the soil temperatures in spring, the long relaxation integrations were extended until the end of May 1996. Time series of the 2 m temperature over Germany and Russia are shown in Fig. 10, which are an extension of Fig. 7 with overlap for the month of January. The day time temperature bias at $2\,m$ has virtually been eliminated with the introduction of the revised LTD scheme and the soil freezing. The difference between the old and the new scheme disappears already at day 450 (25 March, see Fig. 10) over Germany, whereas the old scheme only recovers around day 485 (beginning of May) over Russia. The soil temperature evolution in Germany is shown in Fig. 11 for depths of 20 and 50 *cm*. The 20 *cm* and 50 *cm* soil temperatures are much better throughout the winter. After day 450 only small differences are seen between the two model versions at a depth of 20 *cm*, both being a few degrees too cold. This has virtually no impact on the air temperature forecasts. Deeper soil layers show the delay introduced with the thermal inertia from the soil freezing. The thawing of soil layer 3 (50 *cm*) has finished at about day 480 which is later than observed. However, in spring solar heating during day time is sufficient to maintain a good coupling between the atmosphere and the near surface soil temperatures in order to prevent the boundary layer from going into a permanent stable mode without substantial coupling with the surface.

To see whether the model recovers from the frozen soil in summer, the relaxation experiment ending on 31 May was extended as a normal long run without relaxation until the end of September. Monthly SST's from 1995 were used. The monthly mean temperatures of the four soil layers (0-7 *cm*, 7-28 *cm*, 28-100 *cm* and 100-289 *cm*) were compared with the operational ones of 1995. The differences are rather small for layer 1, 2 and 3. However, layer 4 is considerably colder than in operations which is to be expected, because the delayed warming due to soil thawing will be most noticeable in the deepest layer. The deep layer has a time scale of several months as shown by Viterbo and Beljaars (1995). The monthly mean temperature in September of soil layer 4 (from the experiment with the new model version) is shown in Fig. 12 and compared with climatological information on the permafrost areas. It shows that the areas that are at least partially frozen in the model correspond by and large to the areas that have sometimes permafrost. It means that the model results are rather realistic with probably slightly too much frost remaining in the soil during summer.

The multiyear drift of the model was checked by continuing the end of September soil fields with a second relaxation experiment from 1 October 1995 to 31 January 1996 (as if the same winter situation of 1995/1996 occurs two years in a row). The soil temperatures for January 1996 from the second experiment were compared with those from the first and showed very little drift.

The comparison with the soil temperatures over Germany and the slightly too large extent of the permafrost suggests that the revised model still extracts too much heat from the soil in winter in spite of the considerable improvement from the revised vertical diffusion scheme. The seasonal soil temperature evolution is also much more realistic, mainly due to the introduction of soil moisture freezing. However, the delayed thawing over Germany suggests that the seasonal amplitude of soil heat flux is probably still too large. Such a bias may be due to an overestimation of long wave cooling at the surface, to biases in the short wave radiation or to an underestimation of downward turbulent fluxes in stable situations. These aspects are subjects of ongoing research.

6 Short range forecasts

To allow a more careful inspection of the model performance in general and of the temperature errors in particular, a data assimilation experiment was started from 15 January 1996 at T106 resolution with 10 day forecast every day from 16 to 29 January. The initial soil temperature and soil moisture fields were interpolated from the T63 relaxation run. The forecast model has the revised LTG scheme , the soil freezing scheme and the increased skin layer conductivity. The SYNOP verification for Europe of the 60 and 72 hour forecasts averaged over the second half of January is shown in Fig. 13. The improvement is obvious with a mean bias reduction from -4^o to $-0.2^o C$ at night and from -4.1^o to $-0.9^o C$ at day time. The night time improvement is better than the day time improvement consistent with a reduction in the amplitude of the diurnal cycle. The latter is particularly beneficial over North Africa where a large diurnal cycle is present in January without any effects from soil freezing. The improvement of the amplitude of the diurnal cycle partially due to the revision in the stable diffusion scheme and partially due to the increased skin layer conductivity.

The near surface cold bias, as seen in winter in the operational model, is limited to a very shallow layer near the surface and its impact on the forecast of large scale weather systems is negligible. The model changes are therefore not expected to have impact on the large scale model performance, provided that the vertical diffusion change does not significantly affect the momentum fluxes. As explained in section 3, the vertical diffusion changes are such that the surface momentum fluxes are changed very little. The small impact is confirmed by the r.m.s of the 500 and 1000 hPa height errors (scores) for the Northern Hemisphere (not shown).

In order to see in more detail how the model handles the warming and soil thawing in spring, a short data assimilation experiment was run for April at T106 resolution. Again, the starting values for the soil fields were interpolated from the relaxation runs. The improvement is mainly in the night time temperatures as can be seen from the diurnal cycles in Fig. 14. The scores of the forecasts that were run from this experiment are not shown because the impact is small.

7 Discussion and conclusions

The ECMWF model that was operational between 1993 and 1996 had the tendency to produce a severe cold bias over continental areas in winter. This near surface cold bias was particularly severe over Europe in the winter of 1995/1996. It has been shown that the stable boundary layer scheme has an inherent positive feedback that amplifies a near surface cold bias: if the surface becomes too cold the boundary layer becomes more stable which reduces the downward heat flux, resulting in even more cooling at the surface. This positive feedback only works above a critical Richardson number and since the Richardson number contains the square of the wind speed, the effect will be most noticeable at low wind speeds. It is believed that the positive feedback is more pronounced in the operational model than in the real atmosphere and therefore the parametrization has been changed such that more heat diffusion is produced in stable situations. However, it is very disappointing that the stable boundary layer parametrization cannot be based directly on observational studies. One of the problems is the lack of understanding of the transport mechanisms in very stable situations i.e. at low wind speeds. The observational studies seem to suffer from a common deficiency namely that the surface energy balance cannot be closed. More research is certainly needed to clarify the way the atmosphere satisfies the surface energy balance. It is quite possible that mesoscale variability (i.e. subgrid flow meandering, katabatic flow and surface heterogeneity) plays a key role in the transport of heat and moisture towards the surface.

The process of soil freezing turns out to be an important damping mechanism on the seasonal temperature cycle which is clear from soil temperature observations. In winter the freezing prevents the boundary layer from becoming too stable and the introduction of this process in the model has a clear beneficial effect on the $2\,m$ temperature forecasts. Experimentation with the new revised diffusion scheme and the soil moisture freezing scheme, has shown that the atmospheric data can keep the soil temperature evolution under control when cycling through data assimilation and forecasting.

However, the soil temperatures are still systematically too low and the amount of heat that is extracted from the soil by the atmospheric model in winter is still believed to be too large. Various aspects are under investigation. One obvious reason for the excessive cooling is long wave radiation in the model version used for this study. It is well known that the clear sky radiative cooling is overestimated due to deficiencies in the description of the water vapor continuum and due to an underestimation of cloud cover over land (Gregory et al. 1998). There may also be a direct effect from the soil heat diffusion parametrization itself. The soil heat conductivity is a function of soil moisture content which makes the diffusion of heat more efficient in wet conditions (upward flux in winter) than in dry conditions (downward flux in summer), but in the current ECMWF scheme the dependence on soil moisture is probably too strong (Peters-Lidard et al. 1998). The impact of these model deficiencies are not clear at this stage but will be studied with help of the stand-alone model of the land surface scheme.

One of the remarkable results of this study is that relatively small changes in the stable boundary diffusion have a strong influence on the model's near surface winter climate. At the same time, the scientific knowledge of the very stable boundary layer is not sufficient to give adequate guidelines for parametrization (Mahrt 1998). The impact of soil moisture freezing and stable boundary layer diffusion on temperature is substantial but limited to a shallow layer near the surface and therefore the effect on the large scale flow is limited. Near surface temperature forecasts are obviously important for NWP, but surface temperatures are also important for the radiative balance of the atmosphere which is highly relevant for climate models.

The model improvements, discussed in this paper were introduced in the operational ECMWF system in September 1996. At the time of writing of this paper, operational verification results were available for the winters of 1996/1997 and 1997/1998. Fig. 15 shows the history of 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 and are different for night and day. They have a rich history related to the many model changes that were made over the years. Only a brief discussion is given here on the land surface developments that were made from 1993 onwards.

First of all, in August 1993 the Blondin (1991) scheme with a climatological deep soil boundary condition was replaced by the free running 4-layer scheme (Viterbo and Beljaars 1995), but the impact is not very obvious. The summer day time bias of August 1993 was smaller than that of 1993, but at that time the soil scheme had been running freely for 2 months only (including the July parallel test). 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 the field capacity over vegetated areas. A simple soil moisture analysis scheme was introduced in December 1994 (Viterbo 1995) with a clear beneficial impact on the day time temperature bias for summer 1995. The night time temperatures have been negatively biased 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 wind from the east and very cold temperatures. There is also evidence that the introduction of the major model upgrade in April 1995 (prognostic clouds scheme, mean instead of envelope orography and a new subgrid orographic drag scheme) had a negative impact on the winter time cold drift over land but it is not known which element of this model upgrade is responsible. It is interesting that the reduction of the day time warm bias actually increased the night time bias by moving the entire diurnal cycle to colder temperatures. Soil freezing, increased boundary layer diffusion and increased skin layer conductivity (as described in this paper) was introduced in September 1996 and improved the monthly error statistics considerably. The winter time drift was largely eliminated and the amplitude of the diurnal cycle is at a reasonable level. It is fair to point out that the results of Fig. 15 are averages 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 often related to errors in the cloud forecast and to the presence of snow.

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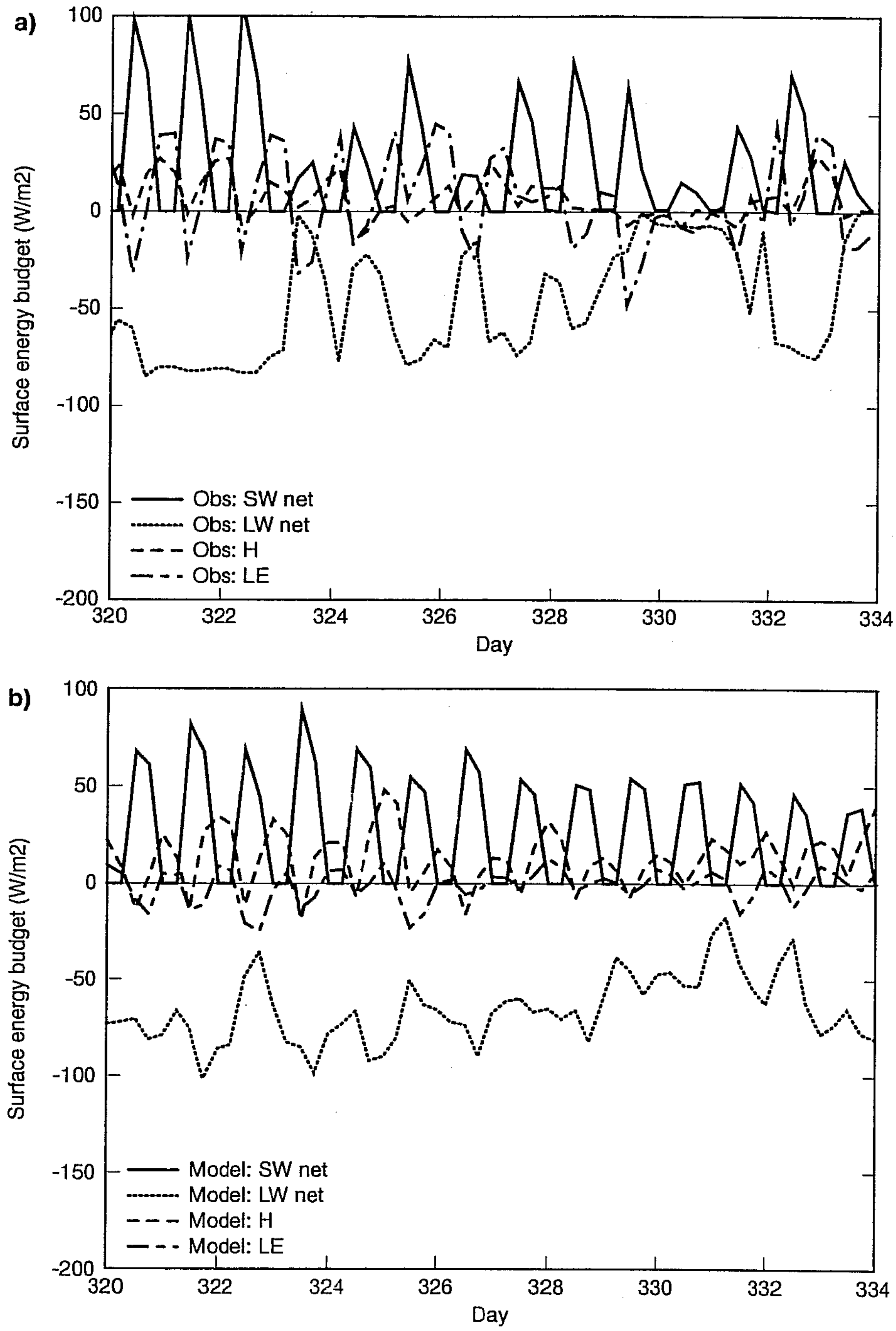


Figure 1: Surface energy budget at Cabauw in The Netherlands for the second half of November 1993. The upper panel shows observed 6 hour averages, the lower panel shows 6 hour averages of the operational 0 to 6, 6 to 12, 12 to 18 and 18 to 24 hour forecasts for successive days. The horizontal axis is the verifying day number. SW stands for short wave radiation, LW represent long wave radiation, H sensible heat flux and LE latent heat flux. Downward fluxes are positive.

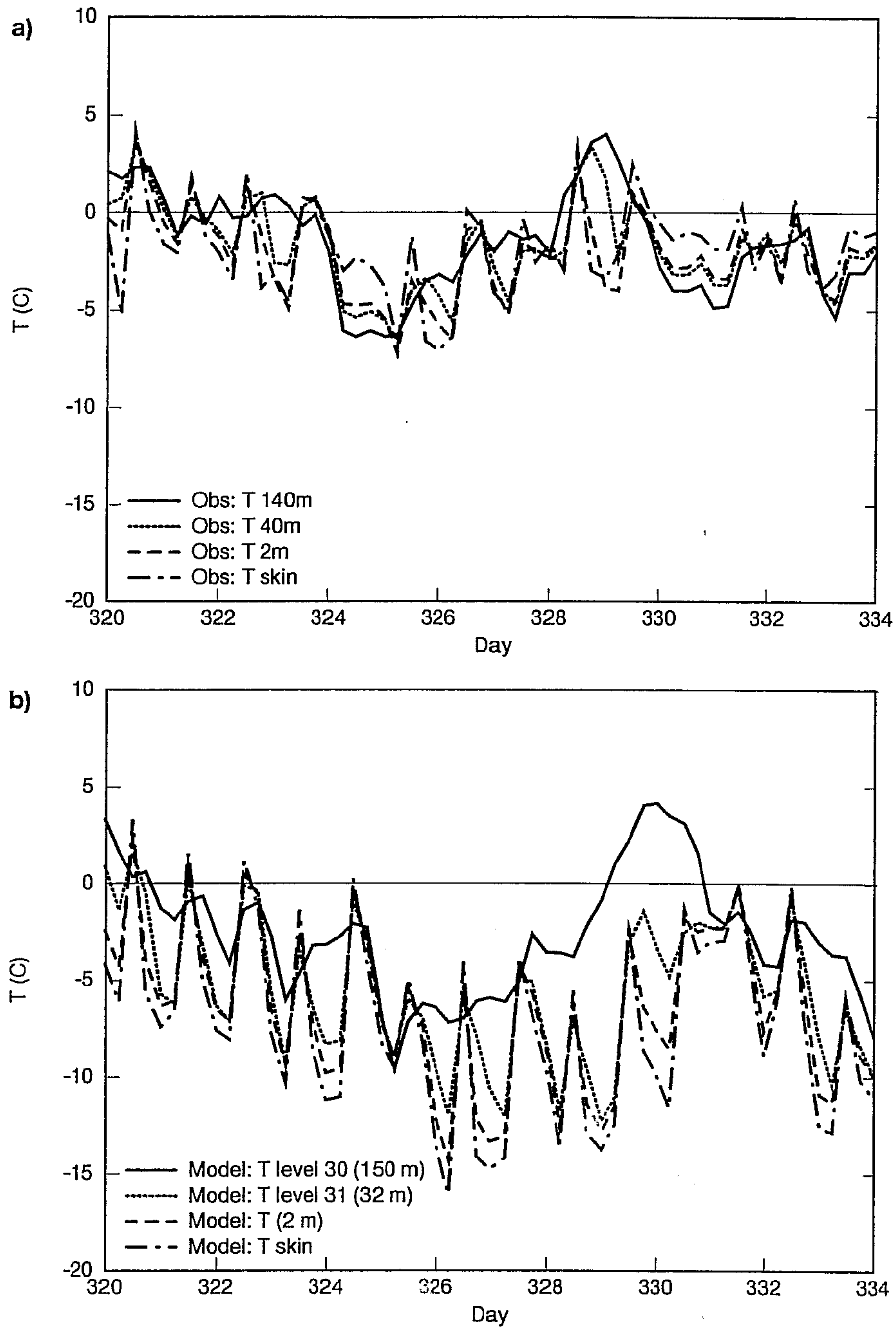


Figure 2: Boundary layer temperatures at Cabauw in The Netherlands for the second half of November 1993. The upper panel shows observed 30 minute averages of the skin temperature (radiative surface temperature), the $2\,m$, $40\,m$ and $140\,m$ temperatures with 6 hour intervals. The lower panel shows the 6, 12, 18 and 24 hour forecasts from successive days of the skin temperature, the $2\,m$, the level 31 ($30\,m$) and the level 30 ($140\,m$) temperatures.

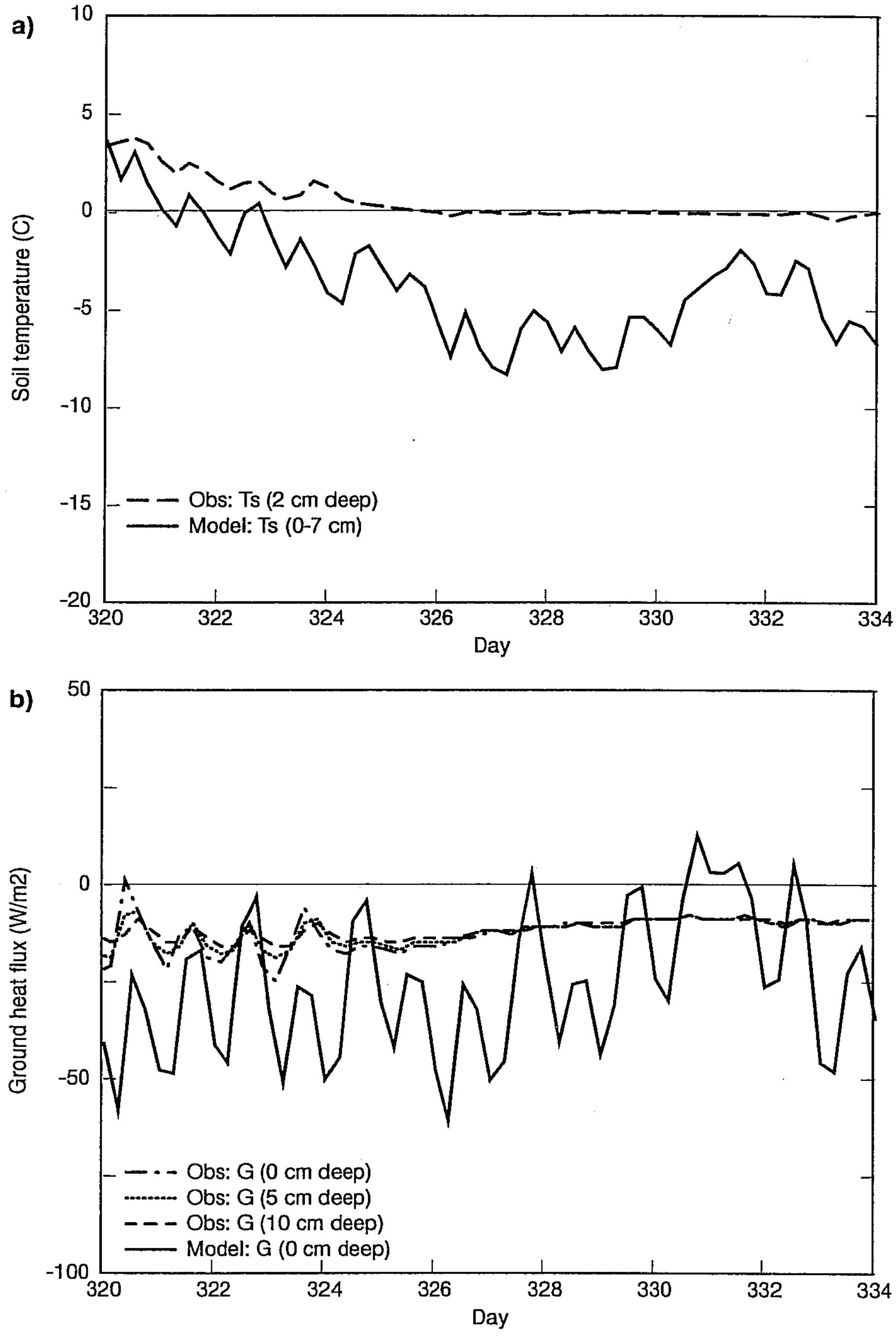


Figure 3: (a) Soil temperature at Cabauw in The Netherlands for the second half of November 1993. Observed 30 minute averages from a depth of 2 *cm* are compared with the model's top soil layer temperatures of the 6, 12, 18 and 24 hour forecast for successive days. (b) Ground heat fluxes at Cabauw in The Netherlands for the second half of November 1993. Observed 6 hour averages at depths of 0, 5 and 10 *cm* are compared with model averages of the 0 to 6, 6 to 12, 12 to 18 and 18 to 24 hour forecasts for successive days.

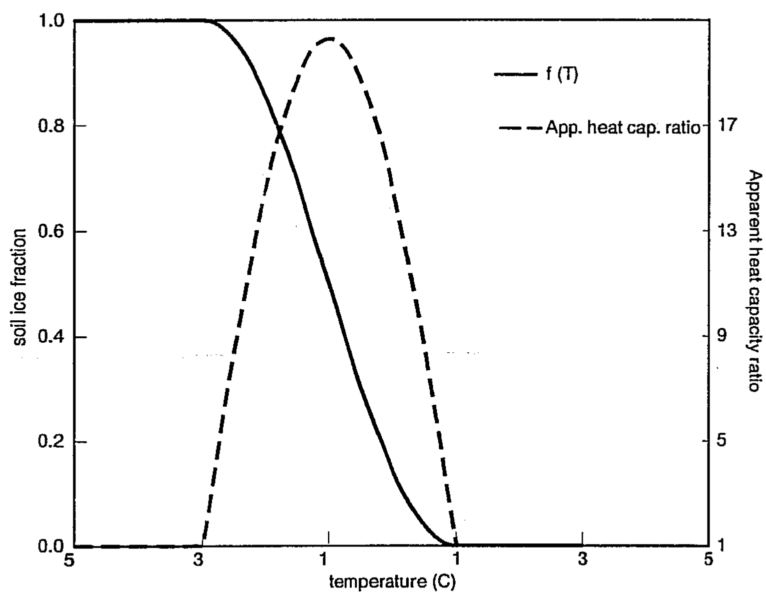


Figure 4: Fraction of water in the ice phase (left axis, solid curve) and ratio of apparent heat capacity to soil heat capacity (right axis, dashed curve) as a function of temperature.

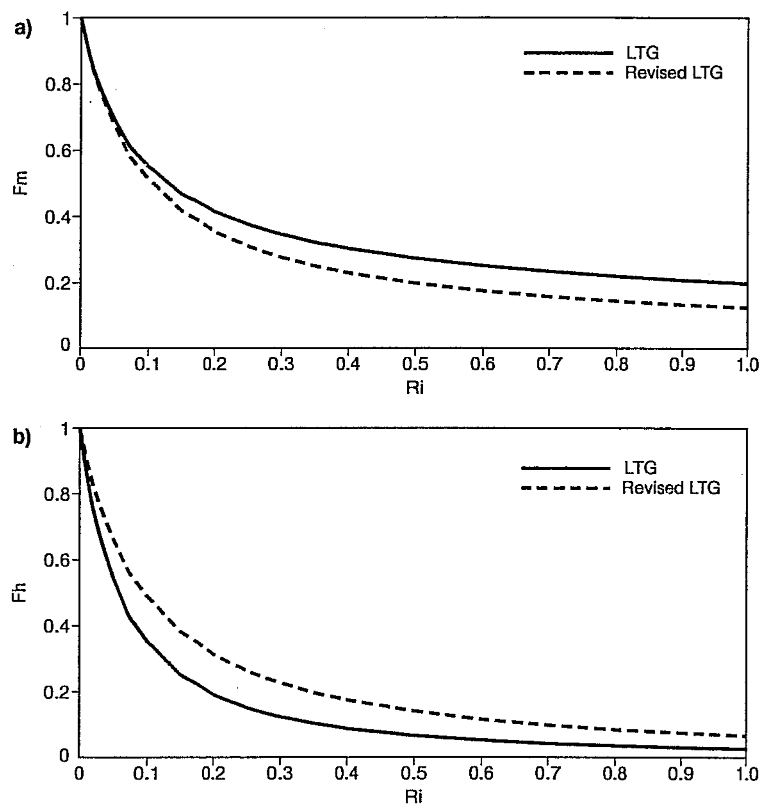


Figure 5: Two different forms of the stability functions for momentum (upper panel) and heat (lower panel) for positive Richardson numbers.

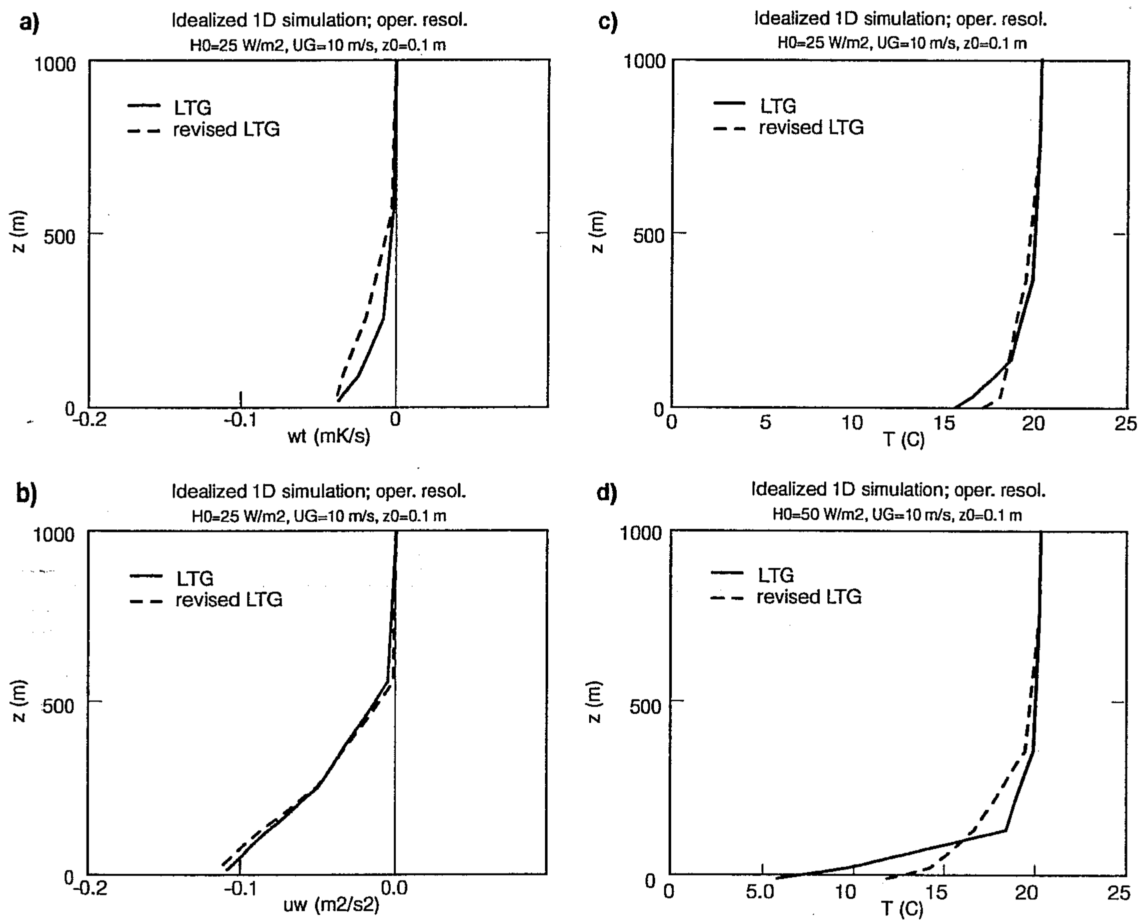


Figure 6: Single column simulations where a neutral boundary layer is cooled by a downward surface heat flux of 25 W/m^2 over a period of 9 hours. The geostrophic wind is 10 $m\,s^{-1}$ in the x-direction and the surface roughness length is 0.1 m. Two different schemes are used: The LTG scheme (Louis et al. 1982) and the revised LTG scheme. Profiles of kinematic heat flux (a), kinematic momentum flux in the direction of the geostrophic wind (b), potential temperature (c) and potential temperature with a surface heat flux of 50 $W\,m^{-2}$ (d) are shown.

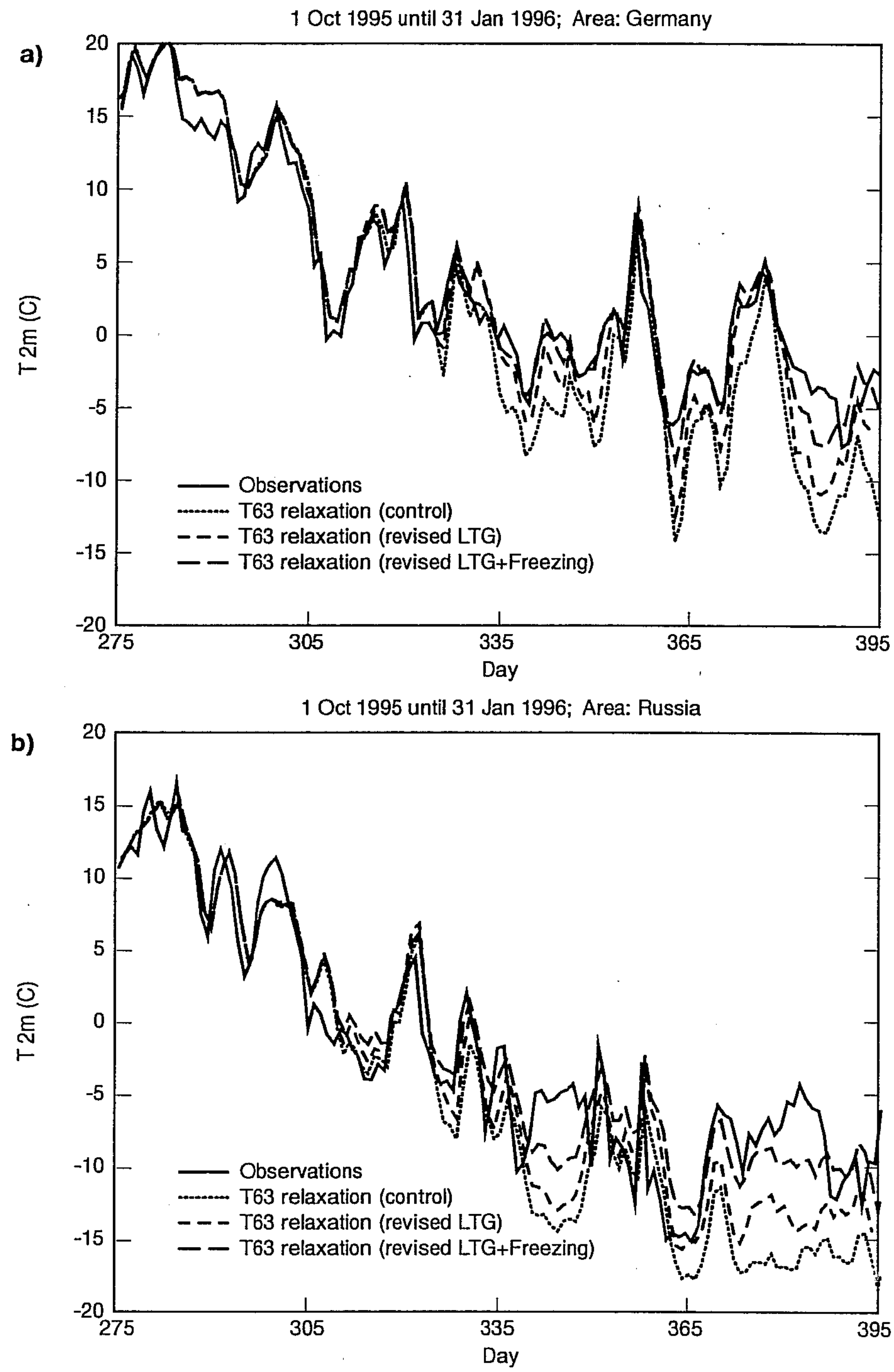


Figure 7: Time series of day time observations of $2\,m$ temperature over Germany and Russia in comparison with the results of T63 relaxation runs from 1 October 1995 (day 275) to the end of January 1996 (day 395).

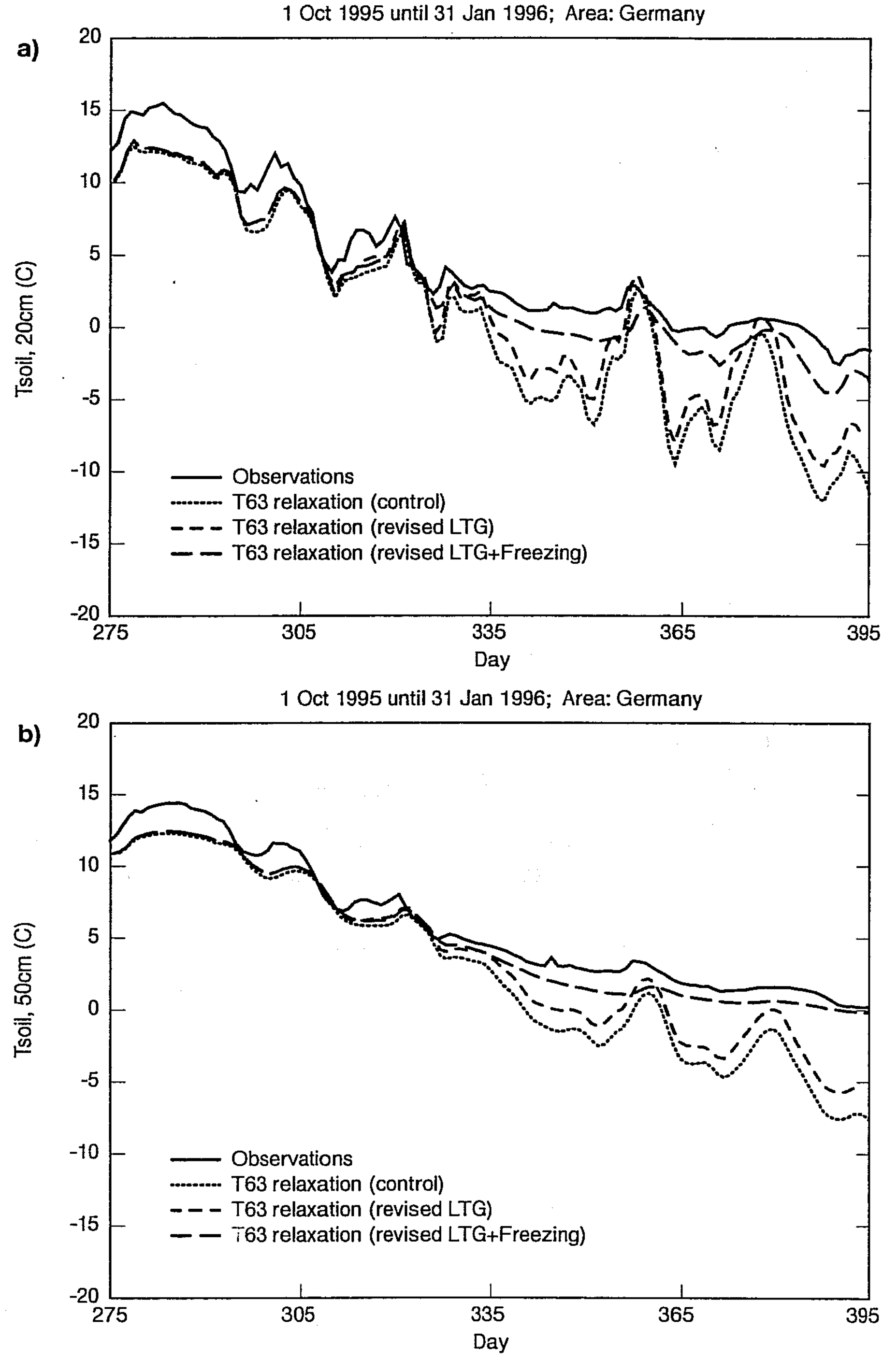


Figure 8: Time series of daily soil temperature observations in Germany at two depths (Figs. a,b) in comparison with the results of T63 relaxation runs from 1 October 1995 (day 275) to the end of January 1996 (day 395).

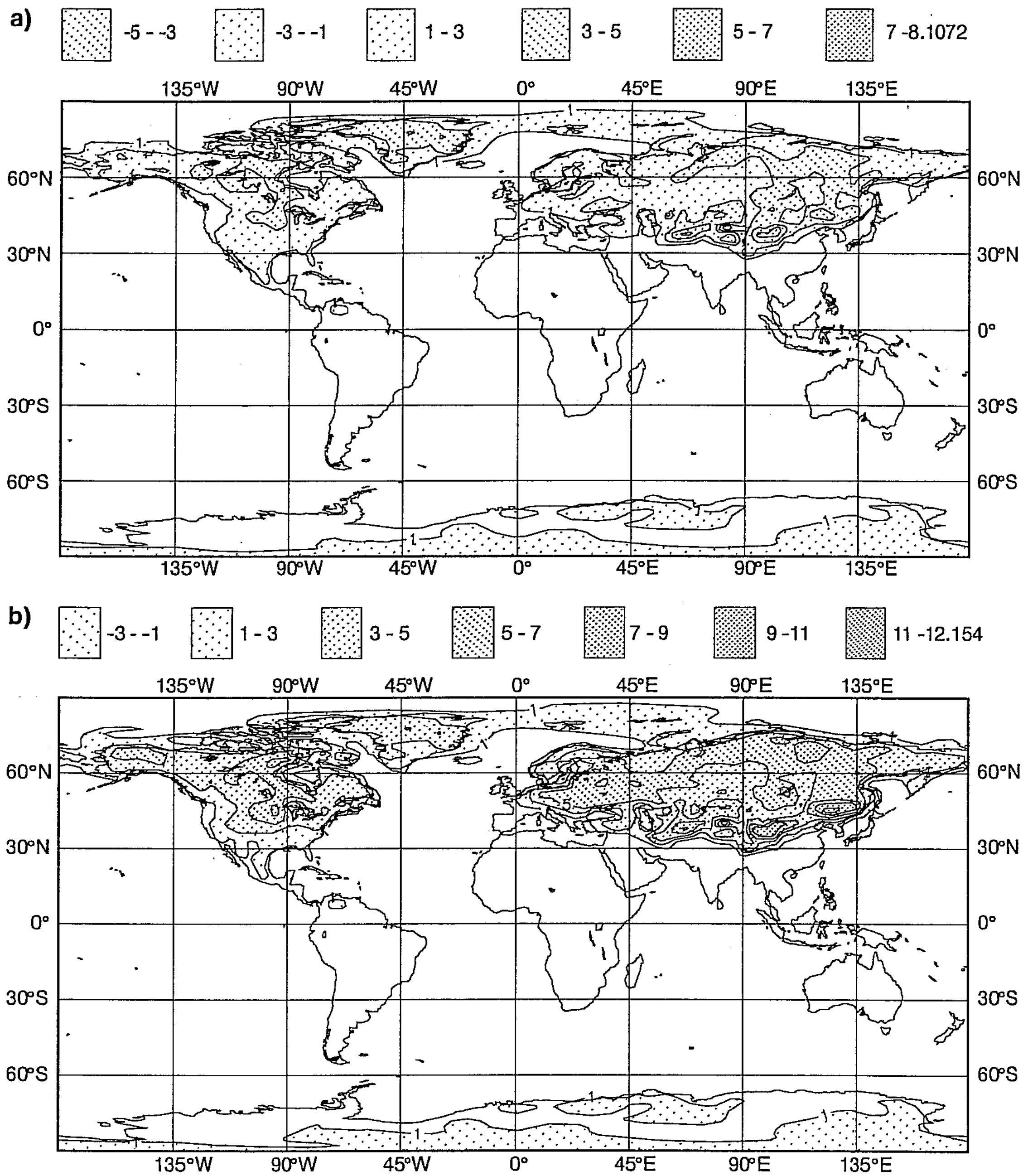
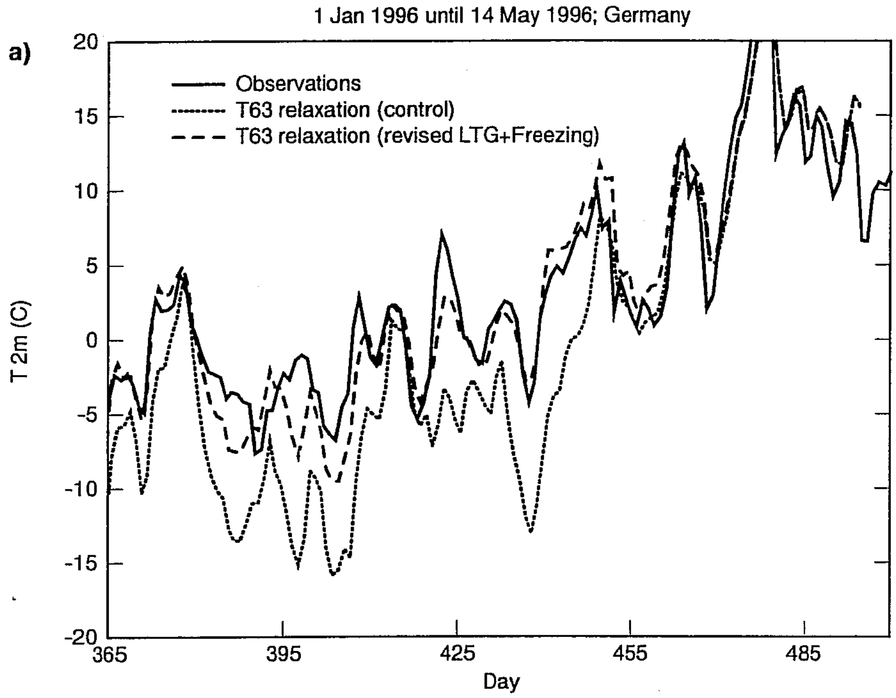


Figure 9: Mean difference in $2\,m$ temperature for the month of January 1996 between the T63 relaxation experiment with "revised LTG" and the control experiment (a) and between "revised LTG+freezing" and the control experiment (b). All long integrations start from 1-10-1995.



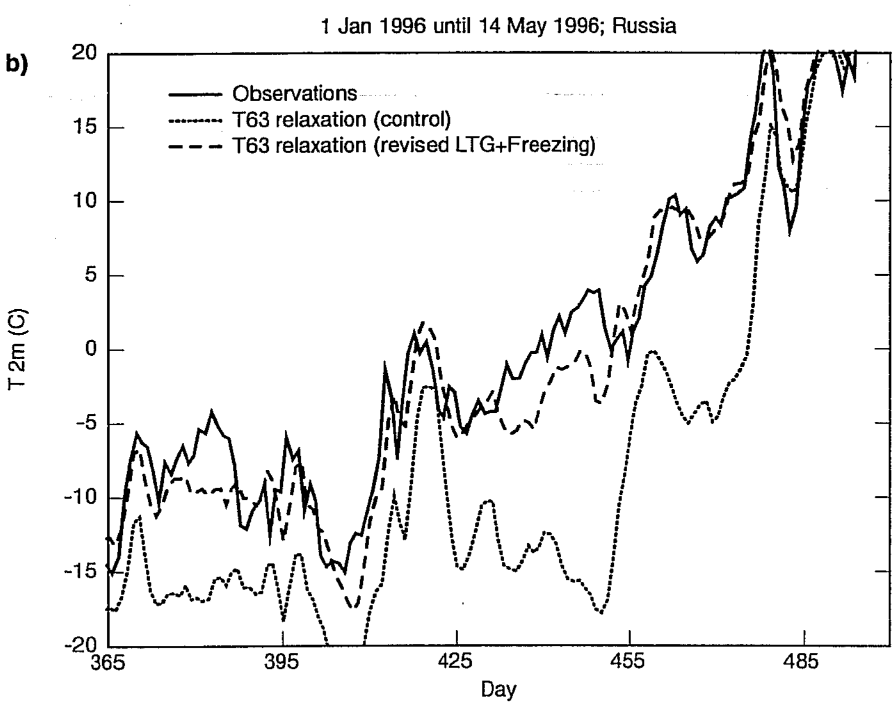


Figure 10: Time series of day time observations of $2\,m$ temperature over Germany and Russia in comparison with the results of T63 relaxation runs from 1 October 1995 to the end of May 1996 (day 517). These curves are a continuation of the ones in Fig. 7 with one month of overlap.

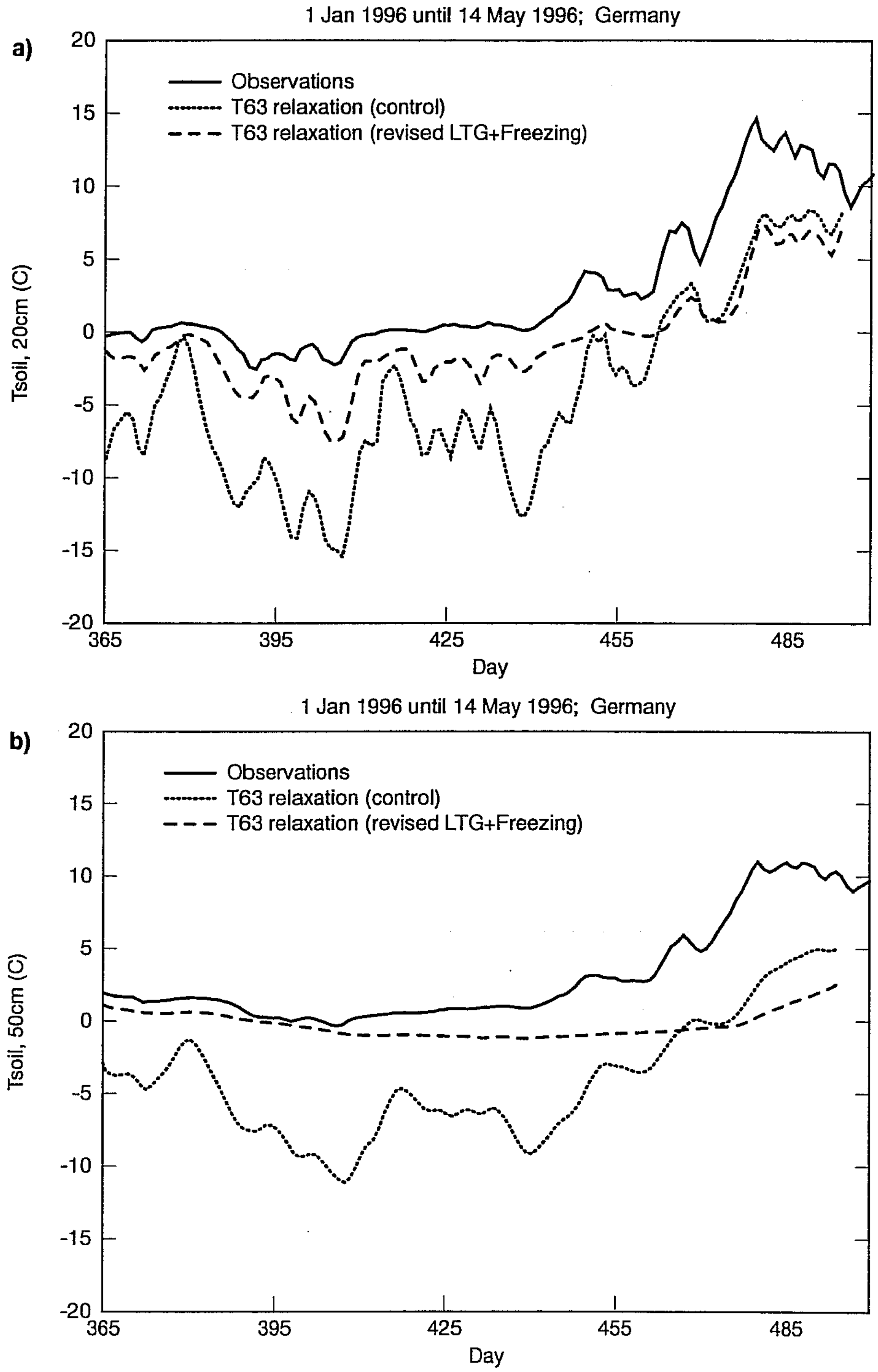


Figure 11: Time series of daily observations of soil temperatures in Germany in comparison with the results of T63 relaxation runs from 1 October 1995 to the end of May 1996 (day 517). These curves are a continuation of the ones in Fig. 8 with one month of overlap.

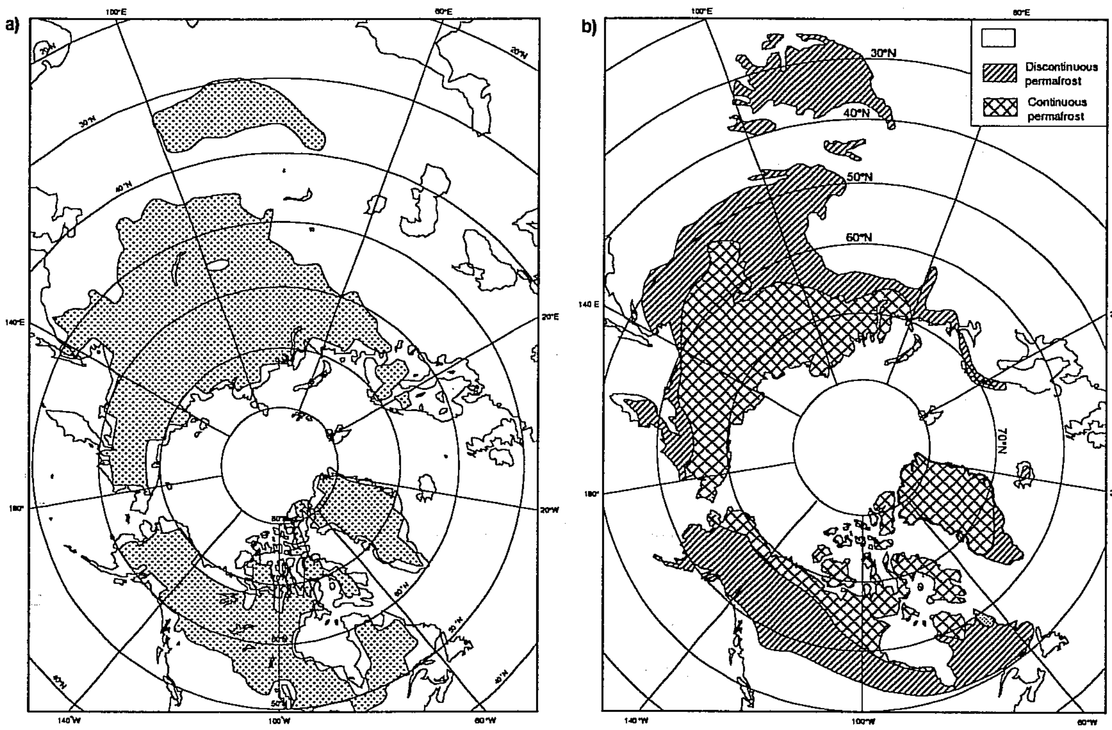


Figure 12: Frost coverage with heavy shading (Fig. a) at level 4 (between 1 and 2.89 m deep) averaged over the month of September 1996. These results were obtained from a relaxation run (including revised LTG+soil moisture freezing) starting from 1 October 1995 until 31 May 1996 and continued as a normal long run without relaxation until 30 September 1996 (with SST's from the year before). The heavily shaded areas over land have temperatures below $0^o\,C$. Fig. b reproduces the climatology of permafrost areas as published by Washburn (1980; see also Untersteiner 1984).

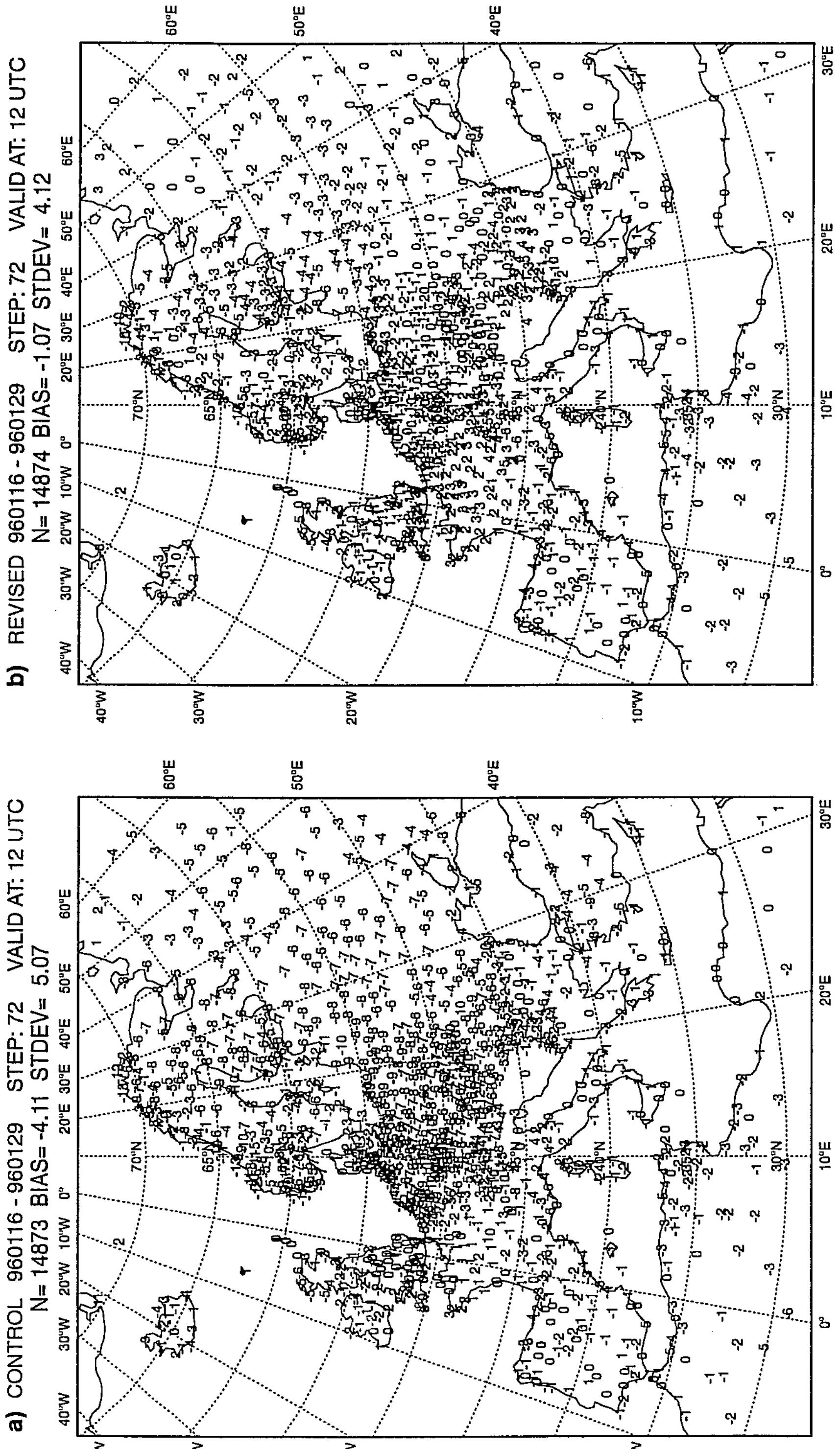


Figure 13: Temperature errors at the 2 m level averaged over 14 successive 72 (a,b) and 60 (c,d) hour forecasts from the data assimilation experiment with "revised LTG", "freezing", and "increased skin layer conductivity" (Figs. b,d) in comparison with the control experiment (Figs. a,c). The printed numbers are the mean errors at SYNOP stations for forecasts with initial dates between 16 and 29 January 1996.

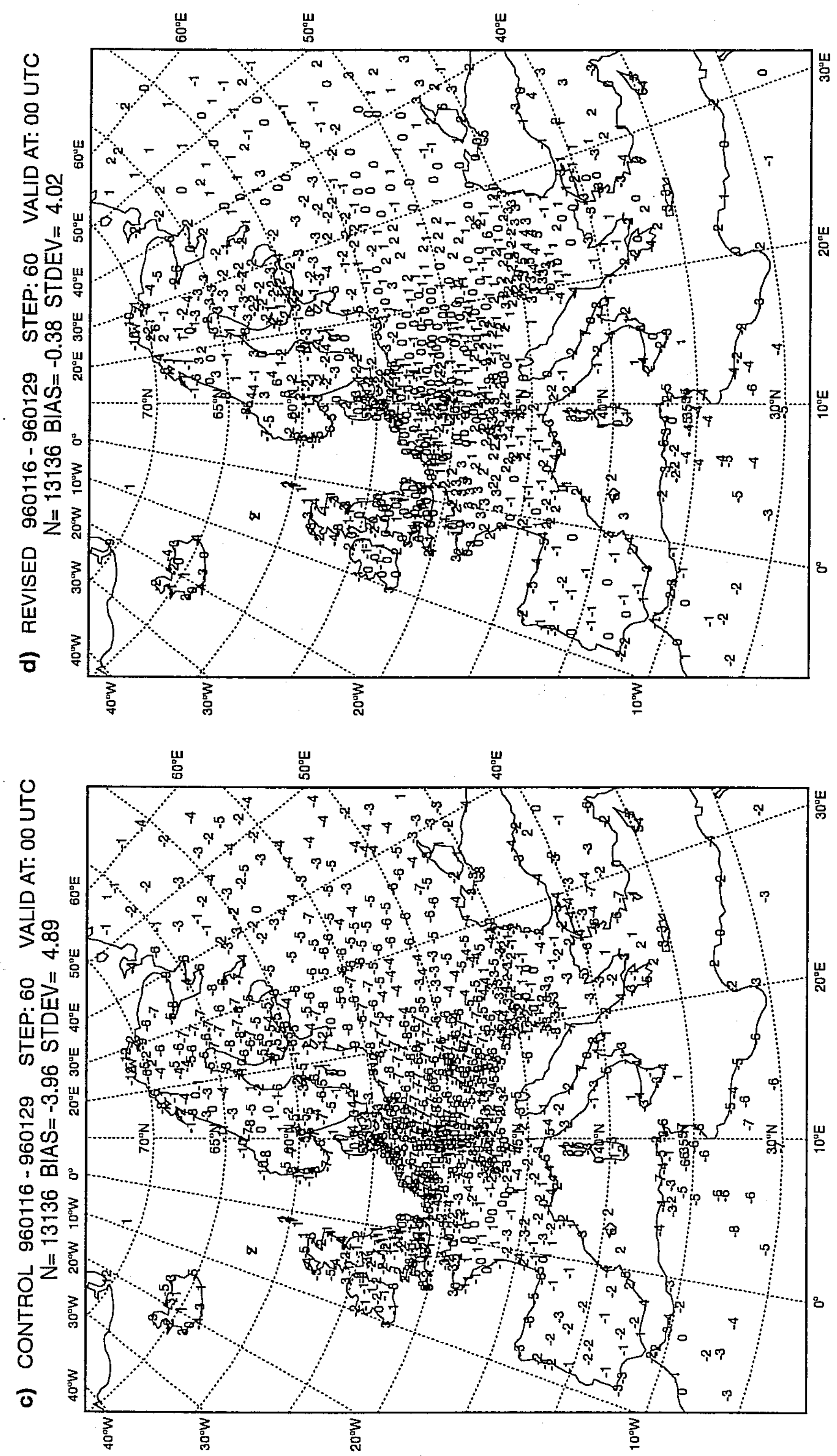

d) REVISED 960116 - 960129 STEP: 60 VALID AT: 00 UTC
N= 13136 BIAS= -0.38 STDEV= 4.02
c) CONTROL 960116 - 960129 STEP: 60 VALID AT: 00 UTC
N= 13136 BIAS= -3.96 STDEV= 4.89

Figure 13: (c,d)

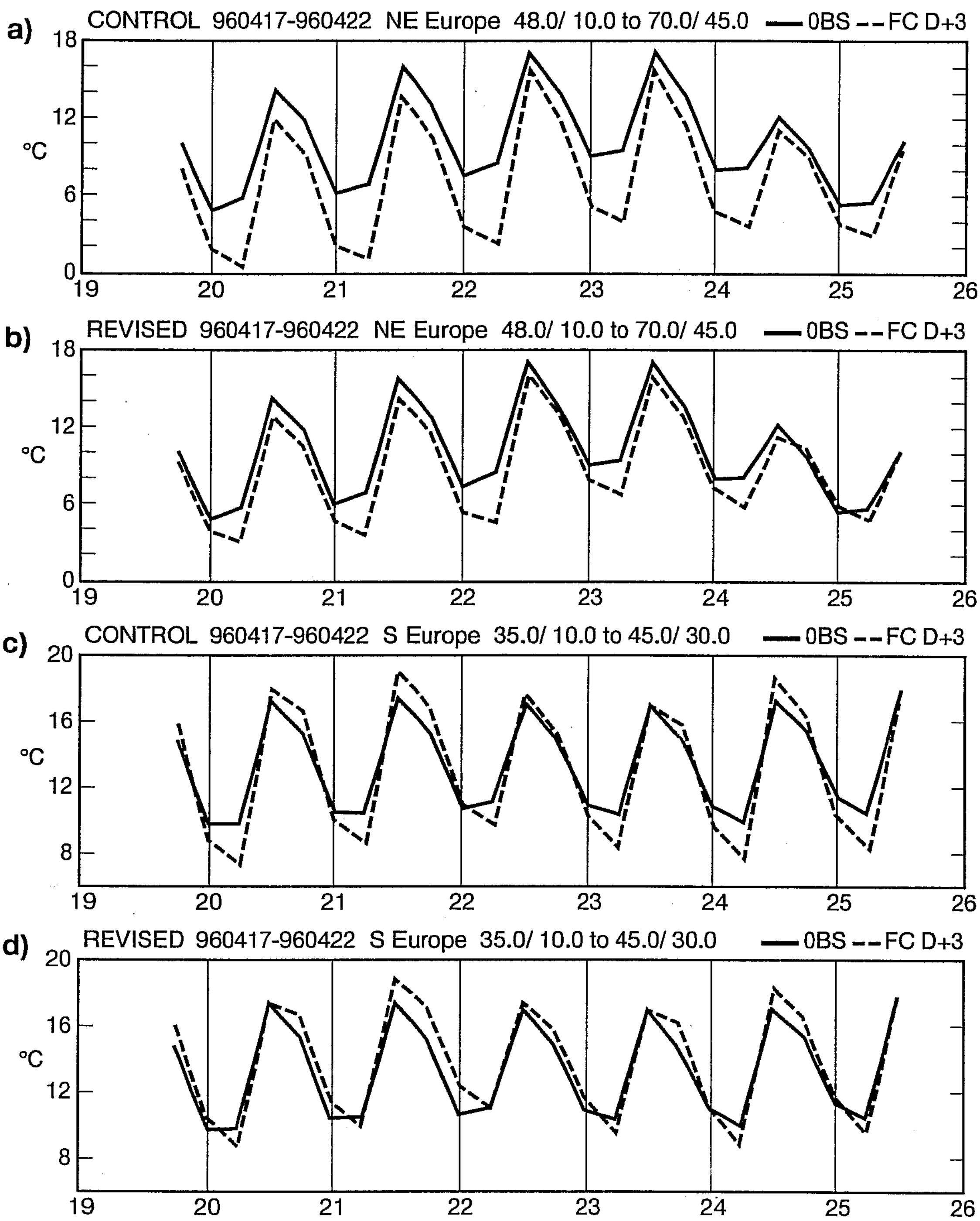


Figure 14: Diurnal cycles of 54, 60, 66 and 72 hour forecasts (dotted) in comparison with observations (solid lines) for North-East Europe (a,b) and Southern Europe (c,d). These are results from T106 data assimilation/forecast experiments. At the beginning of the data assimilation experiment the soil temperatures were initialized with temperatures from the T63 relaxation runs. Figs. a and c correspond to the control experiment, Figs. b and d correspond to the new model version with revised vertical diffusion, soil moisture freezing and increased skin layer conductivity.

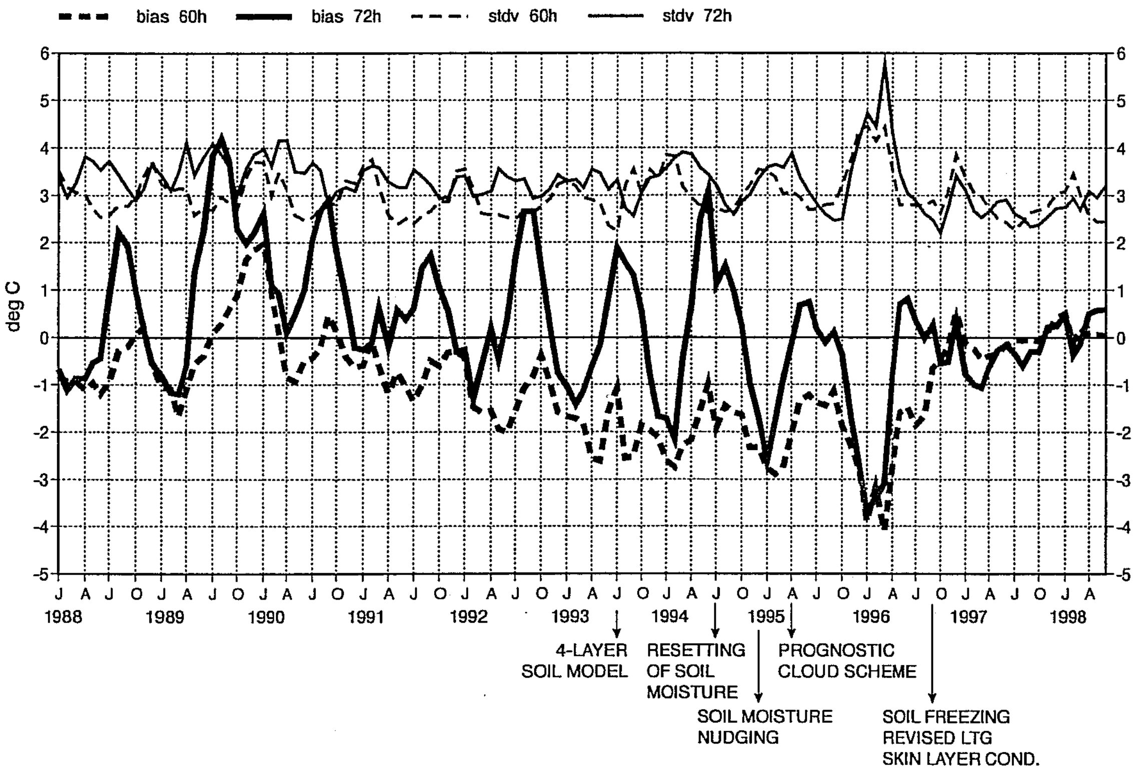


Figure 15: History of monthly bias and standard deviation (with respect to observations) of the day time (72 hour forecasts) and night time (60 hour forecasts) operational 2 m temperature forecasts, averaged over all available SYNOP stations in the European area of 30^o to $72^o\,N$ and $22^o\,W$ to $42^o\,E$.