# Issues in boundary layer parametrization for large scale models

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## 1. Introduction

A realistic representation of the boundary layer is an essential ingredient of state of the art numerical weather prediction (NWP) models for a number of reasons. First of all the boundary layer scheme provides the surface boundary condition for wind, temperature and moisture and controls as such the surface fluxes of momentum, heat and moisture. Secondly, it interacts with other parametrization schemes e.g. convection, cloud and land surface schemes. Finally, the boundary layer scheme provides important forecast products like wind, temperature and humidity at observation level, surface fluxes for wave and ocean models and first guesses (or observation operators) during data assimilation. Tendencies from the boundary layer scheme are typically large near the surface (in zonal averages of the order of 10 m $s^{-1}day^{-1}$, 5 K $day^{-1}$ and 5 g $kg^{-1}day^{-1}$ for wind, temperature and moisture respectively), so even in a short range forecast, substantial drift occurs when boundary layer effects are not represented (for impact in climate models see Garratt et al., 1993, Garratt, 1994).

The purpose of this paper is to review the different aspects of a boundary layer parametrization in large scale atmospheric models and to discuss unresolved issues that are important for models. It will be argued that models are particularly sensitive to turbulent mixing in areas where the gradients are large, e.g. in the surface layer, in inversion layers and in the stable boundary layer (Figure 1). From verification with ocean flux data it will be concluded that the interaction with boundary layer clouds is still a major issue; a substantial part of errors in surface fluxes turns out to be related to an unrealistic representation of boundary layer clouds. It is obviously difficult to define priorities in further work and therefore an attempt is made to identify key issues by considering uncertainty in the parametrization and sensitivity of the ECMWF model to different aspects of the boundary layer scheme. For much of the supporting material, the reader is referred to the earlier papers by Beljaars and Betts (1993), Beljaars (1995b), and Beljaars and Viterbo (1998).

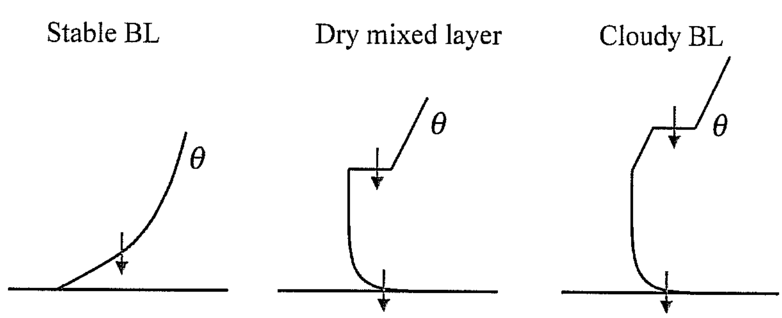


*Figure 1 Schematic potential temperature profiles for the stable boundary layer, mixed dry boundary layer and the cloud topped boundary layer.*

## 2. Surface layer

The surface layer is the layer where the fluxes are close to the surface values and where Monin-Obukhov (MO) theory applies. In large scale models it is usually applied between the surface and the lowest model level even if the lowest model level is not very close to the surface. This is justified, because it turns out that MO profiles are still reasonably accurate above the surface layer. The standard way of expressing surface fluxes of momentum ($u_*^2$; square of the friction velocity), sensible heat ($\overline{w'\theta'_o}$) and moisture ($\overline{w'q'_o}$) into wind, temperature and moisture differences over the surface layer is with help of transfer coefficients (Brutsaert, 1982):

$$u_*^2 = C_m |\vec{V}_1|^2$$
$$\overline{w'\theta'_o} = C_h |\vec{V}_1| (\theta_s - \theta_1)$$
$$\overline{w'q'_o} = C_q |\vec{V}_1| (q_s - q_1)$$

where $C_m$ is the transfer coefficient for momentum (drag coefficient), $C_h$ is the transfer coefficient for heat, $C_q$ is the transfer coefficient for moisture, $|\vec{V}_1|$ is the absolute horizontal wind speed at the lowest model level, $\theta_1$ and $q_1$ are potential temperature and specific humidity at the lowest model level, and $\theta_s$ and $q_s$ are potential temperature and specific humidity at the surface. In accordance with Monin Obukhov similarity theory, the transfer coefficients can be written in terms of profile functions containing a logarithmic part, with roughness lengths as surface characteristics, and a stability function describing the effect of stability as a function of the Obukhov length $L = u_*^3 / \{\overline{w'\theta'_{v\,o}} kg/\theta_v\}$, where $\theta_v$ represents virtual potential temperature and k (=0.4) the VonKarman constant (see Stull, 1988; Garratt, 1992). Three aspects need further specification to close the surface flux parametrization, namely the surface roughness lengths, the stability functions and the wind speed at he lowest model level.

The roughness lengths are in principle a surface characteristic. They control the coupling with the surface in a very direct way and are therefore very important for large scale models. Over land, roughness lengths for momentum are based on land use maps and fine scale orographic data. Although surface roughness values are fairly well known for different vegetation types, area representative values are dominated by heterogeneous terrain effects and orographic features. In principle so called "effective roughness lengths" are defined in such a way that they reproduce the area averaged drag, which implies that the rough and inhomogeneous fractions of the grid box dominate (Mason, 1991). In practice however, information on terrain inhomogeneity is very poor and consequently the roughness length maps in large scale models are not necessarily very accurate. This also applies to the orographic contribution to momentum roughness lengths. Procedures exist to derive roughness lengths from fine scale orographic data (see Mason, 1991 for the so-called silhouette concept), but the results are again highly uncertain with the currently available data sets.

In early models, the roughness lengths for heat and moisture were taken equal to the ones for momentum. A lot of evidence exists now that this is not correct and most state of the art models have different formulations for momentum, heat and moisture. For low vegetation, the roughness lengths for heat and moisture tend to be an order of magnitude smaller than for momentum. However, the ratio is highly dependent on vegetation type and even not necessarily a surface characteristic but dependent on the meteorology (e.g. stability and solar angle; e.g. Malhi, 1996). Furthermore, the transfer coefficients for heat and moisture are less affected by inhomogeneous terrain effects than the transfer coefficients for momentum, and therefore it is necessary to "compensate" the roughness lengths for heat and moisture by reducing them compared to their homogeneous terrain values (e.g. Beljaars and Holtslag, 1991; Hewer and Wood, 1998).

Over the ocean, there is much less uncertainty in transfer coefficients. They are often parametrized using a Charnock relation for the roughness for momentum and smooth surface scaling for the roughness lengths for heat and moisture (see Beljaars, 1997 for the ECMWF formulation and Zeng et al., 1998 for an intercomparison of algorithms). More recently, a fully coupled wave model has been introduced at ECMWF, which makes the Charnock parameter a function of wave age (Janssen and Viterbo, 1996).

The second element in the surface layer formulation is the expression of stability effects. Gradient stability functions have been subject of extensive research in the context of Monin Obukhov similarity theory. Although different formulations are proposed by different authors (see Högström, 1988 for a review), the differences are small compared to other uncertainties. The ECMWF model uses the formulation as presented by Beljaars and Holtslag (1991), which seems to give a reasonable representation of the profiles in different stability regimes. To obtain transfer coefficients, it is necessary to convert the bulk Richardson number in the surface layer into an Obukhov length, which can be done iteratively, or by algebraic approximation (e.g. Abdella and McFarlane, 1997; van den Hurk and Holtslag, 1997).

The third element of the transfer coefficient is the wind speed, which appears straightforward as it is the absolute wind speed at the lowest model level. However, a formulation with the resolved wind speed only has a too weak coupling with the surface at very low winds (even with stability functions that have a proper free convection limit). The reason is that subgrid variability of horizontal wind caused by large eddies maintains a finite surface wind even if the resolved wind components tend to zero. This gustiness needs to be included in the absolute wind speed of the transfer coefficients in order to get a proper free convection limit (Beljaars, 1995a). The gustiness velocity can be parametrized in terms of boundary layer depth and the surface virtual temperature flux. Such a parametrization includes the effect of dry convection and is the minimum gustiness that needs to be considered. However, additional gustiness may be present due to meso-scale variability. The existence and importance of meso-scale variability has been studied for deep convection (Labouille et al., 1996) but is less documented for general situations. Figure 2 shows an example of a wind spectrum from a time series of buoy wind in the Pacific in a stratocumulus area. The spectrum is compared with the corresponding time series from short range forecasts of the ECMWF model. There is a clear indication that the model lacks variability at high frequency, which may require parametrization in the air sea transfer formulation.

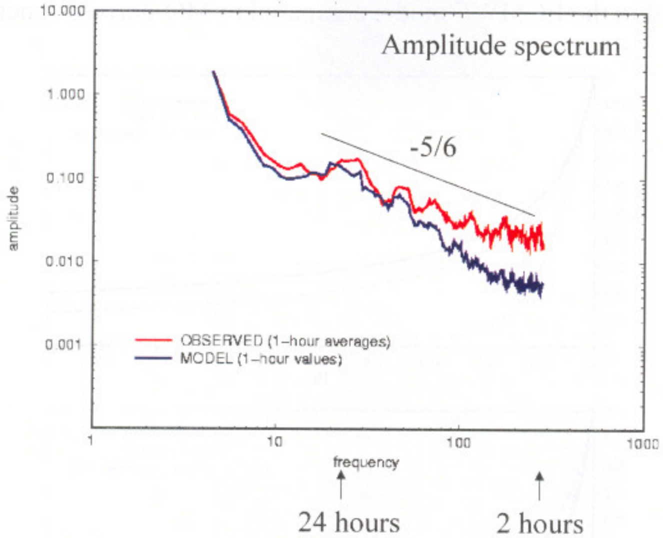


*Figure 2 Amplitude wind spectrum computed from a time series of one hour averages of horizontal wind at the IMET stratus buoy (85°W/20°S). The model equivalent is from daily 12 to 36 hour forecasts written out with a one hour frequency.*

In summary it can be concluded that surface layer parametrization has a firm basis with MO similarity, and that the stability functions are fairly well known. The amount of gustiness that should be parametrized is an unresolved issue. It is very well possible that radiative cooling in the cloud topped boundary layer and effects of shallow convection provide a source for additional gustiness which is not represented in current models. Another unresolved issue is the representation of roughness lengths and its interaction with stability. Heterogeneous terrain and orography are an important source of form drag, but a quantitative representation in large scale models is still highly uncertain. A few examples of sensitivity will be given in the section on momentum fluxes.

## 3. Stable boundary layer

The most popular way of parametrizing the stable boundary layer is with the traditional Louis (1979) scheme in which local fluxes are expressed into gradients with help of diffusion coefficients. The diffusion coefficients for momentum ($K_m$) and heat/moisture ($K_h$) depend on local stability

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

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

where the stability functions $F_m$ and $F_h$ depend on the local Richardson number $Ri = (g/\theta_v)(d\theta_v/dz)|d\bar{V}/dz|^{-2}$ The length scale is an interpolation between the surface layer length scale (height above the surface) and an asymptotic value. The asymptotic length scale is typically between 100 and 150 m, but needs to be reduced to about 30 m well above the boundary layer in order to avoid excessive diffusion in the free atmosphere (e.g. in jet stream areas). The Louis scheme can be seen as an extension of the surface layer formulation with MO similarity where surface fluxes are replaced by local fluxes. This approach found justification through the so-called local scaling concept (Nieuwstadt, 1984), with has as consequence that the stability functions should be compatible with MO functions as observed in the surface layer.

For real application in NWP, the MO stability functions are not very useful because they do not provide enough mixing in the very stable regime. Most formulations for large scale models have much higher values for the stability functions especially for high Richardson numbers. Figure 3 shows two versions of stability functions that have been used in the ECMWF model compared to MO-derived functions.

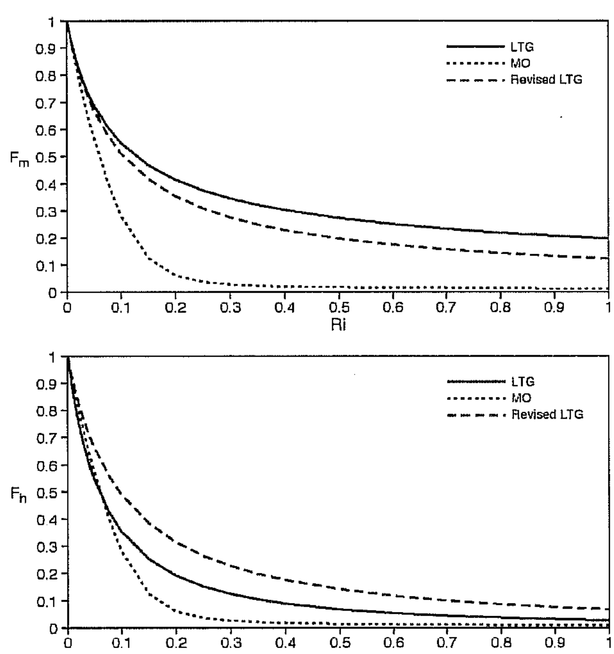


*Figure 3 Three different forms for the stability functions for stable situations as a function of the Richardson number. The top panel shows the functions for momentum, the bottom panel represents the functions for heat and moisture.*

The difference in near surface temperature (2m level) climate for winter between the two versions is reproduced in Figure 4. It is clear that the winter temperatures are very sensitive to the stable boundary layer parametrization, but also the land coupling is important. Viterbo et al. (1999) show that a similar sensitivity exists to e.g. the process of soil moisture freezing which acts as a damping mechanism on the annual temperature cycle due to the thermal inertia of the soil freezing. How much turbulent diffusion should be present on the scale of a grid box can not be concluded from these experiments. It is possible that other aspects of the model have systematic errors, which are compensated for by the turbulent diffusion or by the amount of coupling with the subsurface. There are some indications that the ECMWF model relies to heavily on the damping by the surface, because the depth of the soil that is affected by soil freezing is larger than climatology suggests.

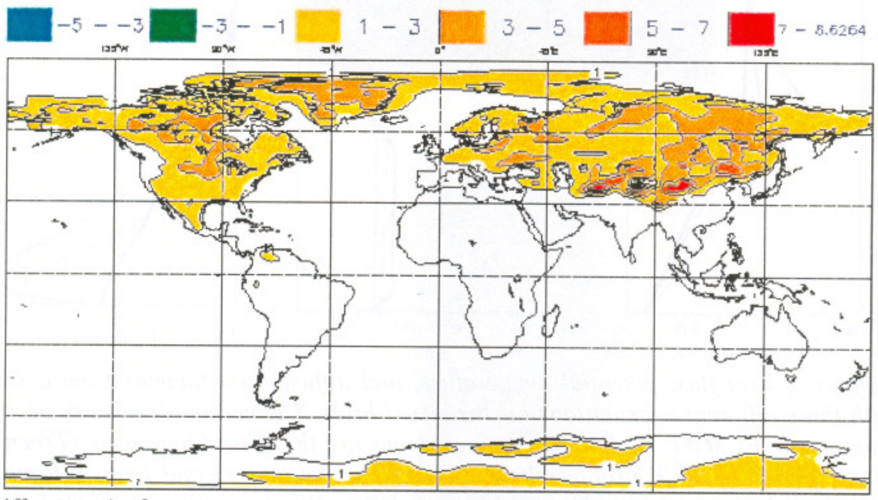


*Figure 4 Difference in 2m temperature model climate for January between a model formulation that uses the revised LTG functions and a model version that uses the original LTG functions for the stable boundary layer.*

On the topic of stable boundary layer diffusion it can be concluded that further research is needed to get more insight in how much diffusion should be present at the grid scale of large scale models. Terrain heterogeneity, subgrid orography and the coupling with the subsoil through the land surface scheme are major issues. Recent work (Zilintikevich and Calanca, 2000) suggests that non-local effects may play a role; stability and gravity wave activity above the stable boundary layer may generate gravity waves that enhance the near surface mixing.

## 4. Inversion layers

Day time boundary layers over land and marine boundary layers are often fairly well mixed in potential temperature and moisture and therefore it is a good idea to think in terms of mixed layer budgets. Not only the surface fluxes are important but also the fluxes through the inversion contribute to the evolution of the mixed layer quantities. Figure 5 shows flux profiles, potential temperature profiles, and profiles of eddy diffusion coefficients for a single column simulation with specified heat flux at the surface and three different parametrization schemes. The first scheme (thick solid) is the traditional Louis scheme with diffusion dependent on local stability, the second one (thin solid) is the K-profile scheme by Troen and Mahrt (1986), in which a profile of diffusion coefficients is prescribed that scales with boundary layer depth and surface fluxes, and the third scheme (dashed) is modified version of the Troen and Mahrt scheme in which among other changes, the boundary layer top entrainment is explicitly prescribed (Beljaars and Viterbo, 1998). The three schemes have rather different profiles of diffusion coefficients, but this is not the cause for the difference in results. The diffusion coefficients are large enough in all three schemes to maintain a mixed layer. The difference in evolution of the mixed layer temperature is entirely controlled by the diffusion in the inversion layer. With the Louis scheme, the diffusion in the inversion is virtually zero because it is controlled by local stability. With the original K-profile scheme, the inversion diffusion depends on the precise position of the diagnosed boundary layer top with respect to the flux levels of the numerical grid. In this particular case the entrainment flux (of potential temperature) is about 60% of the surface flux, but the result is highly variable and depends on vertical resolution. The modified K-profile scheme explicitly prescribes the entrainment flux for virtual potential temperature in the first stable layer as 20% of the surface value. The diffusion coefficient that is needed to achieve this entrainment flux is also applied to the other variables.

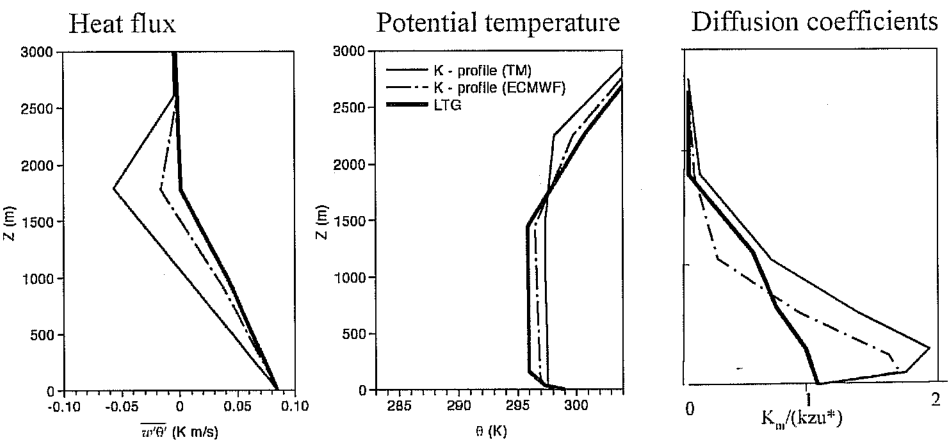


*Figure 5 Profiles of heat flux, potential temperature, and diffusion coefficients from a single column simulation with three different formulations for the mixed layer. The vertical levels are according to the 19 level version of the ECMWF model. The three versions are the K-profile scheme (Troen and Mahrt, TM), the revised version of the K-profile scheme (ECMWF), and the original Louis scheme (LTG). The profiles are obtained by running the single column model for 9 hours with a constant heat flux of 100 W/m$^2$ at the surface.*

When the modified K-profile scheme was introduced in the ECMWF model, it had a big impact on boundary layer moisture and its diurnal cycle over land. Figure 6 illustrates the impact on moisture with an example of a diurnal cycle composite from the FIFE experiment (Beljaars and Betts, 1993). With the modified K-profile scheme, the boundary layer has lower levels of moisture due to the entrainment of dry air from above the boundary layer. Such effects are also seen from shallow convection schemes over the tropical oceans (Gregory, 1996). It is a reminder that dry entrainment is only one aspect of the inversion interaction problem. Moist processes are equally important.

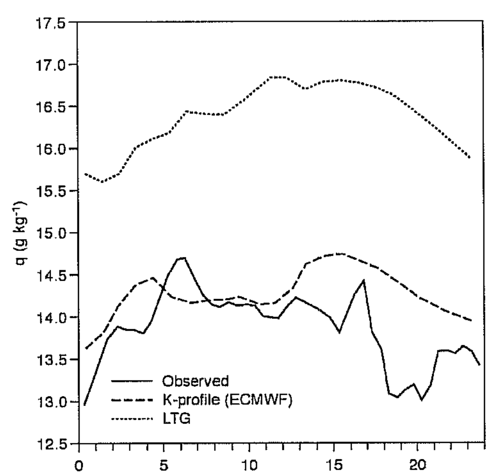


*Figure 6 Composite of 9 diurnal cycles of FIFE observations of specific humidity in August 1987 in comparison with 0-24 hour forecasts from two versions of the ECMWF model (The LTG scheme and the K-profile scheme with a specified entrainment of 20% at the boundary layer top).*

In the context of inversion interaction processes, a number of unresolved problems remain. Although the parameter value of 0.2 for dry entrainment is generally used, also other values are reported (e.g. Betts et al., 1990). It is not clear whether wind shear has a strong influence on entrainment. A further key issue is the entrainment in the cloud topped boundary layer. Inversion erosion by shallow convection is hardly studied, whereas entrainment by stratocumulus is still very uncertain and poorly understood (see the paper by Duynkerke in this proceedings).

## 5. Ocean surface fluxes

Ocean surface fluxes are obviously important for coupled models where biases in the net heat flux lead to ocean temperature drift. Momentum fluxes are also important for coupled atmosphere/ocean circulation models and for ocean wave modeling. Here we focus on heat fluxes.

Figure 7 illustrates the heat ocean flux problems in the ECMWF model. The difference between the JJA model climate and the DaSilva climatology is shown in Figure 7a. The main components to the difference are the surface solar radiation and the latent heat fluxes which are shown in Figure 7b and Figure 7c. The climatology is not a firm reference although a number of differences are most likely due to model deficiencies. The stratocumulus areas off the west coast of N.America, S.America and Africa are clearly visible as areas where too much solar radiation reaches the surface. The same applies to the extra tropical oceans in summer, which is an indication of too little solar reflection from low level clouds. The trades and the ITCZ show the opposite signal; not enough solar radiation is reaching the surface. The precise reason is not very clear. In a diagnostic study Jakob (1997) concludes that the cloud cover is reasonable, but it is likely that the liquid water content is overestimated and/or that cloud inhomogeneity or cloud optical properties are misrepresented.

The latent heat flux is the other important component in the net heat flux biases. The latent heat flux is overestimated in the trades and underestimated in the ITCZ. Although non-negligible, the biases in latent heat flux tend to be smaller than in the solar radiation. Through the surface layer transfer laws, latent heat flux is controlled by wind speed and by boundary layer moisture. Therefore it is interesting to verify wind speed and near surface moisture separately.

Buoy observations are a useful verification source for this purpose. We use the research quality IMET buoys (Anderson and Weller, 1996) for the Pacific locations 125°W/3°S and 85°W/20°S. The first location is at the edge of the ITCZ and has significant convective precipitation from day 315 to 510. After day 510 the sea surface temperature drops (cold tong) and the surface layer becomes stable. The second location is the permanent stratocumulus area off the coast of Chili/Peru. The Figure 8 and Figure 9 show time series for the two locations of wind speed, specific humidity difference between the surface and the observation level of 3.2 m, and the temperature difference between the surface and the 3.2 m level. Both observations and short range model forecasts are shown. Apart from the difference in variability between observations and model, there is no obvious systematic bias in the model. For the 125°W/3°S location there is a small bias in moisture with the model boundary layer slightly too dry. This is consistent with the suggestion from climatology that the model evaporation is too high.

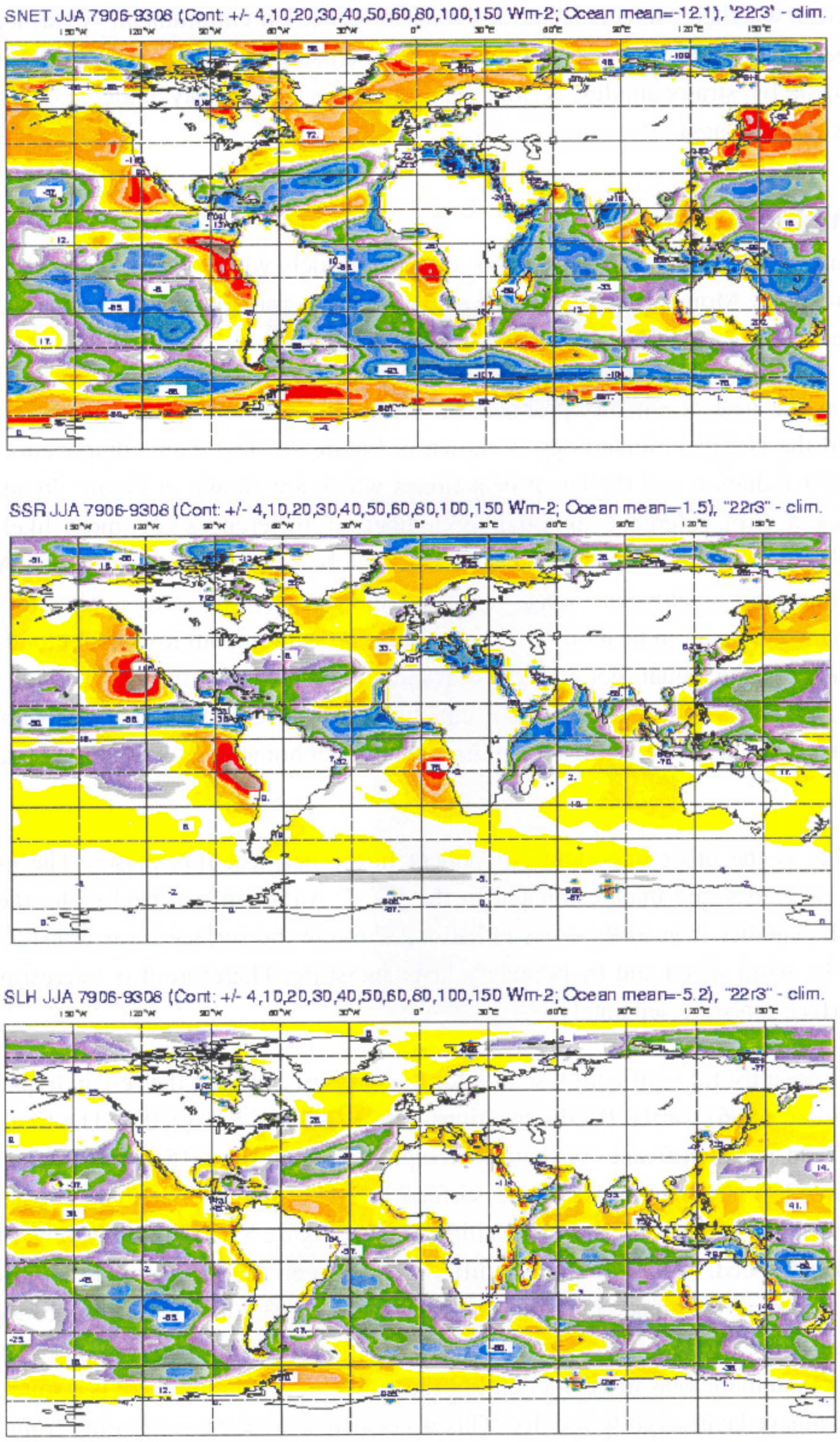


*Figure 7 June/July/August climate of the net heat flux (top panel), the net solar radiation (middle panel), and the latent heat flux (bottom panel) of the ECMWF model (CY22R3) minus the DaSilva climatology of the same quantities. Downward fluxes are positive by model convention.*

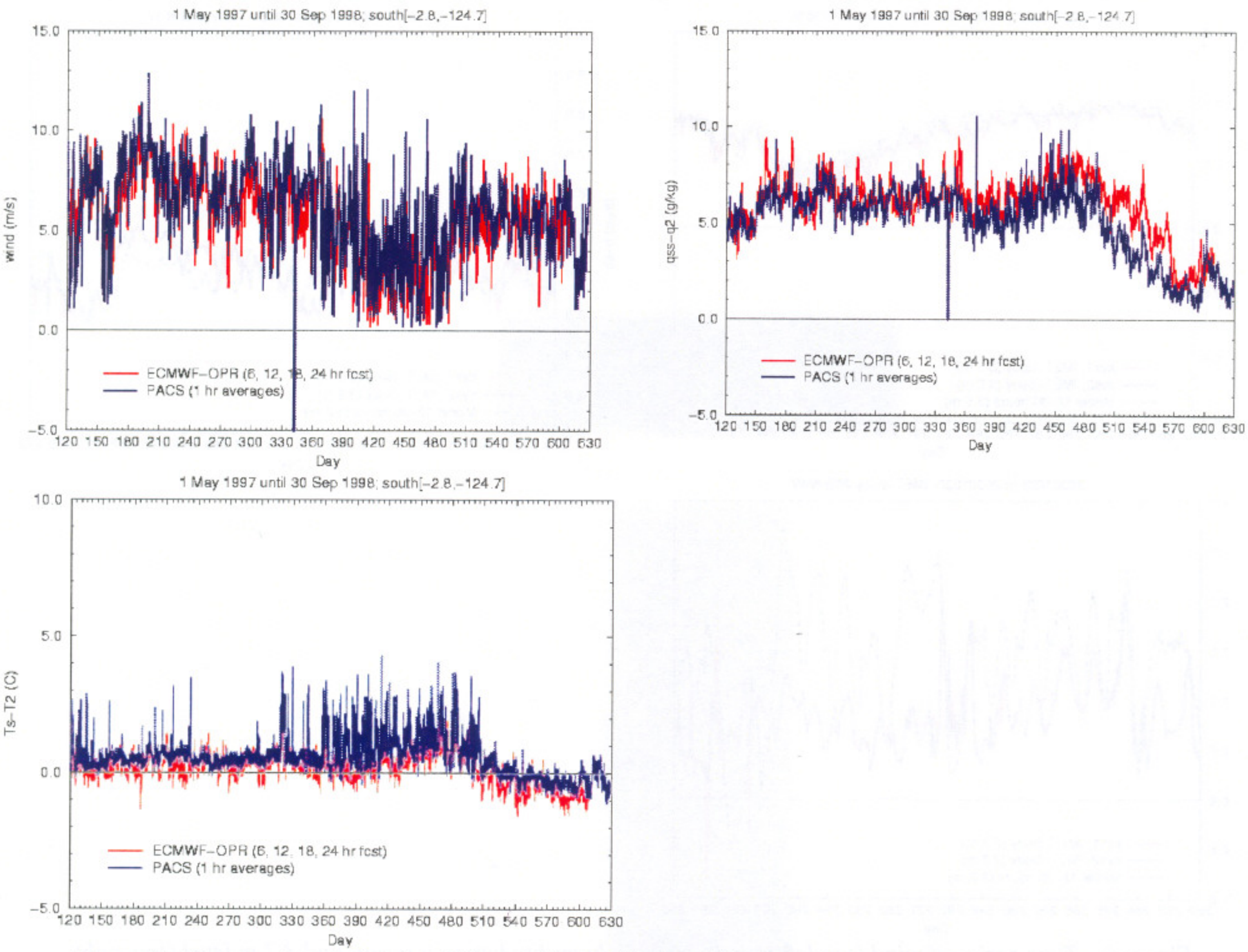


*Figure 8 Time series of wind (top left panel), specific humidity between ocean and 3.2 m level (top right panel), and the temperature difference between sea surface and the 3.2 m level (bottom panel) for the IMET-PACS buoy location 125°W/3°S. The time series covers a period of 17 months, with a data point every 6 hours. The buoy observations (PACS) are one hour averages. The model data is from daily operational, 6, 12, 18 and 24 hour forecasts.*

Such a bias is not visible for the stratocumulus area, where there is perhaps more of a moist bias. Air temperature shows a pronounced cold bias at the stratocumulus location. On the average there is a small warm bias (difference with the SST smaller than observed) at 125°W/3°S location. This bias is particularly large during the observed cold "spikes", which are due to local effects of convection events, which is a sub-grid feature of the real atmosphere and not captured by the parametrization. By and large these fluxes look fairly reasonable, but it should be remembered that these are point verifications which are not necessarily representative for large areas. The biggest error is in temperature in the stratocumulus regime which could be due to errors in cloud top entrainment or radiative effects.

Biases in ocean fluxes are obviously a big issue for coupled atmosphere ocean models. The two main components in the ocean net heat flux are latent heat flux and solar radiation. The biggest bias is in the solar radiation, which is an indication of cloud problems. Also the latent heat flux can show biases, mainly as the result of biases in the boundary layer moisture, which is controlled by ventilation at the top of the boundary layer. Again biases in boundary layer moisture are most likely related to cloud effects in the cumulus as well in the stratocumulus regime.

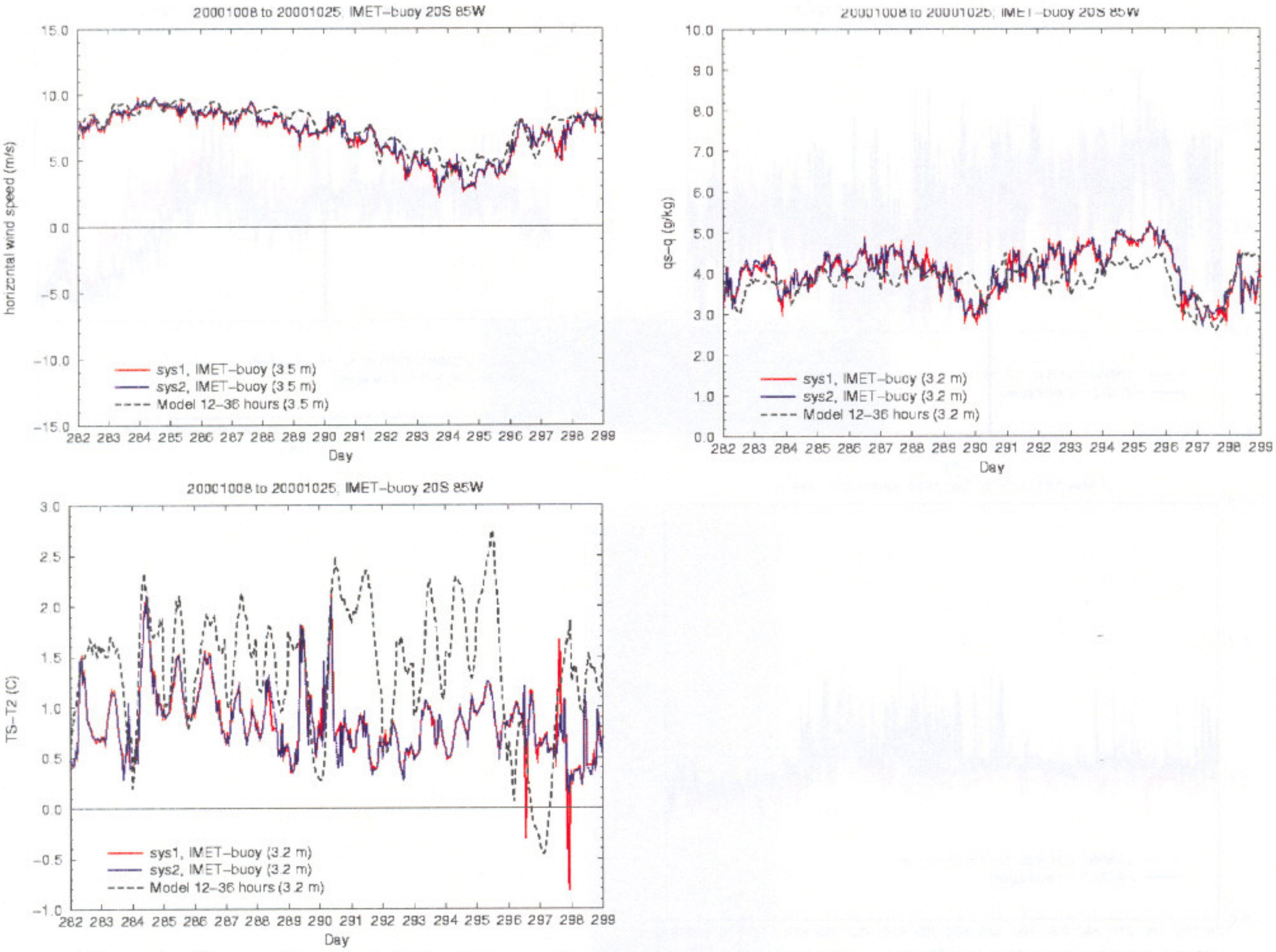


*Figure 9 Time series of wind (top left panel), specific humidity between ocean and 3.2 m level (top right panel), and the temperature difference between sea surface and the 3.2 m level (bottom panel) for the IMET-Stratus buoy location 85°W/20°S. The time series covers a period of 17 days, with a data point every hour. The buoy observations are one hour averages and are given by two independent measuring systems with nearly identical results. The model data is from daily operational, forecasts written out every hour in the forecast range from 12 to 36 hours.*

## 6. Momentum fluxes

Surface momentum fluxes have always had significant impact on the ECMWF model performance. We illustrate the impact by changing the surface roughness lengths over land and by changing the diffusion coefficients in stably stratified flow. In the sensitivity experiments, the roughness lengths over land are divided by a factor 10 leading to less surface drag over land in general, or the stability functions are replaced by the MO-formulation leading to less drag over land in stable situations only. Figure 10 and Figure 11 show zonal means of West-East stress from long integrations for the two sensitivity experiments respectively. Also the zonal stress from the sub-grid orography scheme is displayed. In both cases there is a clear reduction of stress at NH mid latitudes as expected. There is also some compensation from the subgrid orography scheme. The reason is that a reduction of drag in one scheme leads to increased surface winds which results in more drag from the other scheme. It should be remembered that the distribution of stress over land from the two schemes is very different. The turbulence scheme is active everywhere, whereas the subgrid orography scheme is only active in case of stable stratification in areas with significant orography. The effect on the large scale circulation is not very large as can be seen form the Z500 fields and differences in the Figure 12 and Figure 13. Hardly any impact is seen from the reduced roughness length, and some differences are seen in the Arctic from the MO stability functions. This also applies to the near surface temperature climate which are more affected by the stability functions than by the surface roughness lengths (see Figure 14). In a series of winter forecasts (starting from the same initial condition, so no data assimilation), both changes lead to a deterioration of the NH scores, which is consistent with earlier findings (Figure 15). Reduced drag at the surface often leads to degradation of the forecast performance.

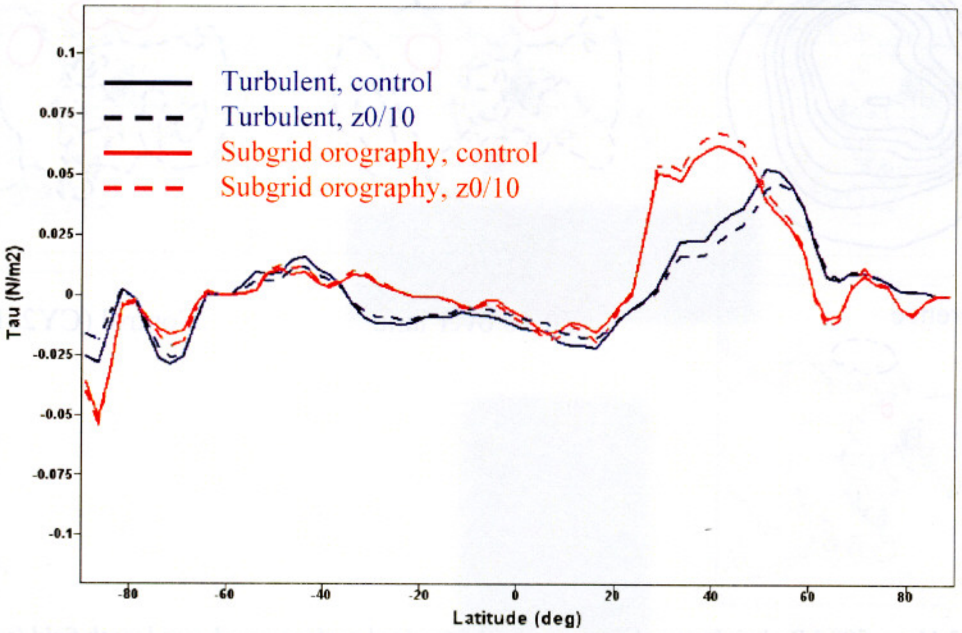


*Figure 10 Zonal mean West-East stress from the turbulence scheme and the subgrid orography scheme with standard surface roughness length field (solid) and with roughness lengths divided by 10 over land. These are results from an ensemble of 6, 120-day integrations averaged over DJF.*

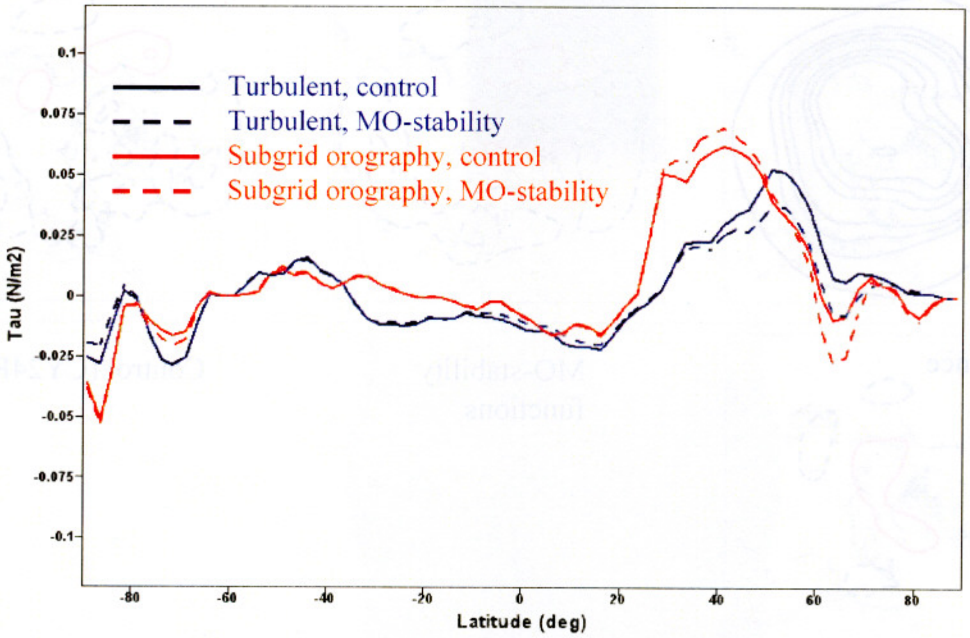


*Figure 11 Zonal mean West-East stress from the turbulence scheme and the subgrid orography scheme with revised LTG in the stable BL (solid) and with MO stability functions for the stable BL. These are result from an ensemble of 6, 120-day integrations averaged over DJF.*

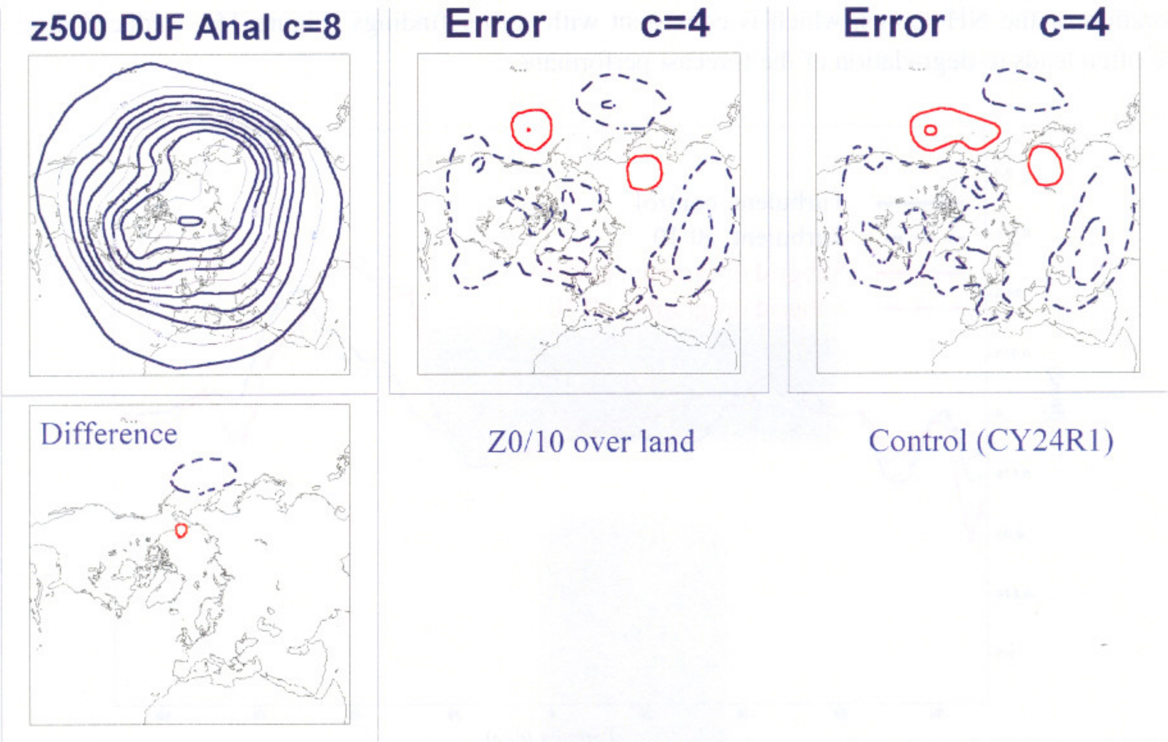


*Figure 12 Mean 500 hPa height (top left), error with standard surface roughness length field (top right) and with roughness lengths divided by 10 over land (top middle), and difference (bottom left). These are result from an ensemble of 6, 120-day integrations averaged over DJF.*

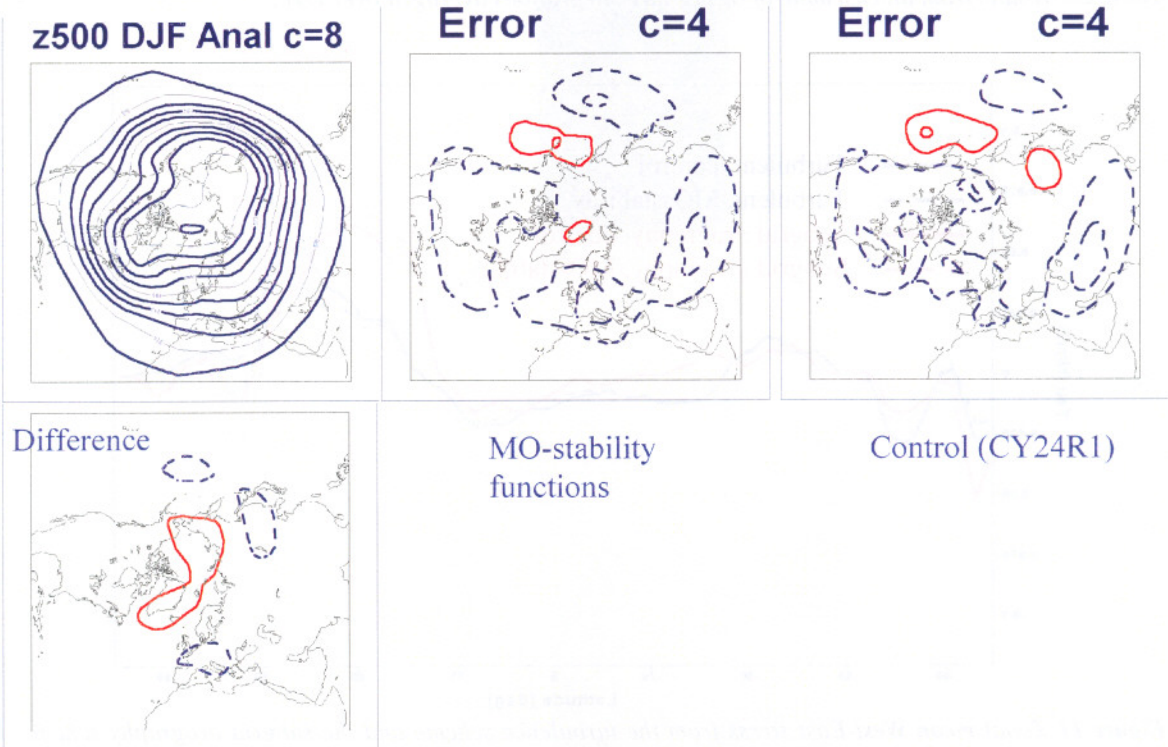


*Figure 13 Mean 500 hPa height (top left), error with revised LTG scheme (top right), error with MO stability functions (top middle), and difference (bottom left). These are result from an ensemble of 6 120-day integrations averaged over DJF.*

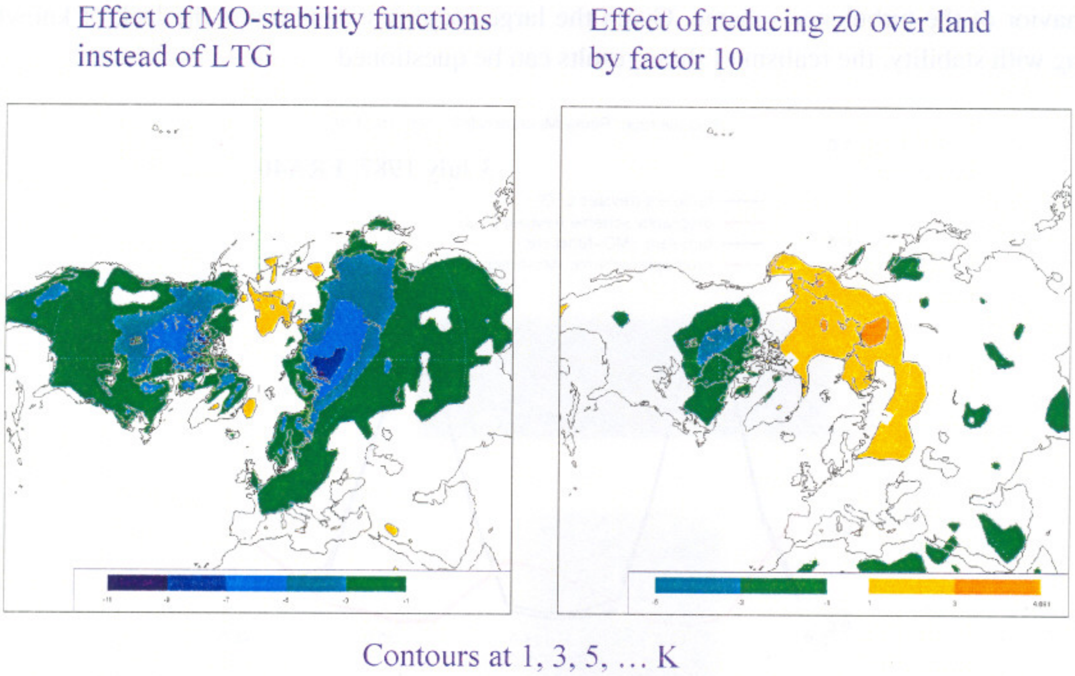


*Figure 14 Difference in 2m temperature climate between model with MO-stability functions and LTG functions for the stable boundary layer (left panel), and difference in 2m temperature climate between model with roughness lengths divided by 10 and the standard model (right panel).*

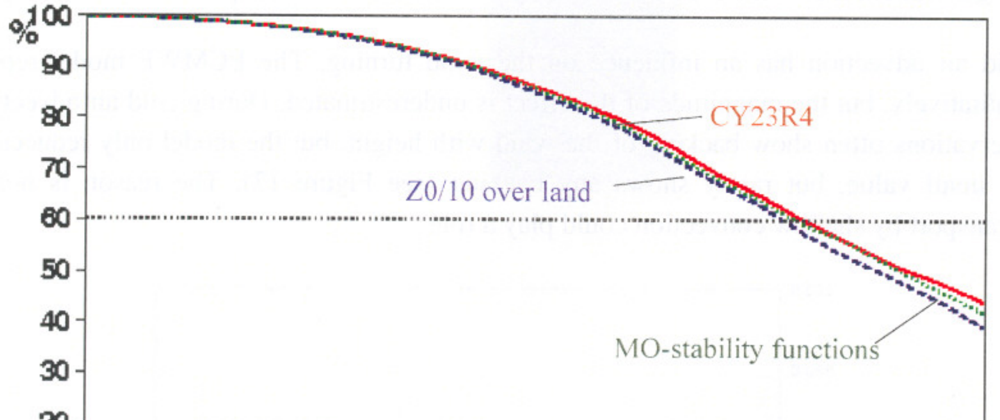


*Figure 15 NH anomaly correlation averaged over 10 winter forecasts with a standard model version (23R4), reduced roughness length over land and MO-stability functions for stable situations instead of the LTG formulation. The different model versions start from the same initial condition.*

To illustrate the behavior of the boundary layer and the subgrid orography schemes in mountainous areas, a plot is made of the diurnal cycle of turbulent and subgrid orography stresses for an area in the Rocky Mountains (see Figure 16). A remarkable feature is the pronounced diurnal cycle in the turbulent drag, which shows only weak sensitivity to the formulation. Again the orography scheme provides partial compensation for the behavior of the turbulence scheme. Given the large roughness length and the lack of knowledge on the coupling with stability, the realism of these results can be questioned.

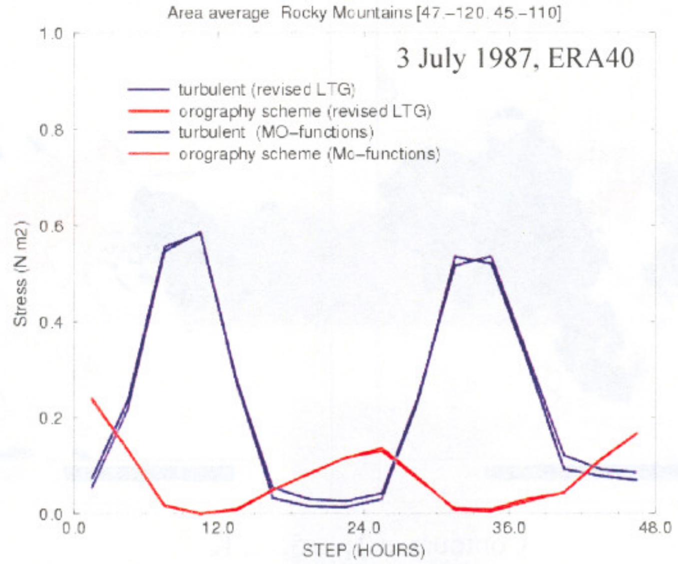


*Figure 16 Two diurnal cycles of West-East turbulent surface stress and the stress from the subgrid orography scheme for an area in the Rocky Mountains. Two model version are used: the standard LTG scheme for the stable boundary layer and the MO-formulation.*

Another feature related to momentum transport is the wind turning with height. From SYNOP verification it is seen that the a-geostrophic angle of the surface wind is systematically underestimated by about 10 degrees. This is a very general feature which is most likely related to momentum transport in the boundary layer.

Warm or cold air advection has an influence on the wind turning. The ECMWF model reproduces the influence qualitatively, but the magnitude of the effect is underestimated. During cold air advection over the Atlantic, observations often show backing of the wind with height, but the model only reduces the normal veering to a small value, but rarely shows any backing (see Figure 17). The reason is not know, but momentum transport by shallow convection could play a role.

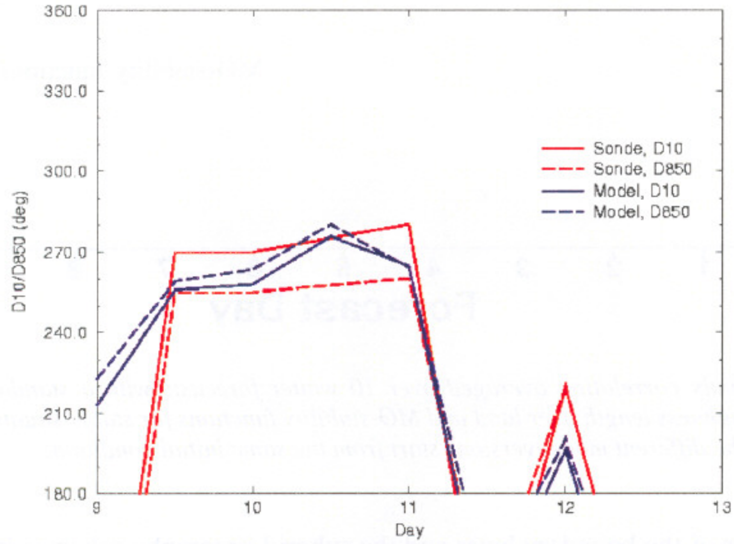


*Figure 17 Wind direction at the surface and at 850 hPa at Ship L (Atlantic) from 9 to 13 Jan 1988. Sonde observations are compared with 12 to 36 hour model forecasts.*

From the sensitivity studies presented here and from earlier experience (Beljaars, 1995b), it is well known that surface drag and subgrid momentum transfer have a strong impact on the large scale model performance as expressed by e.g. NH anomaly correlation. It is also clear that the parametrized fluxes of momentum are highly uncertain, at one hand because of inaccurate boundary conditions (roughness lengths, subgrid orography parameters) and at the other hand due to uncertainties in formulations of the stable boundary layer and subgrid orographic scheme. Quantitative verification on the level of momentum fluxes is virtually impossible. Apart from isolated field experiments (e.g. with PYREX; Lott, 1995), the best hope for verification material is from fine scale model simulations that resolve the subgrid orography. The ECMWF model also has systematic errors in the surface wind direction, which are not well understood and most likely related to momentum transport issues.

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