Experience with the representation of stable conditions in the ECMWF model

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1 Introduction

The formulation of turbulent diffusion in stable conditions has a major impact on the general performance of large-scale models, because stable conditions characterise a wide range of situations, such as nocturnal or hivernal boundary layers (BL), inversions capping the boundary layer, or large parts of the free-troposphere. Most of the formulations used nowadays in large-scale models to represent such conditions are much more diffusive than what one would expect on the basis of observations. This enhanced diffusion is needed to compensate for effects of a crude representation of other processes such as the land-atmosphere coupling, the land surface heterogeneity or the orographic drag. A highly diffusive scheme is beneficial for the performance of models because it helps avoiding a run-away cooling at night (Viterbo et al., 1999) and a too slow decay of cyclones (more diffusive turbulence schemes produce more drag close to the surface, hence decrease the ageostrophic wind angle, therefore reduce the Ekman pumping and thus weaken/shorten the strength/lifetime of cyclones). The counterpart is however that more diffusive schemes also result in too deep boundary layers, too weak low level jets, and too small near-surface ageostrophic wind-angles, hence an underestimation of the wind turning in the boundary layer. Moreover, the inversions capping the boundary layer are often too weak, which leads to an underestimation of the stratocumulus cover and amount both over land and over oceans. As one of the large-scale models using a highly diffusive first-order closure to represent the turbulent diffusion in stable conditions, the ECMWF model (i.e. the Integrated Forecast System - IFS further on) also presents these caveats. Despite the overestimation of the turbulent diffusion in stable conditions, the nocturnal 2m temperatures (T2m further on) forecasted by IFS are often too cold, in particular over Europe during winter time.

Finding a more realistic representation of the turbulent diffusion in stable conditions that would allow correcting these biases has been on the agenda for many years at ECMWF, but a satisfactory solution has not been found so far. In the past year, we examined more thoroughly the model's performance in stable situations and its sensitivity to the representation of turbulent diffusion. We also carefully explored the model's sensitivity to the other element, which is of equal importance for a proper representation of the atmospheric state close to the surface, but whose role is often less emphasized than that of the turbulent diffusion: the land-atmosphere coupling. The aim of this work was to better understand the dependencies among the various processes that control the model behaviour in stable situations and the reasons for its aforementioned caveats. The main insights gained from this investigation are summarized in this extended abstract. We begin by giving two examples of land-atmosphere coupling parameters that have a significant impact on the representation of the atmospheric conditions close to the surface. Then, we illustrate the impact of the representation of turbulent diffusion in the stable conditions on both the boundary layer and the large-scale circulation. Finally, we summarize what we have learned from these numerical experiments.

2 Land-atmosphere coupling

The roughness length for momentum (z_0^m) is one of the land-atmosphere coupling parameters that are important for the representation of near-surface conditions, because it controls the intensity of the 10m wind speed. The persistent tendency of the IFS to overestimate the 10m wind speed compared with synop observations suggested that the values used for the roughness length for momentum for the different vegetation types were too low. Therefore, for each vegetation type we recently used the 10m wind speed forecast errors (relative to the synop observations) typical of both winter and summer conditions, and the logarithmic profile of the wind close to the surface to determine a revised value of z_0^m for which the mean forecast error would be close to zero. We increased thus z_0^m for 10 of the vegetation types (see caption Fig. 1). This change led to a significant reduction of the mean 10m wind forecast errors for the respective vegetation types, and in particular for the predominant ones, i.e. crops and interrupted forest (Fig. 1). As the roughness length for momentum was increased, the roughness length for heat has been decreased in order to account for land-surface heterogeneity. The reduction of the roughness length for heat results in a warming of the air close to the surface during nightime and in a cooling during daytime, and hence in a reduced diurnal cycle of T2m. It therefore corrects for the model tendency to present a cold bias during nighttime, a warm bias during daytime and a too strong diurnal cycle for the T2m over continental areas, and in particular over Europe. The modification of the roughness length for heat also improves the representation of the skin temperature (see the extended abstract of Isabel Trigo). The revised values of roughness were implemented in the cycle 37R3 which became operational in November 2011.

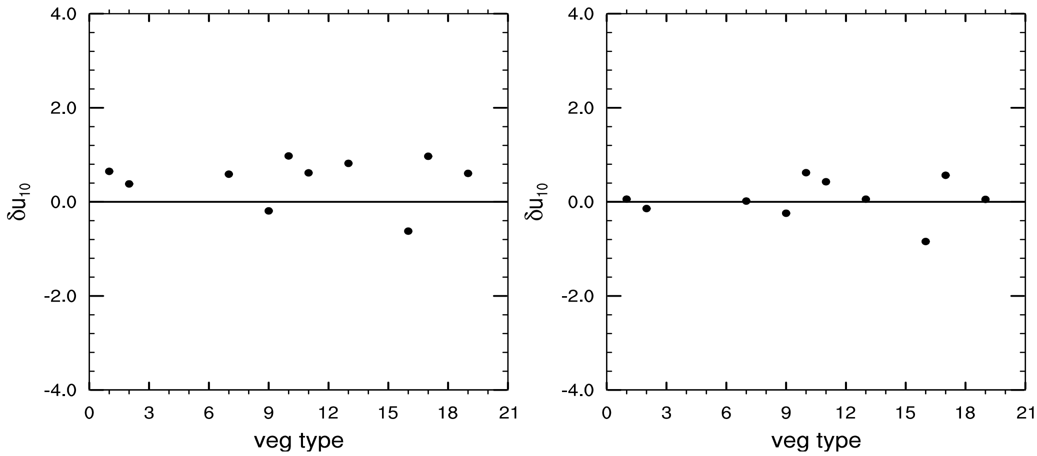

Figure 1: Mean bias of daytime 10m winds (m/s) for snow free locations during August 2010, as a function of the vegetation type of the respective locations, for the old (left) and the new (right) values of the roughness length for momentum. The errors are showed only for the vegetation types for which we changed the roughness length for momentum: 1 - crops, mixed farming, 2 - short grass, 7 - tall grass, 9 - tundra, 10 - irrigated crops, 11 - semi-desert, 13 - bogs and marshes, 16/17 - evergreen/deciduous shrubs, 19 - interrupted forest. The diagnosed winds are taken from steps 24 and 36 of T511L91 forecasts initialized from their own analysis performed with IFS cycle 36R4.

Another parameter playing a role in the representation of the near-surface and soil parameters is the skin layer conductivity (λ_{sk}), which in the IFS simulates the degree of coupling between the surface and the atmosphere. Comparison of 1 year long forecasts relaxed towards ERA-INTERIM above the boundary layer with soil observations from a couple of hundred stations in Germany, show that doubling the current value of these coupling coefficients would significantly diminish (almost halve) the daytime errors for the soil temperature during spring and summer (Fig. 2), while it would not affect its nighttime values. An increase in the coupling would also improve the representation of the nighttime T2m and of its diurnal cycle (not shown). This suggests that the choice of these coupling coefficients, whose current values don't have a solid basis, is very important and should be revised.

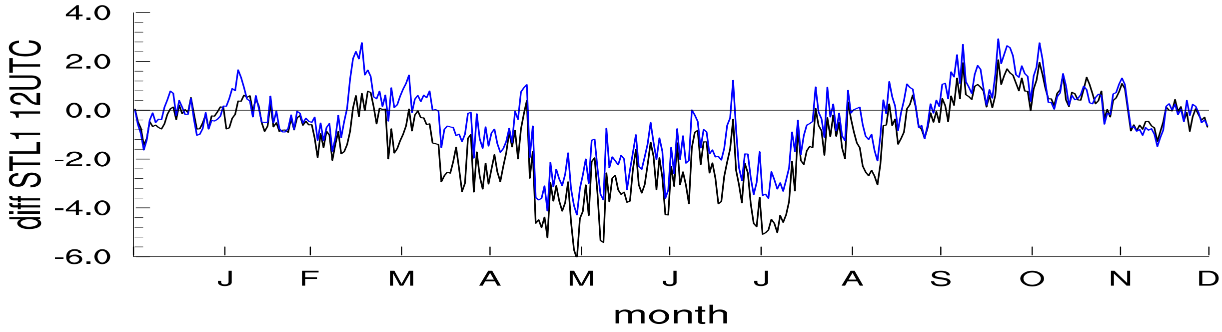

Figure 2: Error on the soil temperature of the first modelled soil layer (i.e. 0-7 cm) with respect to observations at 12 UTC for the control (black) and doubled λ_{sk} relaxation runs (blue). Only snow free grid points were included in this comparison.

3 Representation of turbulent diffusion

3.1 Current representation of turbulent diffusion in the IFS

In IFS, the turbulent diffusion taking place in stable situations is treated with a first order closure based on local stability. The exchange coefficients for momentum and heat K_M, K_H depend on a mixing length l, the gradient of horizontal wind and some stability functions $f_{M,H}$, which are a function of z/L (L being the Monin-Obukhov length) in the surface layer and of the local Richardson number R_i above the surface layer (i.e. in stable boundary layers, inversions capping the boundary layer, stable regions of the free troposphere):

$$K_{M,H} = l^2 |\frac{\partial U}{\partial z}| f_{M,H} \tag{1}$$

The mixing length $l = kz$ used in the surface layer is bounded above this layer by introducing an asymptotic length scale $\lambda = 150m$: $1/l = 1/kz + 1/\lambda$. As for the stability functions, a short tail version of the Monin-Obukhov functions are used in the surface layer (SFMO - blue lines in Fig. 3). Above the surface layer, the exchange coefficients were described until a few years ago by the Louis et al. (1982) functions (long tails) and then by a revised - more diffusive - version of these functions (Fig. 3 - LTG, black line, Viterbo et al. (1999)). Since IFS cycle 32R3, f_H, f_M are given by an interpolation which allows the revised LTG stability functions to be used near the surface and a less difussive (short tail) form of the Monin-Obukhov functions (Fig. 3 - MO, red line) to be used away from the surface (i.e. typically in the inversions capping the boundary layer and in the stable parts of the free-troposphere):

$$f(R_i) = \alpha f_{LTG}(R_i) + (1-\alpha) f_{MO}(R_i) \tag{2}$$

where $\alpha = exp(-z/\lambda_{LTG})$, with $\lambda_{LTG} = 150\,m$. This interpolation was introduced in order to prevent the erosion of stratocumulus due to mixing through inversions, which was far too pronounced when the revised LTG functions were used to describe the inversions capping the boundary layer (Kohler et al., 2011).

This representation of the turbulent diffusion is one way of dealing with the fact that in the stable boundary layer cases the model seems to need more diffusion and in the stable inversions topping the

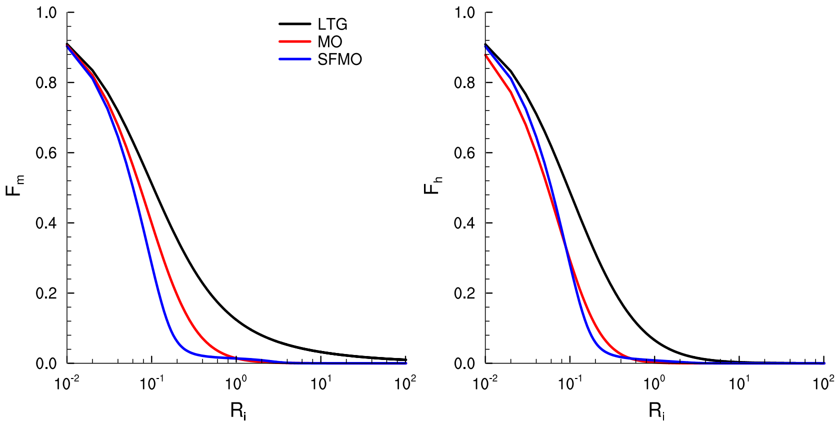

Figure 3: Stability functions for momentum and heat: Monin-Obukhov functions used in the surface layer (blue), revised Louis functions used close to the surface (black) and Monin-Obukhov functions used further away from the surface (red).

PBL it needs less diffusion in order to avoid their erosion and the destruction of PBL clouds, especially of the stratocumulus decks. Another way to cope with these issues is the method chosen in the UK MetOffice models, which consists in using long tail stability functions close to the surface over land and short tail functions over oceans (Brown et al., 2008).

3.2 Experiments

As mentioned in the introduction, overestimating the diffusion over land leads to a number of biases, namely in terms of wind turning in the boundary layer, of low level jets, of diurnal cycles of wind and temperature close to the surface, etc. This is a strong motivation for trying to reduce in future the degree of mixing produced by the model in stable conditions. In other words, we would like to find a less diffusive formulation of the turbulent mixing in stable conditions, that, combined with adjusted values of the parameters describing the land-atmosphere coupling, would allow to improve the representation of the near-surface parameters. A first, important, step towards this revision of the current scheme consisted in investigating how the representation of turbulent diffusion in stable conditions affects different aspects of the system (BL structure, i.e. profiles of temperature, wind, moisture, vertical motions; BL height, stratocumulus cover, large-scale circulation). For this purpose, we performed an extended set of T511L91 ($\approx$ 50 km resolution) 10-day forecast experiments (for both a winter and a summer month) with various ways of representing the exchange coefficients in stable cases. In this abstract, we discuss only three of these sensitivity tests, which illustrate well the impact of changing either the stability functions or the asymptotic mixing length on the model behaviour. In the first test, the interpolation used currently above the surface layer between the long and the short tails is replaced with the short tail functions used in the surface layer (blue lines in Fig. 3, so the blue lines are now used allover), while the asymptotic mixing length is maintained unchanged, i.e. 150m (ST150 experiment further on). In the next two sensitivity tests, the long tails used at present close to the surface (black lines in Fig. 3) are used everywhere above the surface, and the asymptotic mixing length is reduced from 150 to 50 and, respectively, 30m (LT50 and LT30 further on). In the next two sections we illustrate what is the impact of these changes on the boundary layer, and respectively on the large-scale circulation.

3.3 Impact of the turbulent diffusion on the boundary layer

A reduction of the turbulent diffusion in stable conditions impacts the boundary layer structure in many ways. First of all, less turbulent mixing leads to shallower stable boundary layers, which means that the radiative cooling is felt in a more confined layer, hence the near-surface temperature drops more than in deeper boundary layers. This effect is clear for the ST150 experiment where the long tails used now close to the surface are replaced with the much less diffusive functions used in the surface layer. Indeed, for a winter month, when stable conditions are often encountered in the North Hemisphere, the mean T2m is in general between 0.2 and 1 C lower in the ST150 experiment than in the control (CTRL hereafter) run and can even become 3 to 5 C lower in very stable conditions encountered in snow covered regions (top left panel of Fig. 4). Such an effect is obviously not beneficial for the performance of the model, which already tends to predict too cold near-surface temperatures in stable boundary layers (nighttime or winter conditions). For the experiments using long tails and a smaller asymptotic mixing length (λ), the scheme becomes less diffusive than the one used in the operational model only when λ drops to, or below, approximately 30m (instead of 150m), as suggested by the mean change in near-surface temperature in the experiments LT50 and LT30 compared to the CTRL run shown in Fig. 4. What is interesting to notice however is that the strong drop of the near-surface temperature caused by reduction of the diffusion in the ST150 experiment is more than compensated, in all regions except the snow covered areas, by a doubling of the skin layer conductivity (Fig. 5 versus top left panel of Fig. 4). This suggests that these coefficients are as important as the turbulent diffusion scheme for the representation of the near-surface temperature. It also suggests that increasing the coupling, which seems to be supported by the comparison with observations discussed in Section 2, could represent a way to compensate for the deterioration of the quality of the T2m forecasts caused by a reduction of the turbulent diffusion in stable conditions.

Another impact of reducing the turbulent diffusion in stable conditions is the expected, and desired, increase in wind turning in the BL (top and bottom plots, central column of Fig. 4) and the better representation of the low level jet (not shown).

However, another aspect that should not be neglected when changing the representation of the turbulent diffusion in stable conditions is the impact on the inversion layers capping the boundary layer, and hence on the representation of stratocumulus clouds. The IFS, like many other large-scale models, tends to underestimate the stratocumulus cover. Hence, enhancing even slightly the diffusion might have a significantly negative impact on the marine stratocumulus decks, as it is the case for the LT50 experiment (right column, Fig. 4). Maintaining a level of diffusion comparable to the current one for the inversion layers translates unsurprisingly in a small impact on the stratocumulus decks. This is the case for the ST150 and LT30 runs, where the exchange coefficients in the inversions are similar to those obtained with the current version, and therefore only a slight erosion of the stratocumulus decks is seen during the summer months (while during winter the impact is negligible).

If we were to evaluate the model performance only in terms of boundary layer representation, it would seem that using the long tail stability functions plus an asymptotic mixing length of 30m would be the best compromise (from the different options for representing the turbulent diffusion in stable conditions discussed here). Indeed this option improves the representation of the wind in the boundary layer, while it does not cool the air close to the surface too much and it does not reduce significantly the stratocumulus cover. However, in a large-scale model, the impact on the large-scale circulation is also very important and cannot be overlooked. In the next section, we therefore discuss how the changes to the diffusion formulation in stable conditions affect the large-scale circulation and thus the general performance of the model.

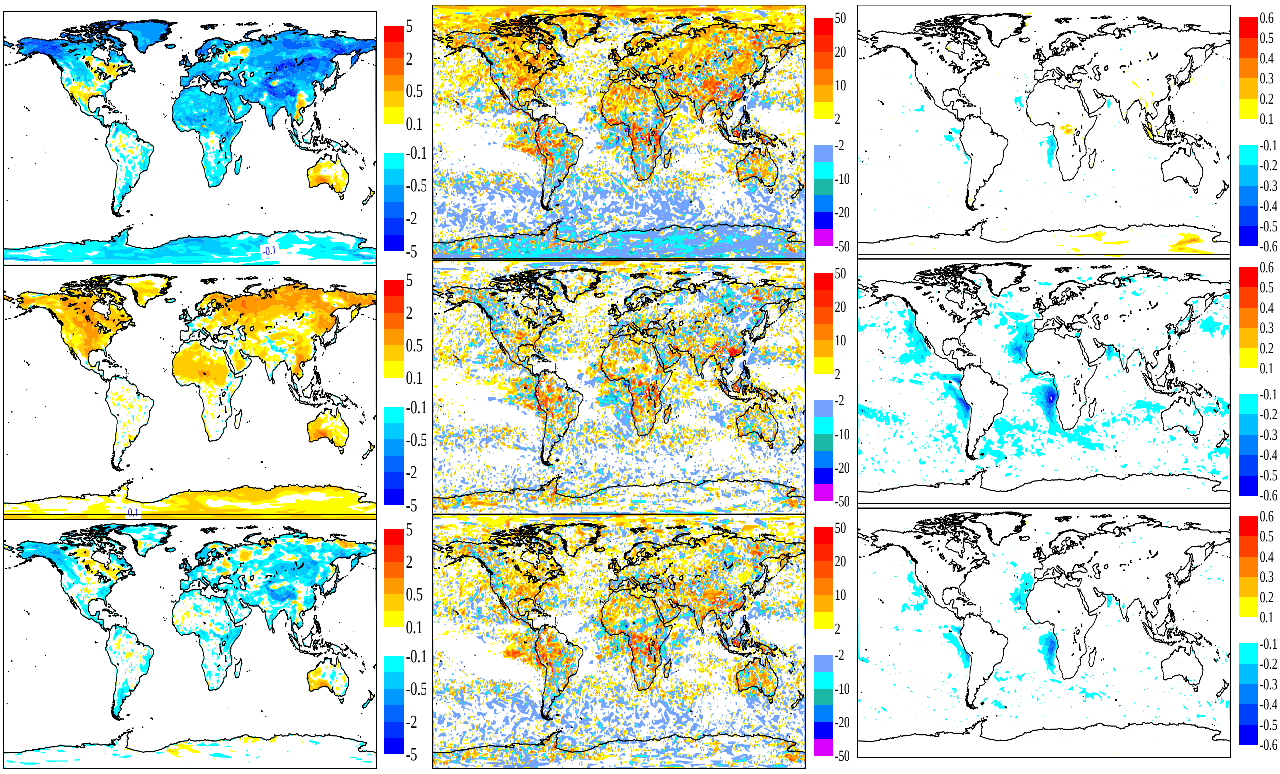

Figure 4: Monthy mean differences in 2m temperature (K, left column), wind turning in the boundary layer (defined as the difference in wind angle between 700hPa and the surface, degrees, centre) and low cloud cover (right column) at step 72 (00UTC) between the ST150 (top row), LT50 (middle row), LT30 (bottom) and the CTRL experiments performed for January 2011 (left and centre columns) and July 2010 (right column).

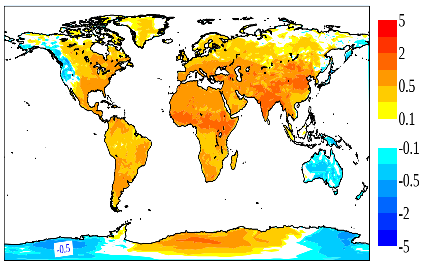

Figure 5: Monthy mean differences in 2m temperature (K) at step 72 (00UTC) between the experiment with doubled skin layer conductivity and the CTRL experiment performed for January 2011.

3.4 Impact of turbulent diffusion on the large-scale circulation

The impact of the various changes to the turbulent diffusion in stable conditions on the model performance at the large-scale is not negligible and both has positive and negative impacts. Reducing the diffusion (ST150 and LT30 experiments) always acts to improve the performance of the model in terms of the headline scores (for e.g. geopotential height at 1000 and 500 hPa) in the summer hemisphere and to degrade it in the winter one. Both the improvement/degradation are significant and can reach a relative value of 5 to 10 %. The effects on the large-scale circulation seem to be related to the impact of changing the turbulent diffusion on the strength of the low and high pressure systems through the modification of the Ekman pumping. Indeed, reduced turbulent diffusion implies less drag, so less ageostrophic wind in the BL, hence reduced Ekman pumping and therefore longer living, or stronger cyclones, or the other way around, slower weakening of the anticyclones, hence stronger anticyclones.

Somewhat surprisingly, the increase in model activity (measured in our case as the mean squared departure of 1000hPa geopotential heights from a climatology) is obvious not only at synoptic scales, but directly at planetary scales, as illustrated in Fig. 6. The top plots in Fig. 6 show the model activity in the control (red) and LT30 experiments (blue) performed for January 2011, for the North (left) and South Hemisphere (right). In the winter hemisphere (i.e. Northern in this case), the activity in the control forecast (red dotted line) is very close to the one in the analysis (blue dashed line), so increasing it in the LT30 experiment deteriorates the performance of the model (blue dotted line), at both scales and increasingly in time. For the summer hemisphere (the Southern in this case), the forecast is characterized by a too weak activity compared to the analysis, so an increase in activity associated with the reduction in turbulent diffusion significantly improves the quality of the forecast.

In order to better understand why the activity increases when the turbulent diffusion is reduced we performed two sensitivity experiments (for the same month, i.e. January 2011). In the first one, we replaced the current formulation of the turbulent diffusion with the LT30 option only for the open ocean grid points. It appears that in the North Hemisphere the activity changes only little in this case compared to the CTRL run (bottom left plot in Fig. 6), suggesting that the degradation of the model performance in the LT30 run (top left plot in Fig. 6) is related to the reduction of the turbulent diffusion in stable conditions over land. Interestingly, this is not the case for the South Hemisphere, where the increase in activity is not associated with changing the diffusion over Antartica as one might expect, but to the changes in diffusion in the storm tracks region (not shown).

In a second run, we corroborated the LT30 change to the turbulent diffusion with an increase by approximately 50 % of the turbulent orographic drag strength. The result is striking: for the North Hemisphere the activity of the forecast is much closer to the one in the CTRL run (or the analysis) than in the LT30 experiment (bottom right versus top left plot in Fig. 6). An analysis of the forecast errors obtained for the North Hemisphere for this winter month showed that in the CTRL run both the depressions (or the ”lows”) are too deep and the high pressure systems (or the ”highs”) are too weak (here we reffer to ”lows” or ”highs” in a climatological sense, i.e. the ”lows” or ”highs” seen in the monthly mean 1000hPa geopotential height, which correspond to the minimum and maximum of the planetary waves train). When reducing the diffusion in the LT30 exp, the ”lows” become even deeper (stronger cyclones) and the ”highs” become stronger (stronger anticyclones). This translates in a reduction of the error compared to the CTRL run in the ”highs” and an increase of the error in the ”lows” (not shown). When on top of the change to the diffusion we increase the drag over the mountains, we get back some of what was lost (in terms of model performance) due to the reduction in diffusion. The ”lows” become less deep, and the ”highs”, somewhat surprinsingly, strengthen even more than in the LT30 run. The errors in the ”lows” are comparable to the ones in the CTRL run (hence much smaller than in the LT30 run) and their reduction in the ”highs” is much stronger than in the LT30 run.

This suggests that the change in drag over orography modulates the amplitude of the planetary waves in the North Hemisphere. And that the deterioration of the model‘s performance in North Hemisphere during a winter period, caused by a reduction in turbulent diffusion, can be at least partially corrected with an increase in the turbulent orographic drag. The counterpart is that in summer, an increase in the turbulent orographic drag would anihilate the improvement of the model performance associated with a reduction of the diffusion in stable conditions. Also, the modification of the orographic drag does not have much effect in the South Hemisphere.

Finally, another aspect, that we found to be sensitive to the formulation of turbulent diffusion in stable conditions, is the representation of the upper troposphere jets. The forecast errors of the wind speed

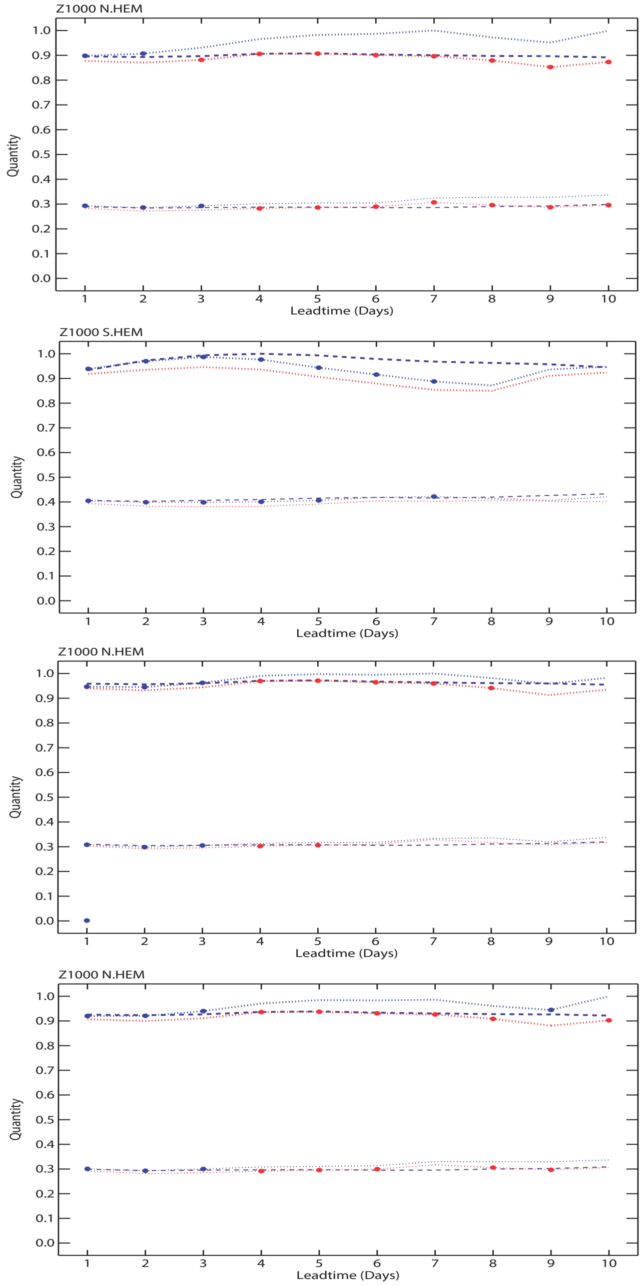

Figure 6: Model activity (computed as the mean square departure of the 1000hPa geopotential height from a climatology, normalized by the maximum value of this parameter) as a function of leadtime, averaged over the North or the South Hemisphere, derived for the forecasts performed in the various experiments (dotted lines) and for the analysis from which all the forecasts are initialized (blue dashed lines). The upper and respectively the lower groups of lines show the activity at planetary and synoptic scales. The red dotted lines correspond in all panels to the CTRL forecasts, and the blue dotted lines to the various experiments as follows: (top left and right) LT30: for the N. and respectively S. Hemisphere, (bottom left) LT30 only over open ocean, for the N. Hemisphere; (bottom right) LT30 (allover) and increased orographic drag, for the N. Hemisphere. The collor of the dotts on the different lines showing the forecast activity indicate in which case the results are significantly better (for e.g. if they are blue it means that the experiment is significantly better than the control, and if they are red it is the other way around.)

at 200 hPa, with respect to the analysis, show that the subtropical jet over Asia is too weak, the ones over the Pacific are too strong, and the winds are too weak in the low wind speed region west of the Gulf of Mexico (not shown). If we either decrease or increase the turbulent diffusion in stable regimes, we will necessary improve the model performance in some regions and deteriorate it in others. For e.g. decreasing the diffusion will improve the representation of the wind in the regions where it is currently underestimated (because less diffusion means an increase in the wind speed), but it will deteriorate it in the regions where it is already too strong. Such effects are reflected in the tropical wind scores at 200 hPa, which are deteriorated irrespective of the change we make in the representation of the turbulent diffusion. This behaviour adds another level of difficulty when chosing a formulation for the turbulent diffusion in stable conditions, and highlights the importance of specifying the level of mixing for the free shear layers.

4 Conclusions

Our investigation of the model sensitivity to the formulation of both the turbulent diffusion in stable conditions and the land-atmosphere coupling showed that the errors in the representation of the near-surface parameters are not entirely related to the formulation of the diffusion in stable conditions. The coupling with the surface also plays a major role for correctly reproducing near-surface parameters such as 10m winds, 2m temperature and humidity.

Reducing the unrealistical turbulent diffusion which is often prescribed for stable conditions in large-scale models would have both positive and negative impacts on the performance of the IFS. It would for example correct for the current underestimation of the low level jets strength, of the wind turning in the boundary layer, the dry bias close to the surface or the underestimation of the stratocumulus cover. But at the sametime, if not corroborated with changes to the land-atmosphere coupling it would further increase the cold bias near the surface in stable boundary layers. In terms of the headline geopotential scores, the impact would be positive for both the North and the South Hemisphere during summer, but it would be negative for winter, especially in the North Hemisphere (in the South Hemisphere a negative impact is seen only near surface, i.e. 1000hPa). The strong impact on the geopotential scores is explained by the fact that a reduction of the turbulent diffusion in stable conditions affects not only the strength of cyclones and anticyclones, at the synoptic scale, but also the amplitude of the planetary waves and thus the large-scale circulation. The tropical upper troposphere wind scores are also very sensitive to changes in the turbulent diffusion in stable conditions, which points towards the importance, but also the difficulty in specifying an appropriate value for the asymptotic mixing length in the free shear layers (e.g. in the upper troposphere subtropical jets).

We also found that the formulation and strength of the turbulent orographic drag is crucial for the level of activity of the model, because changes to the orographic drag affect directly the amplitude of the planetary waves.

This work leads to a few open questions which beg for an answer if progress is to be made in the representation of the boundary layer in large-scale models: (i) can we assess the land-atmosphere coupling parameters from observations?; (ii) is there a sensible way of calibrating the orographic drag (both the gravity wave and the turbulent orographic drag)? (iii) how to prescribe the level of mixing in free shear layers?; (iv) is a simple formulation for the mixing length, such as the one used now in the IFS, sufficient or should it be stability dependent?

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