The stable boundary layer in the ECMWF model

Anton Beljaars

With contributions by: Gianpaolo Balsamo, Alan Betts, Emanuel Dutra, Anna Ghelli, Martin Köhler, Irina Sandu, Soumia Serrar and Pedro Viterbo

ECMWF
Shinfield Park, Reading, Berks, UK
a. beljaars@ecmwf.int

1. Introduction

Many models, including the ECMWF model, show strong sensitivity to the parametrization of the stable boundary layer. The sensitivity is reflected in the night time temperature, the amplitude of the diurnal temperature cycle, the structure of the wind profile (wind direction, surface wind, low level jet), the surface stress, and the impact on the large scale flow. Similarity theory supported by observations is well established for the stable boundary. It is widely accepted for the surface layer with Monin Obukhov similarity, and can be extended to the outer layer through local scaling (Nieuwstadt 1984). This is fully consistent with a closure scheme that expresses the turbulent diffusion coefficients into shear, the distance above the surface as length scale and stability functions dependent on the Ri-number (Louis 1979). However, very few models use functions that are purely observationally based. With an observationally based formulation it is very difficult to control the diurnal cycle of temperature and to obtain optimal large scale scores (Beljaars 2001). The reason for this discrepancy is not well understood. Possible explanations are: (i) similarity theory only applies to the fully turbulent stable boundary layer and not to the intermittent low wind regime, (ii) similarity theory applies to homogeneous terrain only, whereas real terrain is nearly always sloping or covered with inhomogeneous vegetation, and (iii) meso-scale variability may contribute to the vertical transport or enhance turbulence.

This is the reason that many models use so-called “long tail” stability functions to have some level of diffusion at high Richardson numbers in the regime where traditional similarity theory has virtually no turbulent transport. One of the complications is a positive feedback between turbulent diffusion and stability. In case of surface cooling through radiation, the increased temperature gradient leads to an increase of the heat flux towards the surface, but stronger stability opposes such an increase. The result can be a so-called “run-away” cooling of the surface. In that case turbulent diffusion stops altogether and a radiative equilibrium between the atmosphere and the surface is established. The “long tail” stability functions are efficient in controlling such run-away cooling.

This short paper gives an overview of experience at ECMWF with stable boundary layer parametrization issues. The main features of the ECMWF scheme are described in Beljaars and Viterbo (1998) with recent upgrades in Köhler et al. (2011). As illustration, Fig. 1 shows the mean and random errors of the January 2011 night time temperature. The large scale patterns in the mean

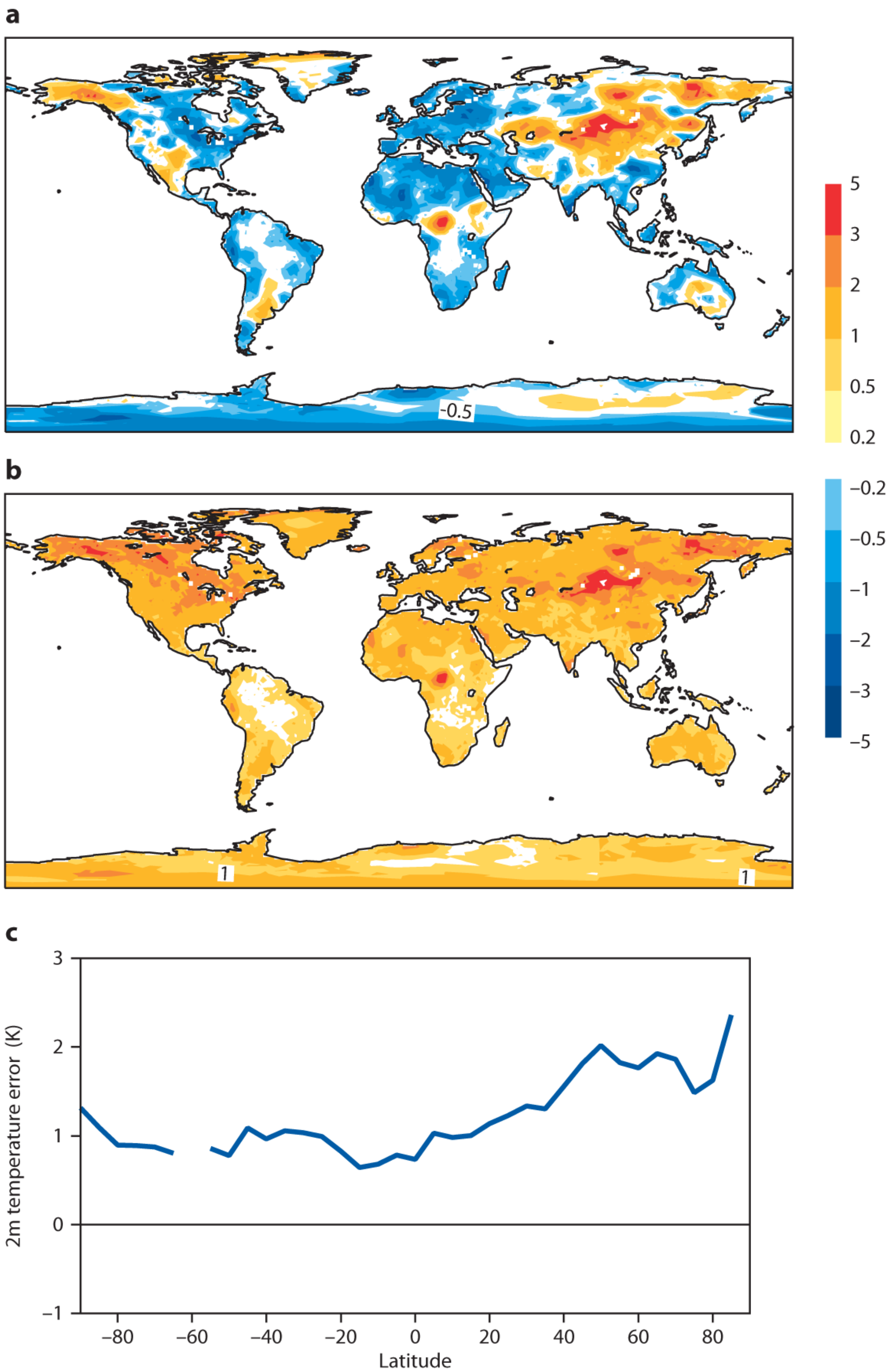

Figure 1: Verification of the night time 2m temperature of the operational ECMWF system against the analyses for January 2011. The 2m temperature analysis draws well to the SYNOP observations except in areas with very few or no SYNOP stations. The night time temperature has been obtained by selecting a verification time of 0,6,12 or 18 UTC (dependent on longitude and latitude) to be closest to the minimum temperature for each location. Daily 24, 30, 36 and 42 hour forecasts have been used. The top panel displays the mean error, the middle panel the mean absolute error, and the bottom panel the zonal mean absolute error over land.

errors are not well understood, but the coherence in the patterns suggests that improvements should be possible. Another clear feature is the increase of errors with latitude which suggests that the most stable regime is the most sensitive to errors. This is also clear from the long term evolution of mean night time temperature errors over Europe as illustrated in Fig. 2. The model changes which had a big impact were in 1996 and 2007, both related to turbulent diffusion in stably stratified flow. In 1996 the diffusion coefficient for heat was increased and in 2007 the diffusion above the surface layer was reduced. The latter was detrimental for the temperature forecasts but was necessary to avoid the destruction of stratocumulus through excessive diffusion in inversions. This illustrates the multi-faceted aspects of the turbulent diffusion parametrization.

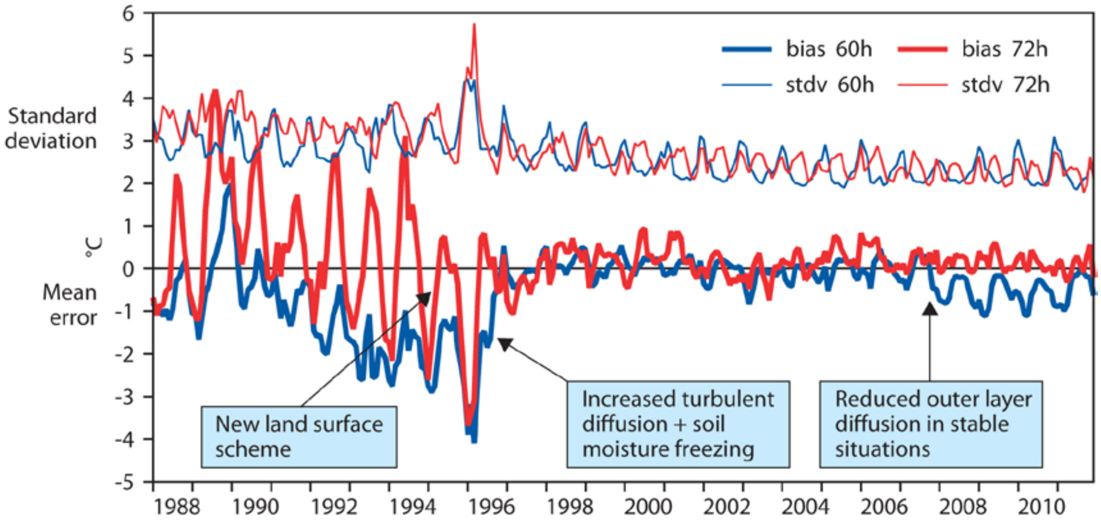

Figure 2: Historic evolution of 2m temperature errors of the operational ECMWF system. These are monthly values of mean and standard deviation of errors for step 60 and 72 hour forecasts initialized daily at 12 UTC, verifying at 0 UTC (blue) and 12 UTC(red) respectively. The verification is against about 800 SYNOP stations over Europe (30°N-72°N/ 22°W-72°E).

2. Thermal coupling between atmosphere and land surface

The thermal coupling between atmosphere and land surface is illustrated in Fig.3. At night the surface temperature T_{sk} drops due to radiative cooling Q_{net} which is negative. It should be noted that the amount of cooling is crucial and depends on the radiation scheme and all the information that goes into the radiation computations namely profiles of water vapor, temperature, aerosols, trace gas concentrations and of course clouds. Given the radiative cooling, the temperature drop at the surface depends on the amount of the turbulent coupling with the atmosphere and the diffusive coupling to the deep soil. These processes are affected by many empirical parameters and processes: turbulent diffusion in the stable boundary layer (dependent on shear, stratification and meso-scale variability), the surface roughness lengths for momentum and heat, coupling between the vegetation layer and the soil (expressed in the ECMWF model by a vegetation type dependent conductivity), the soil diffusion coefficient, the presence of snow (including thickness and density), the occurrence of soil water freezing/thawing, and terrain heterogeneity.

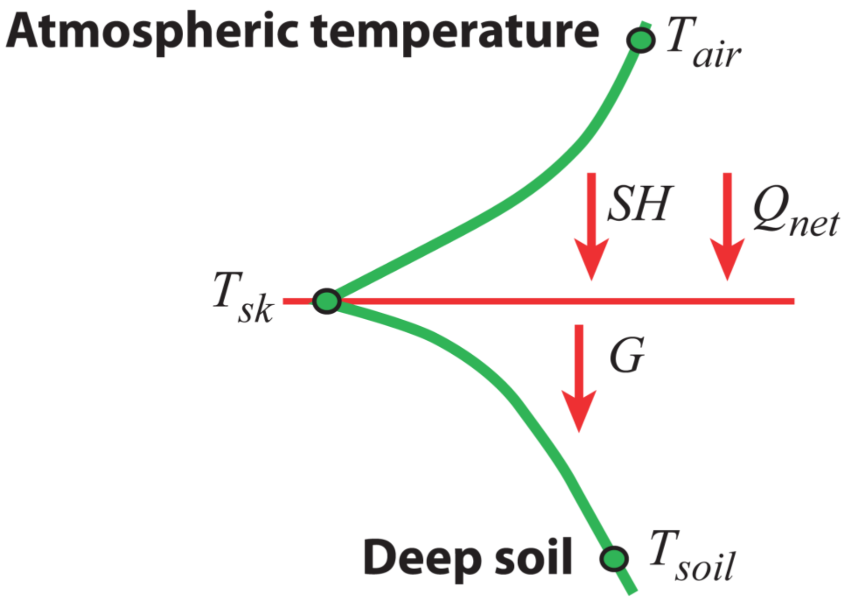

Figure 3: Schematic of thermal coupling between the atmosphere and the deep soil. The T_{sk} level is the radiation intercepting/emitting level which can be the vegetation canopy, the litter layer on top of the bare soil, the snow surface, or a combination of these. The coupling between the skin layer and the atmosphere is controlled by the boundary layer scheme. The coupling with the deep soil is controlled by soil heat diffusion, snow heat transfer, and soil freezing.

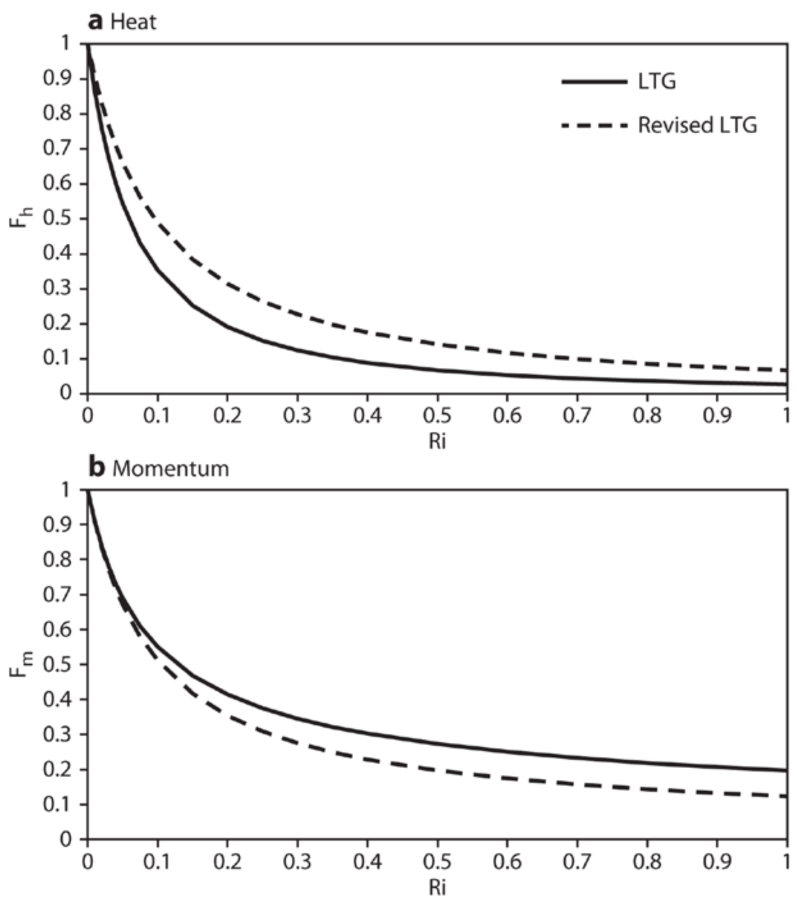

Figure 4: Stability functions for momentum (bottom panel) and heat (top panel) according to Louis, Tiedtke and Geleyn (LTG) and the revised version introduced in 1996.

A clear example is the change in 1996 in which the stability functions were changed from “LTG” to “Revised-LTG” and in which the process of soil water freezing was introduced. Both, the increased diffusion for heat and the additional thermal inertia of the soil due to the freezing process, reduced the winter cold bias. The top left panel of Fig. 5 shows the impact of the revised boundary layer scheme as published by Viterbo et al. (1999) with the 1994 version of the model. This experiment was repeated with the 2011 version of the model by implementing the original LTG and revised-LTG versions. The impact of revised-LTG versus LTG is shown in the top right panel of Fig. 5. Comparing the two top panels of Fig. 5, it is clear that the same boundary layer change has a much bigger impact in the 2011 model version than in old 1994 version. It is impossible to say which model element is responsible, as many model changes were made over the years. However, very likely candidates are the new soil hydrology scheme (Balsamo et al. 2009) and the new snow scheme (Dutra et al. 2010). In the latter, snow is a much better insulator and therefore the winter temperatures are lower. The impact of revised-LTG with the old snow scheme (bottom left panel) is also smaller than with the new snow scheme (top right). Unfortunately, the effect of revised-LTG could not be tested with the old hydrology scheme.

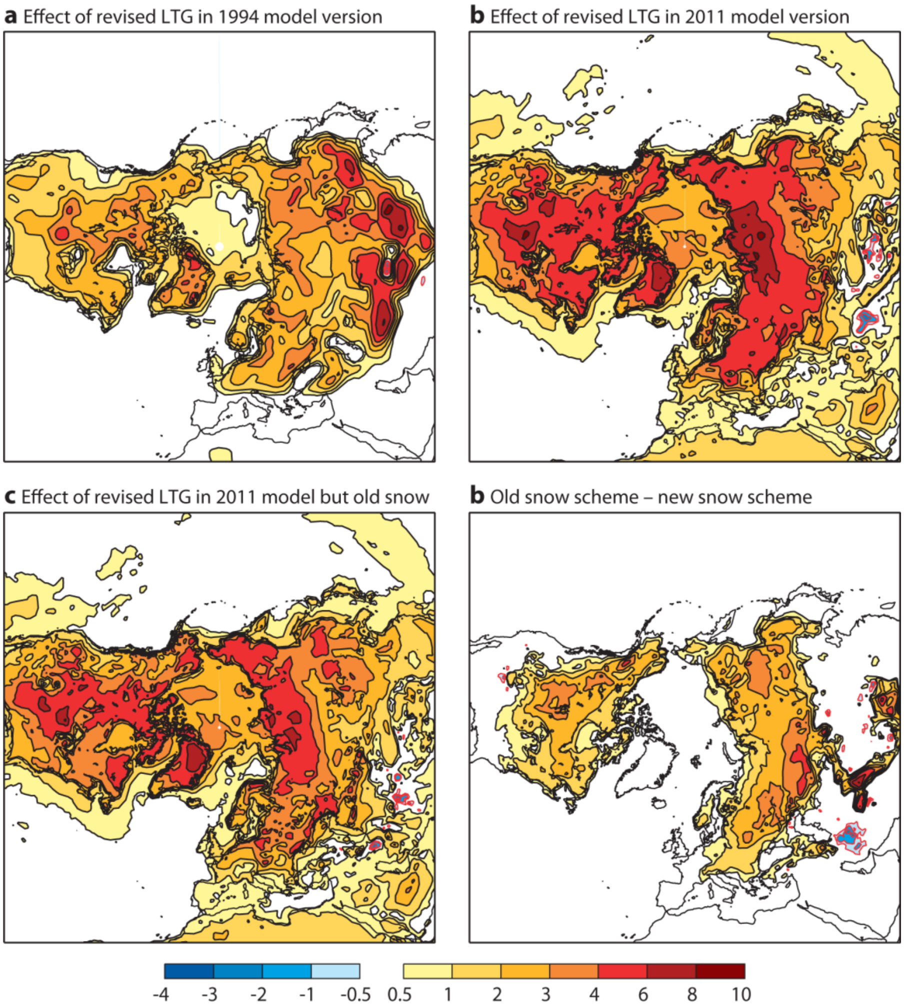

Figure 5: Effect (in terms of mean temperature change) of the model change as indicated by the figure title on averaged January 1996 temperature. These sensitivity experiments were performed by starting a long integration from 1 October 1995 and applying relaxation to the 6-hourly operational analyses above 500 m from the surface. This is an efficient way of doing “deterministic” seasonal integrations without constraining the stable boundary layer.

The conclusion is that the sensitivity of a model to stable boundary diffusion depends on other model aspects. A model with less coupling with the underlying soil (e.g. through insulating snow) is more sensitive to the details of the boundary layer diffusion than a model that has a strong thermal coupling with the deep soil. Given the uncertainty in these processes, it is very likely that reasonable diurnal cycles are obtained with compensating errors. So a key question is: what fraction of the heat flux due to radiative cooling is supplied by the soil and what fraction comes from the atmosphere? This ratio is obviously not constant and depends e.g. on wind speed, land use, soil properties and the presence of snow. The only way to make progress on this aspect is by using observations to constrain models. Betts (2006) has a proposed a scaling framework in which ground heat flux and the sensible heat flux scaled by radiative cooling is presented as a function of wind speed. Such scaling relations should be derived for different data sets (e.g. CEOP, ARM FLUXNET) and compared to models. Also the night time temperature drop and the CO2 increase could be included in such a study (Law et al. 2008).

Unfortunately, due to lack of energy closure in observations, the observed turbulent heat fluxes are not reliable particularly at low wind speed. This is not well understood, but may indicate that meso-scale motions at low winds take over the turbulent transport. Therefore it is better to use observed net radiation and ground heat flux and use the sensible heat flux as a residual. Such fluxes could also be considered as a function of the Richardson number. Hopefully, extensive use of data sets of this nature will lead to optimization of parameters in models and reduce compensating errors.

3. Wind and surface stress issues related to the stable boundary layer

Wind speed and direction are important forecast products, and are strongly affected by the boundary layer parametrization. Figs. 6 and 7 illustrate some longstanding systematic errors of the ECMWF system. The diurnal cycle of wind shows a minimum at night at the 10m level, whereas the wind is strongest at night at 200m. The latter is the result of an inertial oscillation once turbulence stops after sunset. The ECMWF model follows the observed diurnal cycle qualitatively but underestimates the amplitude near the surface and also at the top of the stable boundary layer. This is at least partially the result of the strong mixing that is applied in the model which has the tendency of smearing out the low level jet and bringing too much momentum close to the surface.

Wind direction also shows systematic biases in the ECMWF model with the a-geostrophic angle being too small. This is the case over land as illustrated in Fig. 7, but also over the ocean (Brown et al. 2005). Errors are large (about 10 degrees) at night, and in winter also during daytime. This suggests that the stable regime is the biggest contributor to the wind direction errors.

The boundary layer scheme in an atmospheric model is also responsible for surface drag which feeds back to the large scale flow through the momentum budget and through the a-geostrophic flow. The momentum budget aspect is believed to be an important contributor to the sensitivity of large scale NWP scores to the formulation of the boundary layer scheme. In general over-diffusive stable boundary layer formulations tend to give better performance for the large scale flow. The mechanism is not well understood. However, large drag puts damping on the weather systems and reduces the "activity" of a model, which tends to be good for scores. The challenge is to design a model that has the correct level of activity (i.e. the same level of activity as the analysis) and still has good scores.

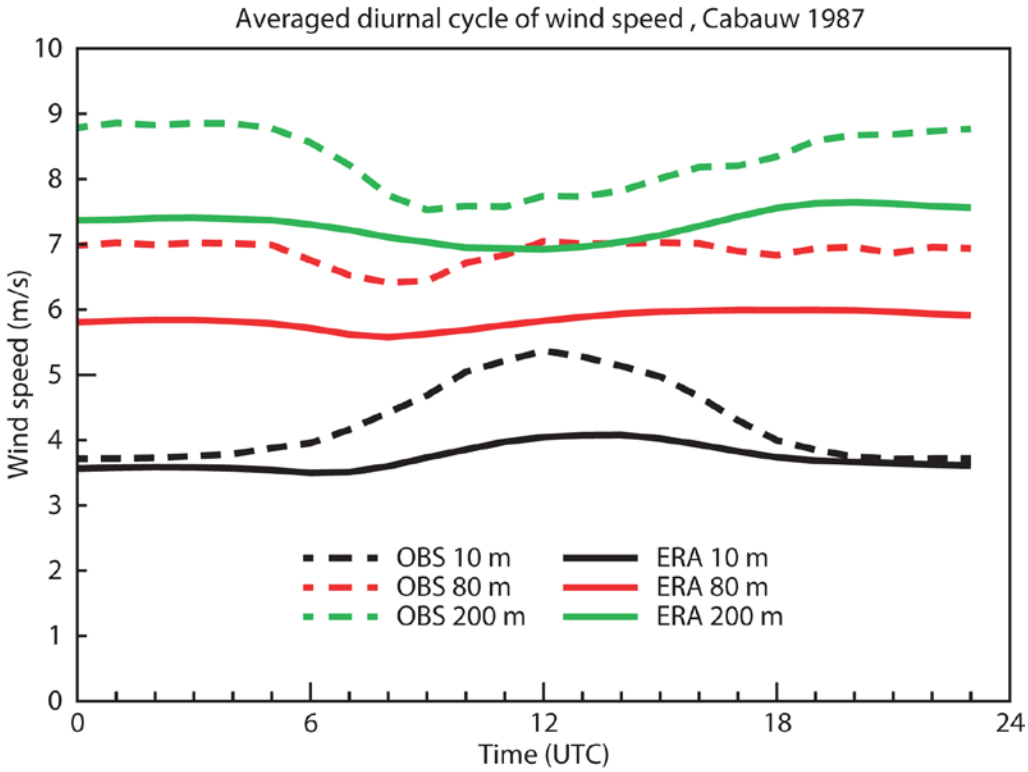

Figure 6: Annually averaged diurnal cycle of wind (1987) in Cabauw in the Netherlands at heights of 10, 80 and 200 m height. The ERA-40 daily 12-36 hour forecasts (solid) are compared to observations (dashed).

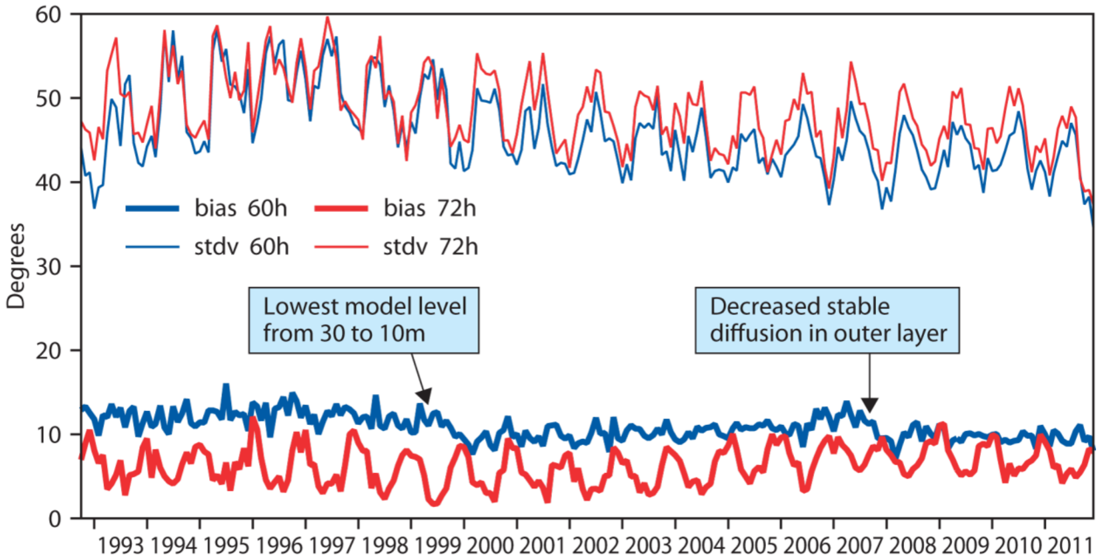

Figure 7: Historic evolution of 10m wind direction errors of the operational ECMWF system. These are monthly values of mean and standard deviation of errors for step 60 and 72 hour forecasts initialized daily at 12 UTC, verifying at 0 UTC (blue) and 12 UTC(red) respectively. The verification is against about 800 SYNOP stations over Europe (30°N-72°N/ 22°W-72°E).

Direct verification of surface drag is very difficult not in the least because in many areas over land it is controlled by heterogeneities with terrain features like high vegetation or topography exerting form drag on the flow. These aspects are characterized in models by empirical parameters like roughness length, coefficients in the turbulent orographic form drag (TOFD, Beljaars et al. 2004) and the sub-grid orography schemes (Lott and Miller 1997). This is again an area where compensating areas are likely to occur. Progress could be made by fine scale modelling of real terrain in order to find “drag laws” that can be used as a reference for parametrized models.

Often forgotten are the scales of variability between model resolved scales (say of the order of 50 km) and the turbulence scales (say a few hundreds of metres and smaller). Turbulence parametrization can be justified by a spectral gap, such that turbulence is always in quasi-equilibrium with the large scale forcing. However, such a gap is seldom found in practice. Of particular interest is the meso-scale variability in the stable boundary layer, which raises the question whether meso-scale variability (e.g. horizontal meandering of the flow with poor correlation in the vertical) can maintain sub-grid shear that contributes to the production of turbulence? To illustrate this, Fig. 8 shows the time spectrum of wind shear at about 190 and 170 m for model and observations respectively. The model shows less variability at all time scales but particularly at the shortest time scales (e.g. 1 hour) which are not part of the turbulence spectrum. It is also clear that the level of meso-scale shear is highly dependent on the model version, with particular sensitivity to the boundary layer scheme. The lack of meso-scale shear in the ECMWF model was the motivation to add such a term in the turbulence parametrization. However, this is not very satisfactory as it is not clear how such a term should scale with e.g. wind speed and how it should vary in the vertical (Mahrt and Vickers 2006).

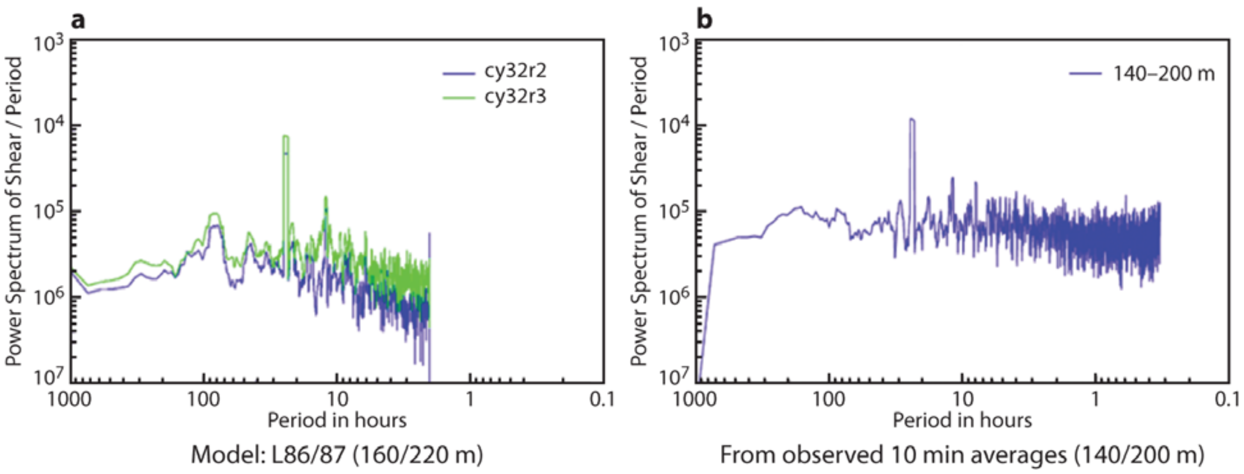

Figure 8: Time spectrum of “wind shear” from two model versions of the ECMWF model (CY32R2 in blue and CY32R3 in green; left panel) and from observations at Cabauw (right panel). The model time series has been produced by concatenating hourly output from daily 24-hour forecasts. Therefore the model data corresponds exactly in time and location to the observational data. The spectrum in the graphs have been multiplied by frequency (i.e. divided by period) to facilitate the interpretation in terms of variance as integral of the spectrum on a log scale.

Another question related to meso-scale variability is whether it could be directly responsible for some of the heat transport to the surface and whether it would explain the lack of energy closure in observations particularly at low wind speeds. All closure schemes are based on turbulence observations, but none of these observations close the surface energy budget. Is such variability related to terrain heterogeneity or is it an internal mode of the coupled atmosphere/surface system?

It can be concluded that many questions remain unanswered and that more research is needed to resolve them. The good news is that we have the model tools (e.g. LES) and the observations to address some of these questions.

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