Chapter 13

Role of the Boundary Layer in a Numerical Weather Prediction Model

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Abstract

The role of a boundary layer scheme in a numerical weather prediction model is to provide the coupling with the surface, to give realistic forecasts of near surface parameters as temperature, specific humidity and wind and to provide input to the cloud and convection schemes. It is argued that models are sensitive to surface fluxes and that turbulence parametrization is particularly sensitive in layers where steep gradients occur e.g. in the surface layer, the inversion layer and in the stable boundary layer. Developments in the ECMWF boundary layer scheme are described to illustrate this.

1 Introduction

A realistic representation of the boundary layer is an essential ingredient of current 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, 1993 and chapter 14).

The purpose of this chapter is to illustrate the role of different aspects of the boundary layer in a NWP model. This will be done by reviewing some changes that were made to the ECMWF boundary layer scheme over the years and by showing the impact in the model. The current vertical diffusion scheme is a combination of sub-schemes that apply to different situations. The surface layer, the stable boundary layer, the convective boundary layer and the upper troposphere each have formulations that were introduced to optimize different aspect of the model performance. In this chapter the different aspects of the parametrization are reviewed and compared with the original scheme as documented by Louis, Tiedtke and Geleyn (1982), hereafter referred to as LTG.

It is postulated that the main impact of the boundary layer scheme on the large scale flow is through the surface fluxes. Changes to the scheme that have impact on surface fluxes normally lead to a strong model response. It does not mean that all the emphasis is on the surface layer parametrization only. Surface fluxes are also determined by the boundary layer structure. If we think of the boundary layer as a reservoir (e.g. a mixed layer) of momentum, heat and moisture, then it becomes clear that all processes affecting the contents of the reservoir have impact on the surface fluxes. If for instance the boundary layer is ventilated from the top through dry entrainment or through moist convection, then the boundary layer becomes dryer and the surface evaporation over the ocean increases (Tiedtke et al., 1988). Also the stable boundary layer has a complicated interaction between its structure and the surface flux. The temperature structure affects the near surface temperature forecasts but through the stability effects the main impact on the large scale flow is through the momentum flux.

Another way of looking at the importance of different aspects of the boundary layer parametrization is by considering the location where steep gradients occur e.g. in the surface layer, in the stable boundary layer and at the top of the mixed layer. The fluxes are obviously very sensitive to the diffusion that is applied in such layers. On the other hand, changes to the amount of mixing in well-mixed layers, have very little impact. This will be illustrated in the sections that discuss the different parts of the parametrization.

2 The surface layer formulation

The standard way of expressing (kinematic) surface fluxes of momentum (u_*^2; square of the friction velocity), sensible heat ($\overline{w'\theta'}_o$) and latent heat ($\overline{w'q'}_o$) into wind, temperature and moisture differrences over the surface layer is with help of transfer coefficients (Brutsaert, 1982):

$$u_*^2 = \quad C_m \, |\vec{V}_1|^2 \, , \tag{1}$$
$$\overline{w'\theta'}_o = C_h \, |\vec{V}_1| \, (\theta_s - \theta_1) \, , \tag{2}$$
$$\overline{w'q'}_o = C_q \, |\vec{V}_1| \, (q_s - q_1) \, , \tag{3}$$

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, θ_1 and q_1 are potential temperature and specific humidity at the lowest model level, and θ_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 (see chapter 1). The transfer coefficients can also be written as

$$C_m = C_{mn} \, F_m(Ri_b, z_1/z_{om}, z_1/z_{oh}) \ , \tag{4}$$
$$C_h = C_{hn} \, F_h(Ri_b, z_1/z_{om}, z_1/z_{oh}) \ , \tag{5}$$
$$C_q = C_{qn} \, F_h(Ri_b, z_1/z_{om}, z_1/z_{oq}) \ , \tag{6}$$

where C_{mn}, C_{hn} and C_{qn}, are the neutral transfer coefficients i.e. the ones that contain the logarithmic part of the profile functions only, and F_m, F_h and F_q are the stability functions dependent on bulk Richardson number Ri_b $(=(g/\theta_v)\, z_1(\theta_{v1} - \theta_{vs})/|\vec{V}_1|^2)$, and the ratios z_1/z_{om} and z_1/z_{oh}. Parameter k is the Von Karman constant, z_1 is the height of the lowest model level above the surface (about 32 m in the ECMWF model), and z_{om}, z_{oh} and z_{oq} are the roughness lengths for momentum heat and moisture. Note that z_{om} is used as a displacement height to avoid singularities for high roughness lengths e.g. over steep orography. It can easily be shown that Ri_b is related to z_1/L, so the functions F_m and F_h can be derived from the Monin Obukhov stability functions for which the main body of empirical information exists (see Högström, 1988 for a review).

The parametrization introduced by Louis (1979) and modified by LTG (1982) has empirical functions for F_m and F_h which are based on scaling arguments, but their precise form is inspired by the model's ability to forecast large scale flow patterns rather than by observations of Monin Obukhov functions. The dependence on roughness lengths is simplified and exists only for unstable situations and no distinction is made between z_{om}, z_{oh} and z_{oq}. The neutral transfer coefficients for momentum, heat and moisture are the same in the LTG model i.e. the roughness lengths for heat and moisture are equal to the one for momentum. For the ocean, these roughness lengths are computed with help of the Charnock relation ($z_{om} = z_{oh} = z_{oh} = 0.018\, u_*^2/g$), resulting in an increase of transfer coefficients with wind speed, which is realistic for momentum, but rather unrealistic for heat and moisture (DeCosmo, 1991).

The current ECMWF formulation uses Monin Obukhov similarity functions dependent on z/L and solves the relation between z_1/L and Ri_b iteratively giving maximum flexibility for the specification of stability functions and surface roughness lengths[1]. The Paulson (1970) functions are used for unstable situations and the ones documented by Beljaars and Holtslag (1991) are used for stable situations.

Apart from the stability functions also the surface roughness lengths have been changed. Over land the roughness lengths for heat and moisture have been reduced with respect to momentum. A factor 10 reduction is used for vegetation and a further reduction is used for areas with orographic enhancement of the aerodynamic roughness length (see Mason, 1991). Over the ocean, smooth surface scaling is added to the Charnock relation for momentum and smooth surface scaling only is used for heat and moisture (see Beljaars, 1995a):

$$z_{om} = 0.11\,\nu/u_* + 0.018\,u_*^2/g \quad , \tag{7}$$

$$z_{oh} = 0.40\,\nu/u_* \quad , \tag{8}$$

$$z_{oq} = 0.62\,\nu/u_* \quad , \tag{9}$$

where ν is the kinematic viscosity of air ($1.5\,10^{-5}\,m^2/s$).

The free convection behaviour of the transfer formulation is improved by introducing a gustiness component in the absolute horizontal wind in addition to the resolved horizontal velocity $(U_1^2 + V_1^2)^{1/2}$:

$$|\vec{V}_1| = (U_1^2 + V_1^2 + \beta w_*^2)^{1/2} \; , \quad w_* = (z_i\, g/\theta_v\, \overline{w'\theta'_{vo}})^{1/3} \; , \tag{10}$$

with z_i for boundary layer height and $\beta = 1$. The choice of the boundary layer height is not very critical so a fixed value of $1000\,m$ is used. The w_* term is based on the idea that when the grid averaged large scale velocity components U_1 and V_1 become very small, the convection itself maintains a finite horizontal velocity at the subgrid scale. This motion is driven by boundary layer size eddies which scale with the free convection velocity w_* (Liu et al., 1979; Schumann, 1988; Godfrey and Beljaars, 1991; Beljaars, 1995a). The advantages of the new surface layer formulation are: (i) it has a better free convection limit for evaporation which is highly relevant for the tropical circulation (Miller et al., 1992), (ii) the ocean transfer coefficients for heat and moisture agree better with recent observations (Bradley et al., 1991; DeCosmo, 1991), and (iii) the evaporation from wet land surfaces is reduced which is in better agreement with observations (Beljaars and Viterbo, 1994).

The impact of the improved free convection limit was substantial when it was implemented in 1990. The reason is that the tropical circulation is very

[1] A number of recent papers discuss alternatives to the iterative conversion from bulk Richardson numbers to Obukhov lengths e.g. Buyn, 1990; Launiainen, 1995; Mascart et al., 1995; Uno et al., 1995; Lo, 1996; van den Hurk and Holtslag, 1997.

sensitive to the SST's in the so-called warm pool of the Western Pacific where wind speeds tend to be low. Because winds are weak, the moisture jump over the surface layer is large and depends strongly on the specification of air-sea transfer coefficients in the surface layer. To illustrate the magnitude of the change, an example is given in Fig. 1. For a typical air-sea difference of 1.5 K and 7 g/kg, the latent heat flux is increased from 5 to 40 W/m^2 at zero wind speed. Initially (in 1990) the improved free convection limit was not implemented as described above, but in a more empirical way restricting the change to winds below 10 m/s (see Fig. 1). The impact was most noticeable in the tropical precipitation and the tropical wind errors. The old model had a dry zone over the warm pool in the Western pacific and a tendency to produce a double Inter Tropical Convergence Zone (ITCZ). The improved scheme increases the precipitation over the Western Pacific and virtually eliminates the split ITCZ (Fig. 2). The increased latent heat release in the concentrated ITCZ enhances the strength of the Hadley circulation and leads to reduced Easterly errors in the upper troposphere as shown in Fig. 3 (Miller et al., 1992). The impact on the tropical circulation can be interpreted in terms of Gill's linear model as a response to the latent heat release in the Western Pacific (Gill, 1982).

Not all these changes in the surface layer formulation were directly beneficial in the ECMWF model as shown by Beljaars (1995b). The reduced

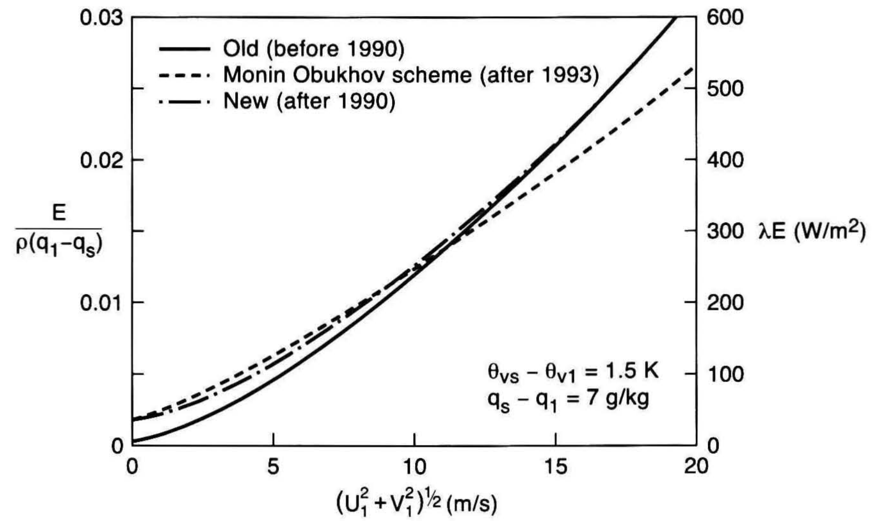

Fig. 1. The conductivity of the lowest model layer for moisture transfer (left hand scale) as function of wind speed. The reference height is 32 m, which is the height of the lowest model level in the ECMWF model. The right hand scale indicates the latent heat flux for a typical temperature difference of 1.5 K and a specific humidity difference of 7 g/kg. The solid line represents the old scheme (as used by Louis et al., 1982), the dashed line represents the current scheme (operational after 1993) and the dash-dot line represents the empirical implementation of 1990 which limits the impact to low wind speeds only.

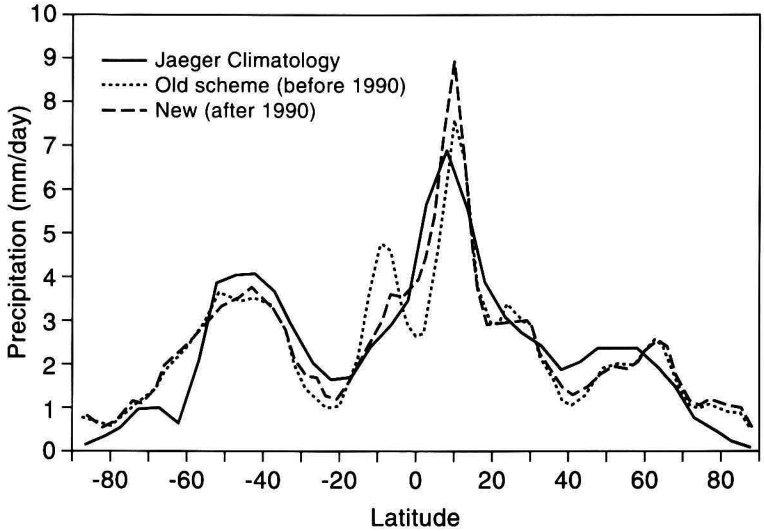

Fig. 2. Zonal mean rainfall averaged over 90 days of a T42, June/July/August model integration with the old scheme, and the new scheme (as introduced in 1990) in comparison with the climatology by Jaeger.

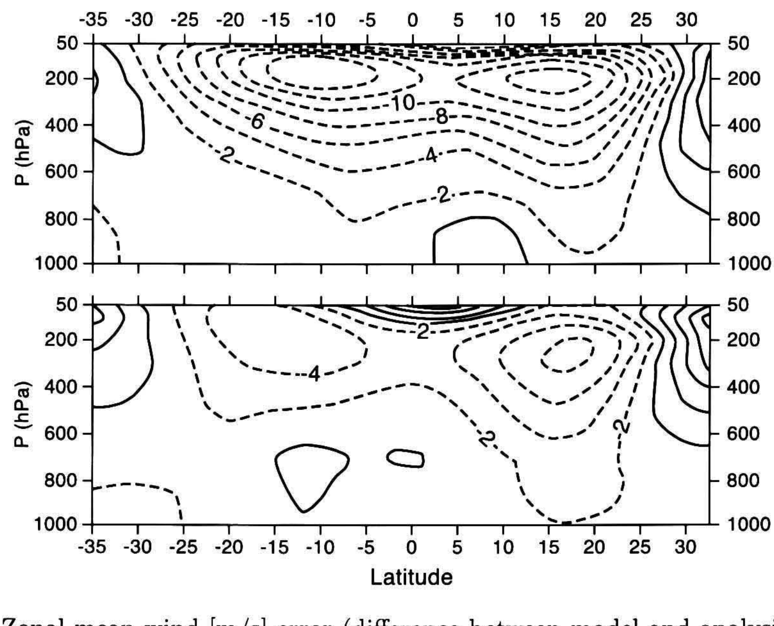

Fig. 3. Zonal mean wind [m/s] error (difference between model and analysis) averaged over 90 days for a December-January-February T42 integration with the old scheme (upper panel) and the new scheme with enhanced air-sea coupling at low winds as introduced in 1990 (lower panel).

air-sea interaction at high winds as the result of much smaller sea surface roughness for heat and moisture than for momentum (introduced in 1993), was clearly in better agreement with observations (Beljaars, 1995b; DeCosmo, 1991), but was detrimental in a winter data assimilation/forecast experiment. This was due to compensating errors; the mass flux scheme for convection scheme did not extract enough moisture from the boundary-layer, leading to an underestimation of evaporation which was compensated for by large transfer coefficients. A change in the closure of the convection scheme corrected this problem. This is a clear reminder that boundary-layer turbulence is not an isolated process in a large scale model, but it interacts with other processes as convection and land surface processes. It also implies that different schemes may perform differently in different models.

In conclusion, it can be stated that the surface layer formulation based on Monin Obukhov similarity functions is well established now in NWP. It works very well over homogeneous terrain, provided that the boundary-layer scale eddy motion is accounted for in the near surface wind and that proper surface roughness lengths are prescribed in particular over the ocean. The model sensitivity to air-sea transfer is large because the gradients near the surface are large.

For heterogeneous terrain, large uncertainties exist with respect to the atmosphere-surface transfer. The concept of effective roughness lengths can be used i.e. roughness lengths are modified in such a way that the correct surface fluxes are obtained in the model (see Mahrt, 1996 for a review). In practice, however, it is difficult to derive such effective parameters because the available global data sets do not have sufficient resolution. The momentum fluxes over land are of particular concern, because model sensitivity to these fluxes tends to be large and the impact of terrain heterogeneity on momentum fluxes is considerable (Beljaars, 1995b).

3 Mixed layer and entrainment formulation

Diffusion in the unstable boundary layer of the ECMWF model used to be parametrized with help of diffusion coefficients dependent on local stability (LTG, 1982). This implies that potential temperature has to decrease with height in the unstable boundary layer in order to maintain upward diffusion. In practise, these gradients are very small so the LTG formulation still maintains a rather realistic well mixed structure for potential temperature. The main disadvantage is that diffusion in the inversion layer is modeled with the stable part of the parametrization and therefore entrainment is negligible. This deficiency was illustrated by Betts et al. (1993) in a comparison with observations from the FIFE experiment. It was shown that the boundary layers produced by the LTG scheme were too moist and too shallow due to a lack of entrainment at the top of the boundary layer.

In order to improve on entrainment, the K-profile scheme propose by Troen and Mahrt (1986) was tried without the counter-gradient term (fluxes just proportional to gradients). This scheme determines the boundary layer height h and expresses the diffusion coefficients as a function of z/h. For $z/h > 0.1$ the formulation is as follows

$$K_m = \; k\, z\, w_s (1 - \frac{z}{h})^2 , \tag{11}$$

$$K_h = k\, z\, w_s (1 - \frac{z}{h})^2 / Pr , \tag{12}$$

where $w_s = (u_*^3 + C_1\, w_*^3)^{1/3}$, $C_1 = 0.6$, Pr is the turbulent Prandtl number evaluated from surface layer similarity at $z/h = 0.1$. One of the crucial elements of the scheme is the estimation of h, which is done by lifting a parcel from the surface layer with excess temperature $\Delta\theta = D\, \overline{w'\theta'_{vo}}/w_s$ with $D = 6.5$ until a Richardson criterion is met, which is always somewhere in the capping inversion.

From single column simulations it became apparent that the implied entrainment from this scheme was rather strong and very much dependent on the details of the boundary layer height algorithm and also dependent on vertical resolution. In order to get a well controlled entrainment rate and a less aggressive erosion of inversions, the scheme was modified in the following way: (i) the point of zero buoyancy is determined rather than using a Richardson criterion, (ii) the parcel is lifted from the minimum virtual temperature level instead of the surface layer, (iii) coefficient D for the excess parcel temperature is reduced from 6.5 (Troen and Mahrt, 1986) or 8.5 (Holtslag and Boville, 1993) to a value of 2, and (iv) the diffusion in the entrainment layer (the layer between model levels where h happens to be) is prescribed such that a negative buoyancy flux is obtained of 20% of the surface buoyancy flux:

$$K^E_{m,h} = C_E \frac{\overline{w'\,\theta'_{vo}}}{(d\theta_v/dz)_E} \quad , \quad C_E = 0.2 , \tag{13}$$

where, $(d\theta_v/dz)_E$ is the virtual potential temperature gradient in the inversion layer. The reason for prescribing a diffusion coefficient rather than a flux is technical; a specified K together with an implicit solver guarantees numerical stability for long time steps. Details of this scheme are given by Beljaars and Betts (1993).

How different parametrization schemes handle the mixed layer is illustrated in Fig. 4. Two versions of the K-profile scheme and one version of the local stability closure are used in a single column simulation of a hypothetical mixed layer. The K-profile scheme is used in its original form but without counter-gradient term (Troen and Mahrt, 1986), and in the form described above including prescribed entrainment in the inversion layer (equation 13).

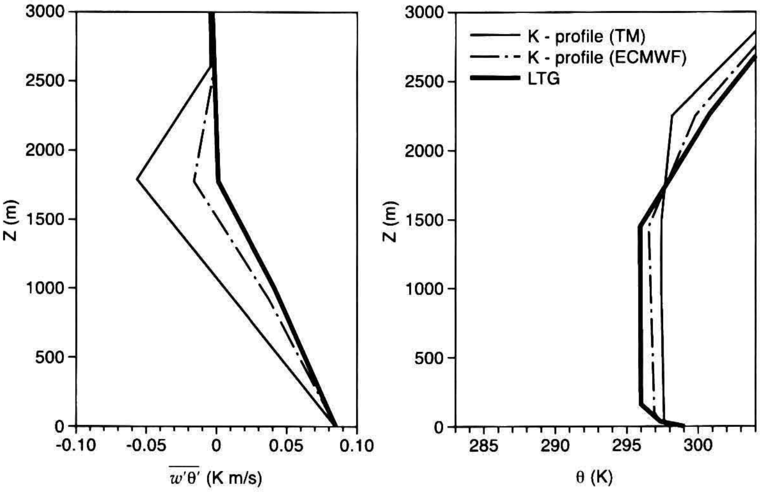

Fig. 4. Kinematic heat flux (left panel) and potential temperature (right panel) profiles after 9 hours of integration with three versions of a single column model. A stable layer is heated from the surface with 100 W/m^2 over a period of 9 hours. The geostrophic wind is 10 m/s and the surface roughness length is 0.1 m. The three models have different boundary layer schemes: (i) the K-profile according to Troen and Mahrt (1986) without the counter-gradient term, (ii) the K-profile scheme with entrainment parametrization as implemented in the ECMWF model, and (iii) LTG scheme (Louis et al., 1982) with diffusion dependent on local stability.

The model version with local stability closure uses the LTG functions. The hypothetical case consists of a stable layer that is eroded by heating from the surface with 100 W/m^2 during 9 hours. Profiles of heat flux and potential temperature are shown at the end of the 9 hour interval (Fig. 4). The flux profiles are nearly linear and the heating rate is determined by the heat flux at the surface and by the heat flux in the entrainment layer.

In this example with specified heat flux at the surface, differences in temperature at the end of the 9 hour interval can only occur as the result of differences in entrainment. The local stability closure has no entrainment and the two K-profile schemes have different levels of entrainment. The original version by Troen and Mahrt (1986) has an entrainment that is determined by subtle details of the boundary layer height computation and by vertical resolution which makes the resulting entrainment rather arbitrary and far too big in this particular example. With the scheme implemented in the ECMWF model, the entrainment is controlled and determined by coefficient C_E which is set to 0.2 (Stull, 1988).

Holtslag et al. (1995) argue that the mixed layer evolution and the boundary layer height are sensitive to the presence of a counter-gradient parametrization. This sensitivity was also found for the example described above (results not shown). However, our interpretation is rather different. It turns out that the counter-gradient term affects the diffusion in the inversion (entrainment) with the original K-profile scheme. The reason is that with

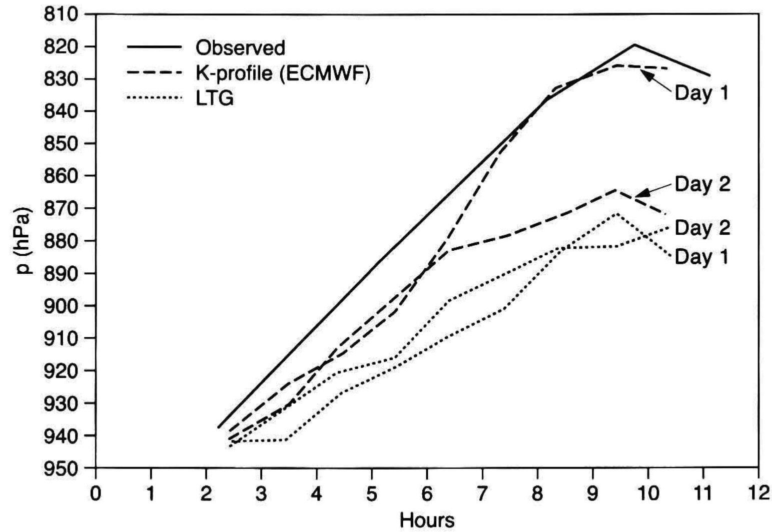

Fig. 5. Composite of 9 diurnal cycles of FIFE observations of boundary layer height in August 1987 in comparison with forecast (0-24 and 24-48 hours) from two versions of the ECMWF model (the LTG scheme and the K- profile scheme with entrainment parametrization).

the counter-gradient term the potential temperature profile tilts slightly and therefore the lifted parcel reaches the critical Ri-number at a slightly different level in the inversion, resulting in a different diffusion coefficient in the inversion layer and therefore a different entrainment.

It is also interesting to note that schemes with diffusion dependent on local stability are not very sensitive to the precise form of the stability functions. LTG and M.O. functions have been tried and they result in substantial differences in diffusion coefficients but the heat flux and temperature profiles can hardly be distinguished (results not shown). The reason is that both closures have virtually no entrainment and therefore the heat flux profile and the heating rates are the same. The diffusion is sufficiently strong to maintain a well mixed layer independent of the precise magnitude of the diffusion coefficients in the mixed layer.

Experimentation with the full 3D ECMWF model has shown that the entrainment formulation has a clear beneficial impact on the boundary layer evolution and on boundary layer moisture over land. Results, presented by Beljaars and Betts (1993), indicate a drying and deepening of the boundary layer which is in better agreement with the FIFE observations (Figs. 5 and 6). The K- profile/entrainment sheme still underestimates the boundary layer depth which is consistent with the analysis of FIFE data by Betts et al. (1990) who suggest that for the FIFE situation the downward buoyancy flux at the top of the boundary layer is much bigger that 20% of the surface value. Operational verification of the ECMWF model with synoptic observations also shows the benefit of the entrainment parametrization; particularly the diurnal cycle of

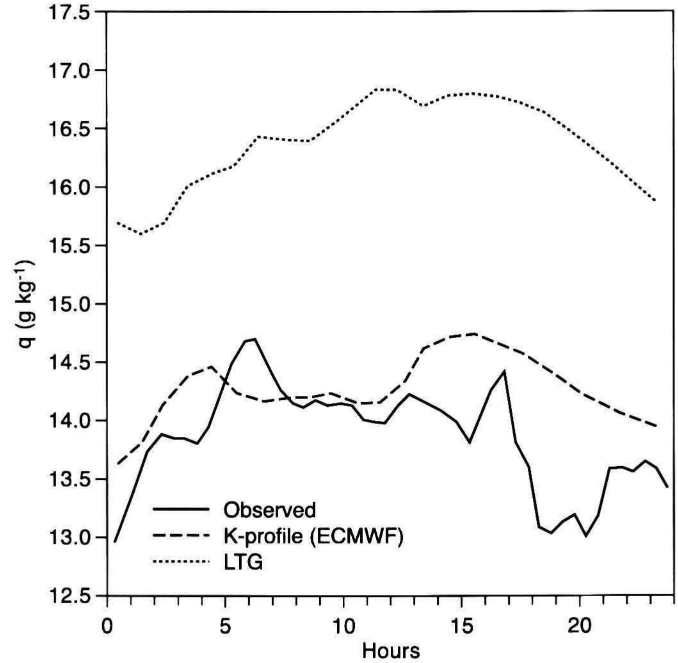

Fig. 6. Composite of 9 diurnal cycles of FIFE observations of specific humidity in August 1987 in comparison with 0-24 hour forecast from two versions of the ECMWF model (the LTG scheme and the K-profile scheme with entrainment parametrization).

specific humidity shows clear improvements (Beljaars and Betts, 1993).

4 The stable boundary layer

The capability of the ECMWF model to describe large scale synoptic developments is very much influenced by the stable boundary layer formulation particularly in winter over the Northern Hemisphere. This is related to the atmospheric momentum budget which is affected by the drag at the surface as produced by the vertical diffusion scheme. The impact of turbulent momentum transfer is felt at the largest scales through drag from entire continents, but also at the synoptic scale where the drag at the surface provides so-called Ekman damping on the cyclones. The structure of the stable boundary layer and the drag at the surface is controlled by the stability functions whether they are formulated directly in terms of the Richardson number or in terms of the Obukhov length. The LTG functions are not based on observational material but are inspired by model performance. Experiments with more observationally based stability functions gave always bad results (e.g. in Northern Hemisphere 500 hPa anomaly correlation, see Beljaars 1995a). This also applies to the stable part of the Troen and Mahrt (1986) scheme and therefore these two options were never considered for operational implementation in the ECMWF model.

Another aspect of the stable boundary layer parametrization is its capability to simulate the night time and winter time near surface temperatures and to simulate the sensible heat flux towards the surface. A problem with the LTG scheme (as operational in the ECMWF model for the stable boundary layer before 1996) became apparent after introduction of a new land surface scheme in 1993 (Viterbo and Beljaars, 1995). This new land scheme does not have a deep soil climatological boundary condition for temperature, but computes the diurnal and seasonal soil temperature evolution through successive short range forecasts. The 2 m temperature forecasts for winter turned out to be too cold and the soil temperatures were drifting cold on a seasonal time scale.

Diagnostics with help of tower observations from Cabauw in The Netherlands, indicated that the lowest model level temperature and the soil temperatures decoupled in an unrealistic way from the upper air temperatures (see Beljaars, 1995b). This result suggested a lack of diffusion of heat in very stable situations. The decoupling mechanism is related to a very fundamental characteristic of the stable boundary layer namely a positive feedback that amplifies excessive surface cooling (Derbyshire, 1997). If the surface cools, the heat diffusion will normally increase and compensate for the cooling, but in very stable situations the effect may be the opposite because the layer becomes more stable and therefore heat diffusion becomes less efficient.

The traditional Louis scheme expresses the fluxes into gradients with help of diffusion coefficients where the diffusion coefficients for momentum (K_m) and heat/moisture (K_h) depend on local stability

$$K_{m,h} = l^2_{m,h} |d\vec{V}/dz| F_{m,h}(Ri)\,, \tag{14}$$

$$l^{-1}_{m,h} = (kz)^{-1} + \lambda^{-1}_{m,h}\,, \tag{15}$$

where the stability functions F_m and F_h depend on the local Richardson number $Ri = (g/\theta_v)(d\theta_v/dz)|d\vec{V}/dz|^{-2}$ and the asymptotic mixing lengths λ_m and λ_h have values of 150 and 450 m respectively in the LTG scheme. The stability functions have the following form

$$F_m = \frac{1}{1 + 2bRi(1 + dRi)^{-1/2}}\,, \tag{16}$$

$$F_h = \frac{1}{1 + 3bRi(1 + dRi)^{1/2}}\,, \tag{17}$$

where $b = 5$ and $d = 5$.

It is interesting to note that equation (14) is fully consistent with local

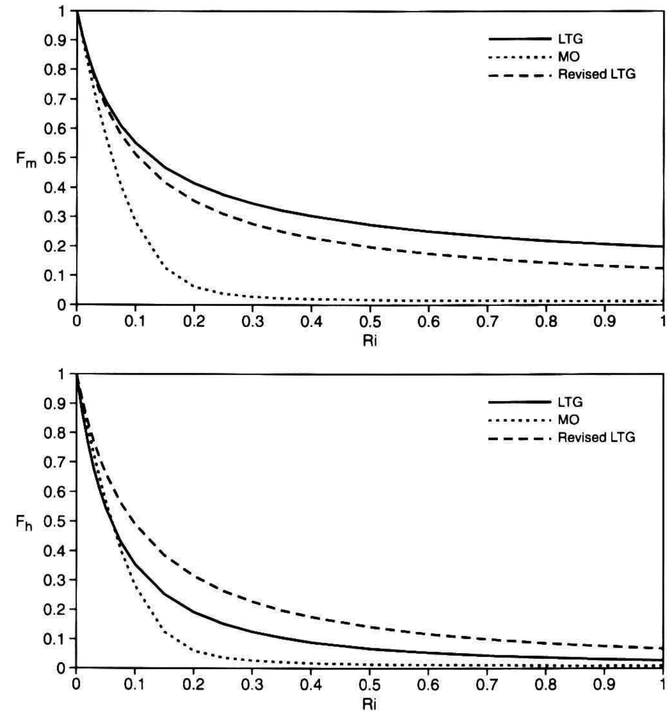

Fig. 7. Three different forms of the stability functions for momentum (upper panel) and heat (lower panel) for positive Richardson numbers.

scaling as suggested by Nieuwstadt(1984) and that functions F_m and F_h can be derived from observed Monin Obukhov stability functions. Fig. 7 shows three different forms of such stability functions. The first set of closure functions are according to LTG (equations 16 and 17), and show strong diffusion for momentum and weak diffusion for heat. The rather high diffusion coefficients for momentum at high Richardson numbers is to maintain sufficient levels of surface drag in the model in very stable situations (see also Delage, 1988a,1997).

The second set is labeled Monin Obukhov scheme because the functions have been derived from the Monin Obukhov similarity functions as they are used in the surface layer (Beljaars and Holtslag, 1991). This particular form is supported by observations and is in the range of experimental material reviewed by Högström (1988). The main point of the MO stability functions is that they are based on observations relating gradients to fluxes. At the same time it has to be emphasized that the 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 subgrid variability (i.e. due to katabatic flow or surface heterogeneity), are not represented (see e.g. Mahrt, 1987; Derbyshire, 1995). Characteristic of the MO formulation is that the turbulent Prandtl number ($Pr = K_m/K_h$) is close to one and that the diffusion for high Ri-numbers (e.g. above 0.2) is very weak.

Pr-numbers near 1 are supported by observations at low Ri- numbers, but highly uncertain at high Ri-numbers, where gravity waves may transport momentum without transporting heat (Mahrt, 1985; Kim and Mahrt, 1992).

The third form is the so-called revised LTG scheme, which is a retuned LTG form with smaller diffusion for momentum and stronger diffusion for heat. This has been obtained by changing coefficient d in equations (16) and (17) from 5 to 1, and by changing the factor 3 in the denominator of equation (17) to 2 as in the equation for F_m. Also the asymptotic mixing length was put to 150 m for both momentum and heat, but this has little impact on the boundary layer. The resulting turbulent Prandtl numbers 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 is entirely empirical and the result of a pragmatic approach to the surface temperature drift problem in the ECMWF model. The idea was to increase the diffusion of heat without affecting the momentum budget in the model. To limit the impact on the momentum budget it was necessary to decrease the diffusion for momentum, because the increased heat diffusion makes the boundary layer generally less stable.

To show the characteristics of the stable boundary layer with these three closure functions, equilibrium solutions have been generated in single column mode. The Ekman equations are integrated in time for 9 hours for a geostrophic wind of 10 m/s, a surface roughness length of 0.1 m and downward surface heat flux of 25 W/m^2. The initial profiles are uniform for wind (equal to the geostrophic wind) and potential temperature (20^o C). Profiles of heat flux, momentum flux, and potential temperature are shown in the Figs. 8a, 8b, and 8c respectively after 9 hours of integration. Both the LTG and the revised LTG schemes have a boundary layer that is about 500 m deep, whereas the MO scheme results in a momentum boundary layer depth of about 200 m. The impact of the schemes on the heat flux is very different: LTG and MO have shallow temperature boundary layers, whereas the depth over which heat is transported with the revised LTG scheme is much larger. This is also clear from the potential temperature profiles: LTG and MO show a cooling from 20^o C to 15.5^o C whereas the revised LTG scheme shows less cooling (temperature drops to 17^o C). To illustrate the positive feedback from the stable boundary layer parametrization, a second integration is done with a 50 W/m^2 downward heat flux at the surface. The temperature drop is not doubled but tripled, which is the result of the positive feedback from stability (see temperature profiles in Fig. 8d).

One of the remarkable features of the LTG and the revised LTG schemes is that the boundary layer depth is about 500 m, which is excessive for a geostrophic wind of 10 m/s. The MO scheme has the much more realistic boundary layer depth of about 200 m (e.g. Zilitinkevich, 1972; Nieuwstadt, 1981). Unfortunately, a realistic boundary layer depth does not seem to be compatible with reasonable surface drag and surface heat flux characteristics

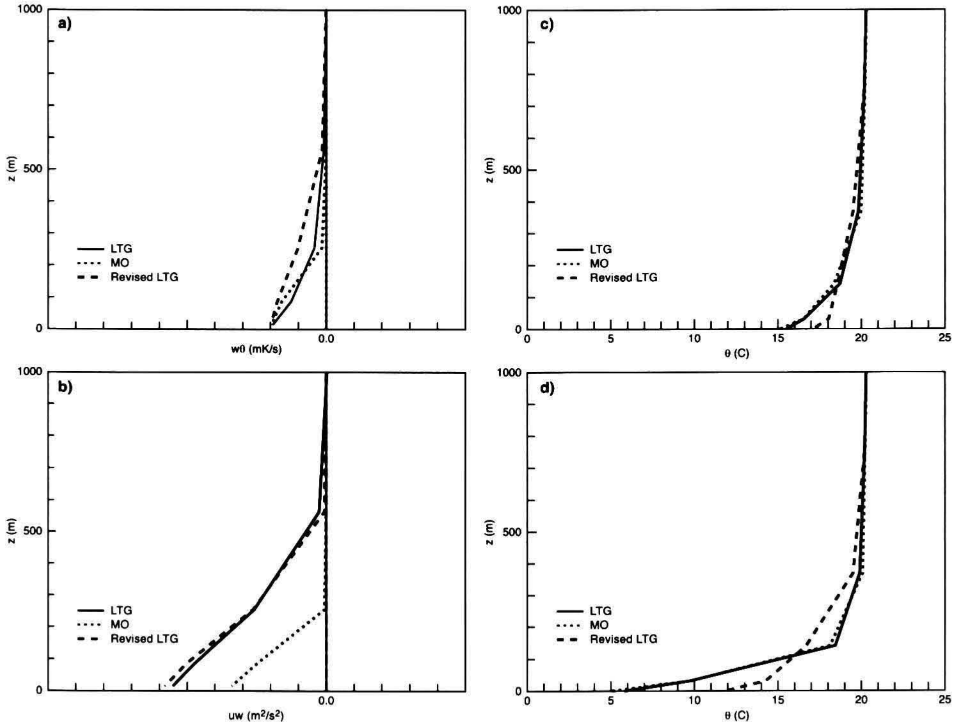

Fig. 8. Single column simulations where a neutral boundary layer is cooled by a downward surface heat flux of 25 W/m^2 over a period of 9 hours. The geostrophic wind is 10 m/s in the x-direction and the surface roughness length is 0.1 m. Three different schemes are used: The LTG scheme (Louis et al., 1982), the Monin Obukhov scheme (MO) and the revised LTG scheme. Profiles of kinematic heat flux (a), kinematic momentum flux in the direction of the geostrophic wind (b), the potential temperature profile (c) and the potential temperature profile with a surface heat flux of 50 W/m^2 (d) are shown.

in the ECMWF model. The poor vertical resolution could be blamed, but it is not a major issue in this case as shown by Beljaars (1992). Delage (1988b) explores vertical resolution in more detail and concludes that compensating errors make the results more accurate than could be expected from a numerical solution with such poor vertical resolution.

The revised LTG scheme reduces the cooling at the surface (compared to the original LTG scheme) and therefore reduces the risk of the surface to decouple from the atmosphere. Another way to prevent this decoupling is by increasing the thermal inertia of the land surface on the seasonal time scale. This is done in the ECMWF model by introducing the latent heat release due to soil moisture freezing/thawing. The practical implementation is through an apparent heat capacity in the soil temperature range between $-3^o\ C$ and $1^o\ C$. This mechanism puts a thermal heat barrier around freezing and obviously reduces the annual cycle of soil temperatures in areas where the soil temperature drops below zero (Viterbo et al., 1998).

The effect of the revised LTG scheme and the introduction of soil moisture freezing in the ECMWF model are shown in Fig. 9. These are time series of temperature at the 2 m level from long integrations from the 1^{th} of October 1995 to the 31^{th} of January 1996 in which the upper air is relaxed towards the operational analysis. The advantage of the upper air relaxation is that the same realization of the atmosphere can be obtained in every long integration and it also allows a comparison with observations for that particular season. Fig. 9 shows a time series of day time temperature over Germany for a control run, one with the revised LTG scheme, and one with the revised LTG scheme plus soil freezing in comparison with observations. It is clear that both model changes reduce the temperature errors. Half of the reduction at the 2 m level comes from the vertical diffusion change and the other half from the soil freezing. The impact on soil temperatures (not shown) is also beneficial and comes predominantly from the introduction of soil freezing (Viterbo et al., 1998).

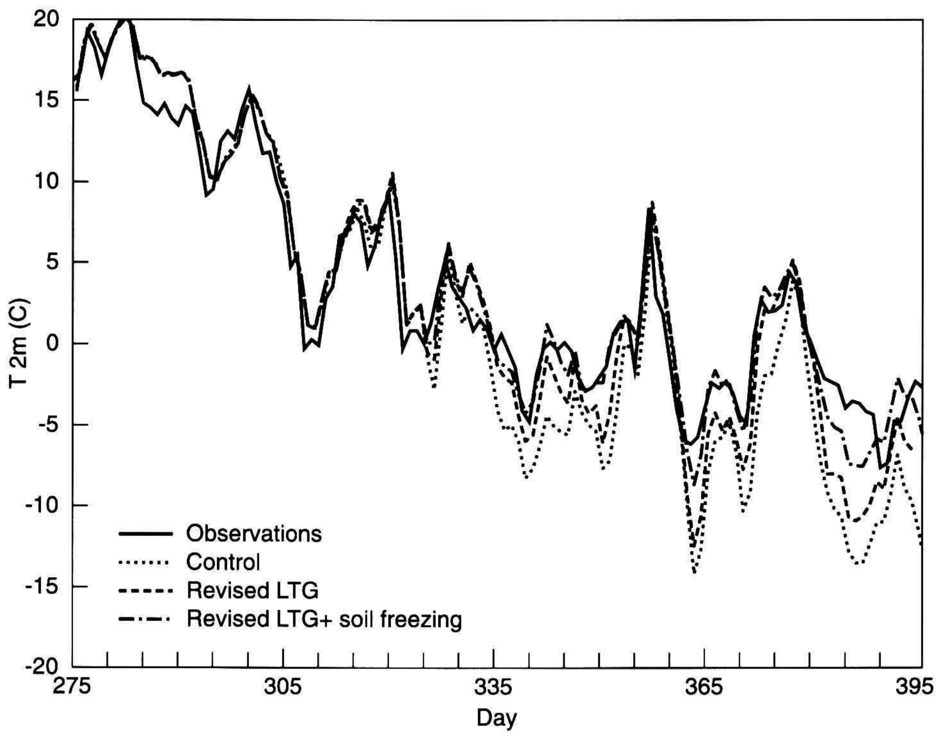

Fig. 9. Daytime (12 UTC) temperature at 2 m height over Germany from long integration with three different versions the ECMWF model in comparison with observations. The upper air is relaxed towards the observed atmospheric state in order to reproduce the time evolution of the large scale flow from 1 October 1995 to 31 January 1996. Three different stable boundary layer formulations are used: (i) LTG which is the control experiment, (ii) revised LTG, and (iii) revised LTG plus soil moisture freezing.

5 Discussion

The current formulation of the ECMWF boundary layer has evolved from the original Louis, Tiedtke and Geleyn (1982) scheme to different schemes for different parts of the boundary layer. Three fundamental changes are described: (i) a new surface layer formulation based on Monin Obukhov similarity with gustiness to account for the free convection limit, (ii) an entrainment parametrization for the dry mixed layer, and (iii) modified stability functions for the stable boundary layer. A common feature in these model changes is that mixing is changed at a location in the profile where steep gradients occur. This is believed to be the reason that strong impact is seen. The surface layer formulation influences the surface fluxes over the ocean with impact on the tropical circulation. The boundary layer top entrainment influences boundary layer ventilation and significant impact can be seen on the diurnal evolution of boundary layer height and on boundary layer moisture. Finally, the stable boundary layer formulation has a strong influence on the momentum budget and on the near surface temperature evolution of the atmosphere in winter over continental areas.

In contrast to this, model sensitivity to the details of the mixed layer parametrization is small. As long as the parametrization maintains a well mixed structure, the exact magnitude of the diffusion coefficients is of secondary importance. Also the direct impact of the counter-gradient effects is small. Indirect effects were found in the Troen and Mahrt scheme, but they were caused by changes in the entrainment due to small differences in the boundary layer height computation as the result of tilting of the potential temperature profile.

The surface layer formulation (at least for homogeneous surfaces) and the entrainment formulation have a sound background in the sense that ample research exists to support these schemes. This is important because model performance is not always a reliable source of information: incorrect formulations may be compensating for other model deficiencies and therefore improvements on a single process can lead to worse forecasts. The situation with the stable boundary layer is rather different. The formulation that is currently in use in the ECMWF model is not based on observational material (e.g. observed Monin Obukhov stability functions), but on model performance (see also Delage, 1997). The formulation has been adjusted in such a way that reasonable night time and winter time near surface cooling is obtained, but the simulated boundary layers tend to be too deep over flat terrain. This is very unsatisfactory, because it is clear that the stable boundary layer interacts strongly with other parts of the model and therefore compensating errors are not to be excluded. For instance inclusion of soil moisture freezing in the land surface scheme alleviated the winter time cooling problem considerably. The stable boundary layer may also suffer from biased radiation in the model, which is a common feature of many atmospheric models (Garratt et al., 1993, Garratt, 1994; Wild et al., 1995). At this stage it is not known how it influences the near surface winter temperatures and the night time cooling.

To what extent the atmosphere controls the near surface temperatures by turbulent transport, by radiative transfer or by meso-scale exchange is still an unsolved problem (Kim and Mahrt, 1992). Parametrization in large scale models could benefit considerably from more research on subgrid scale transport in the stable boundary layer including full consideration of the surface energy balance and the effects of heterogeneous terrain. For instance, many observations fail to satisfy the surface energy balance to any degree of accuracy and are therefore difficult to interpret. The problem is a highly coupled one, in which turbulent transport needs to be studied in conjunction with land surface processes, radiation and mesoscale effects (Derbyshire, 1997).

The interaction of the boundary layer with other processes should have more attention in general. We have seen that the interaction with the surface plays an important role, but also boundary layer ventilation at the top is an important aspect (see Tiedtke et al., 1988 for the moist aspects of boundary layer ventilation). It is well known that clouds at the top of the boundary layer interact strongly with the subcloud layer, but many large scale models still treat dry and moist processes with separate schemes. This also applies to the current ECMWF model and it is seen as a major shortcoming. Some attempts have been made with the formulation of turbulent transport in terms of moist conserved variables combined with distribution functions for moisture to describe the subgrid fraction that contains condensate (LeTreut and Li, 1988; Smith, 1990; Brinkop and Roeckner, 1993; van Meijgaard and van Ulden, 1998). In practice, these schemes are difficult to couple with prognostic cloud variables and with mass flux convections schemes.

Tiedtke (1993) has followed an alternative approach. He implemented prognostic variables for condensate (water and ice) and cloud fraction with source and sink tendencies from all processes in the ECMWF model (e.g. production of clouds by detrainment from the updrafts in the mass flux convection scheme). This provides a clean coupling of the cloud scheme with the mass flux convection scheme (Tiedtke, 1989), but the coupling with the boundary layer is still rather artificial. Much more research is obviously needed in order to reduce the uncertainty in the parametrization, to unify the schemes and to improve the quality of cloud forecasts. The GEWEX Cloud System Study (GCSS) is an important initiative in this context, because it involves close co-operation between observational work, large eddy simulation (LES) and parametrization for larges scale models (Browning, 1993). Such co-operation is vital, because complicated parametrizations (e.g. the Tiedtke cloud scheme) use parameters that can only be derived with help of parametrization-dependent diagnostics on observations or LES data (e.g. the production of clouds by convective detrainment).