PART II ATTACHMENT 3

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A new surface/boundary layer fromulation at ECMWF

Introduction

This article describes the model changes introduced operationally on 4 August 1993 and results of forecast and data assimilation experiments with the new model version (Cycle 48). Many of these changes were introduced to cure problems that were identified by comparison with field data (see Betts et al., 1993 and Beljaars and Betts, 1992). The problems of the previous model version (Cycle 47) can be summarized as follows:

- the ground heat flux over land surfaces was too large by a factor of 2 to 3 and had a large phase error in the diurnal cycle;
- the diurnal cycle of the sensible and latent heat flux had a phase error of about 2 hours due to thermal inertia of the 7 cm surface soil layer in the model;
- the boundary layer depth was generally too small, indicating a lack of boundary layer entrainment;
- the boundary layer was too moist, even if the surface fluxes were correct. This was again due to lack of boundary layer entrainment;
- the evaporation from the surface was too large in wet conditions and too small in dry conditions;
- the soil moisture was excessively dominated by the climate layer;
- runoff tended to be a constant fraction of precipitation, resulting in relatively large runoff even when the soil was dry.

These model deficiencies inspired a set of changes in the boundary layer parametrization and in the land surface parametrization.

In addition, the parametrization of air-sea interaction has been modified. The principles of air-sea interaction have already been described by Miller et al. (1992), together with a description of the enhancement of evaporation at low wind speeds that was introduced in Summer 1990, with an empirical formula restricted to low wind speeds. The revised air-sea interaction is now part of the Monin-Obukhov formulation.

Model changes

Boundary layer above the surface layer

We have to distinguish two different regimes: (i) the stable and (ii) the unstable regime.

(i) The stable regime

Two different versions of the vertical diffusion scheme were used for the stable regime in Cycle 47 of the operational model. Below a generous estimate of the boundary layer height, the stability-dependent exchange coefficients were a function of the Richardson number (Louis, 1979, Louis et al., 1982). It was discovered that, when applied to the free atmosphere, the Richardson number formulation results in excessive mixing and is detrimental to the wind jet structures in the vicinity of the tropopause (revised in model version Cycle 29, January 1988). In Cycle 47 a MoninObukhov (M0) formulation (Beljaars and Holtslag, 1991) was used above the boundary layer height. An iterative procedure is used to calculate the Obukhov length given the Richardson number. The diffusion coefficients used by the model are much smaller in the free troposphere than below the boundary layer height.

The new model (Cycle 48) follows the same approach; the only difference is that the stability functions of the new MO formulation are slightly modified to get a better Richardson-dependence of the turbulent Prandtl number (ratio of turbulent diffusivities for momentum and heat) in the range of Ri from 0.2 to 1. Originally an attempt was made to use the MO scheme in the boundary layer as well (which makes the estimation of the boundary layer height obsolete), but this was abandoned because of a detrimental impact on the European objective scores from the reduced diffusivities in the stable case.

(ii) The unstable regime

The formulation of vertical diffusion in Cycle 47 (Louis, 1979, Louis et al., 1982) in the unstable regime is similar to the stable regime, with Richardson number dependent stability functions below the boundary layer height. The diffusion coefficients in the unstable regime are large leading to rapid dry adiabatic adjustment. This process is relatively insensitive to the formulation since mixed profiles of dry static energy are always produced, provided that the diffusion coefficients are sufficiently large.

It was shown with help of FIFE data that the lack of entrainment through the stable capping inversion results in mixed layers that are too shallow and too moist. An entrainment parametrization has been introduced in Cycle 48, by specifying the diffusion coefficient in the capping inversion such that the buoyancy flux in the entrainment layer becomes proportional to the surface buoyancy flux. The entrainment constant is 0.2. Since the heat flux becomes negative in the upper part of the mixed layer, the traditional local closure cannot be used any more (since the stable regime of the closure would be selected). A profile of diffusion coefficients is prescribed in the mixed layer as proposed by Troen and Mahrt (1986). Details of how this scheme performs in comparison with FIFE data are given by Beljaars and Betts (1992).

The surface layer

In Cycle 47, the transfer coefficients for heat. momentum and moisture were based on the Richardson number formulation, and a single roughness length was used for all the fluxes (Louis, 1979). Over sea there was a correction to the heat and moisture transfer, enhancing the transfer rates at low wind speeds.

In Cycle 48, the transfer coefficients which are used to parametrize the surface fluxes of momentum, heat and moisture consist of a neutral part determined by the logarithmic profile and separate roughness lengths for momentum and heat (the moisture roughness length is identical to that for heat), plus a stability correction. Over the ocean the roughness length for momentum has been modified at low wind speeds according to smooth surface scaling (see Miller et al. 1992). This concept has been applied for all wind speeds in the roughness lengths of heat and moisture. The high wind part of the heat and moisture transfer coefficients used is virtually independent of wind speed, in accordance with recent reviews of data (see de Cosmo, 1991, for a summary of the HEX0S experiment results).

The roughness lengths over land have been recomputed from vegetation and orographic where new empirical formulae have been used the orographic contribution (see Mason 1992) . Where an orographic contribution applies to the roughness length for momentum, the neutral transfer coefficients for heat and moisture are kept constant. This results in orders of magnitude reduction of the roughness lengths for heat and moisture, when compared to the momentum values.

The stability corrections applied to the neutral transfer coefficients are now expressed as a function of the Obukhov length instead of the Richardson number This allows for a consistent treatment of different roughness lengths for heat and momentum in combination with stability corrections.

The skin temperature

In order to have a faster response of the sensible and latent heat fluxes to the radiative forcing and to reduce the heat flux into the ground, the concept of a skin layer has been introduced. The skin layer has no heat capacity and adjusts its temperature instantaneously to the radiative forcing. The heat transfer to the underlying soil is parametrized with the help of an empirical conductivity, the value of which determines the amplitude or the diurnal cycle in the soil heat. flux. The skin temperature is calculated implicitly as part of the vertical diffusion scheme in order to reduce time truncation errors. The skin layer together with the reduced roughness length for heat is responsible for a weaker coupling between surface and atmosphere and for a much reduced coupling between surface and soil. The surface temperature is allowed to increase during daytime without increasing the temperature at 2m and without increasing the soil heat flux. Likewise, during night time, the skin layer cools radiatively, leading to lower minimum surface temperature. Although this is realistic, two-metre temperatures are currently adversely affected, giving rise to negative biases (see section on parallel run). The problem lies in the post processing of two-metre temperatures and is under investigation.

Land hydrology

The Cycle 47 surface scheme was based on the heat and water budget of two a active soil layers plus an additional surface layer underneath (Blondin, 1991, Viterbo and Illari,1993). The fluxes of water and energy between the layers are based on constant diffusion coefficients. The climate values, kept constant during the forecasts, were used as lower boundary conditions and updated at the beginning of every month. The "Mintz and Serafini" climate (Mintz and Walker, 1993) is used for soil moisture, while for surface temperature the RAND climatology is used (Brankovic and van Maanen, 1985).

From comparisons with FIFE data it was concluded that the land hydrology in Cycle 47 was inaccurate and dominated excessively by the climate fields. This was confirmed by single column simulations where the model's atmospheric forcing is replaced by observational data (see Fig. 1, described below).

The new scheme has 4 prognostic layers, to represent the diurnal to the annual time scales. The diffusivities and conductivities of soil moisture are non-linear functions of the soil moisture. This allows precipitation to penetrate fairly quickly into the soil and in dry conditions the upward diffusion of water becomes slower. The runoff in the new model is mainly due to gravitational drainage. Boundary conditions at the bottom are zero energy flux and free percolation. The new hydrology scheme has been extensively tested in one column mode with the help of long data sets (see Fig. 1 for FIFE). Results of these comparisons will be published elsewhere. The main conclusion is that the new scheme maintains evaporation in the drying season for a longer time and that it loses less water in runoff when the soil is dry. In general the new scheme tends less to extremes. It produces less evaporation in wet conditions and the soil dries out less quickly.

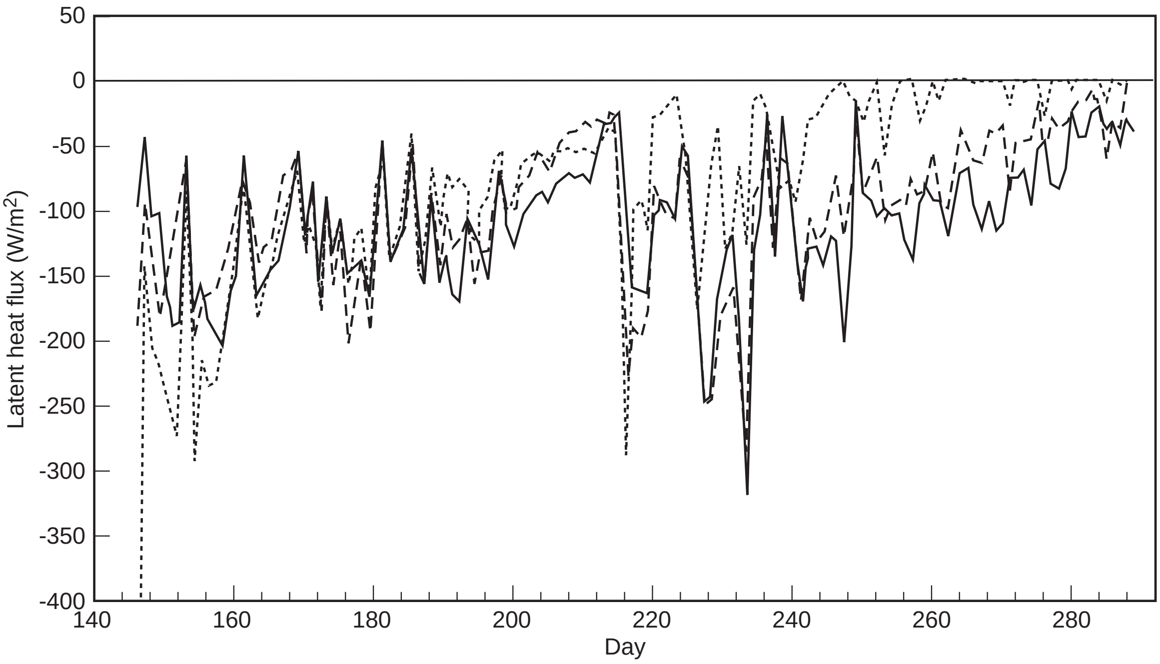

Fig. 1 Latent heat fluxes for FIFE; time in Julian days, from the end of May to mid -October. Solid line: Observations, Dotted line: One-column simulation using the old operational system, Dashed line: One column simulation with the new scheme.

Clouds

The entrainment in shallow cumulus clouds has been increased and the relative humidity criterion for inversion clouds has been modified. The effect of increased entrainment in shallow convection is to have a more rapid mixing of the updrafts with surrounding air, resulting in less deep penetration.

Long runs

Two types of long integrations wrercarried out with the new version of the model at T63 resolution:

- a multi-year run (4.5 years) to examine the long term stability of the model;
- 120 day summer and winter runs.

Multi-year integrations (T63L31)

From the multi-year integration, monthly averaged surface fields were computed (for 12 UTC) and used to make time series for different areas. Fig. 2 shows the time series for soil wetness averaged over Central Europe for the control and the NEW model. The first impression is that the NEW annual cycle is physically more realistic in that we see a decreasing amplitude of the annual cycle with increasing depth and that phase differences occur. The magnitude of soil moisture (in mm of water per 70 mm of soil) in the two model versions cannot be compared directly because it has to be interpreted in relation to the settings of field capacity and wilting point. (At field capacity the evaporation is not limited by soil moisture availability; at the wilting point, evaporation stops).

To understand the difference between control and NEW it should be realized that the soil moisture processes are quite different. In the control model, the top layer and the deep layer have roots (7+42 cm of soil); in NEW, layer 1, 2 and 3 have roots (7+21+72 cm of soil). The supply of water from the climate layer in the control run is therefore through diffusion when the difference in soil moisture between the climate layer and the deep layer is large. In NEW, the roots have direct access to water from a lm deep soil layer; the supply from layer 4 is through diffusion, but this is relatively slow.

120 day runs (T63L19)

Summer and winter 120 day integrations were done to study the model climate. Soil moisture and soil temperature were initialized with monthly averages from the multi-year runs. The zonal mean wind and moisture errors for the NH-winter run are shown in Fig. 3. The errors are generally reduced in the tropics: the NEW model reduces the easterly errors, the Hadley circulation is enhanced (reduced V and W errors, not shown) and the cold bias is reduced (not shown). Also the moisture bias is reduced considerably in the tropics. With respect to this it is interesting to note that the negative bias at

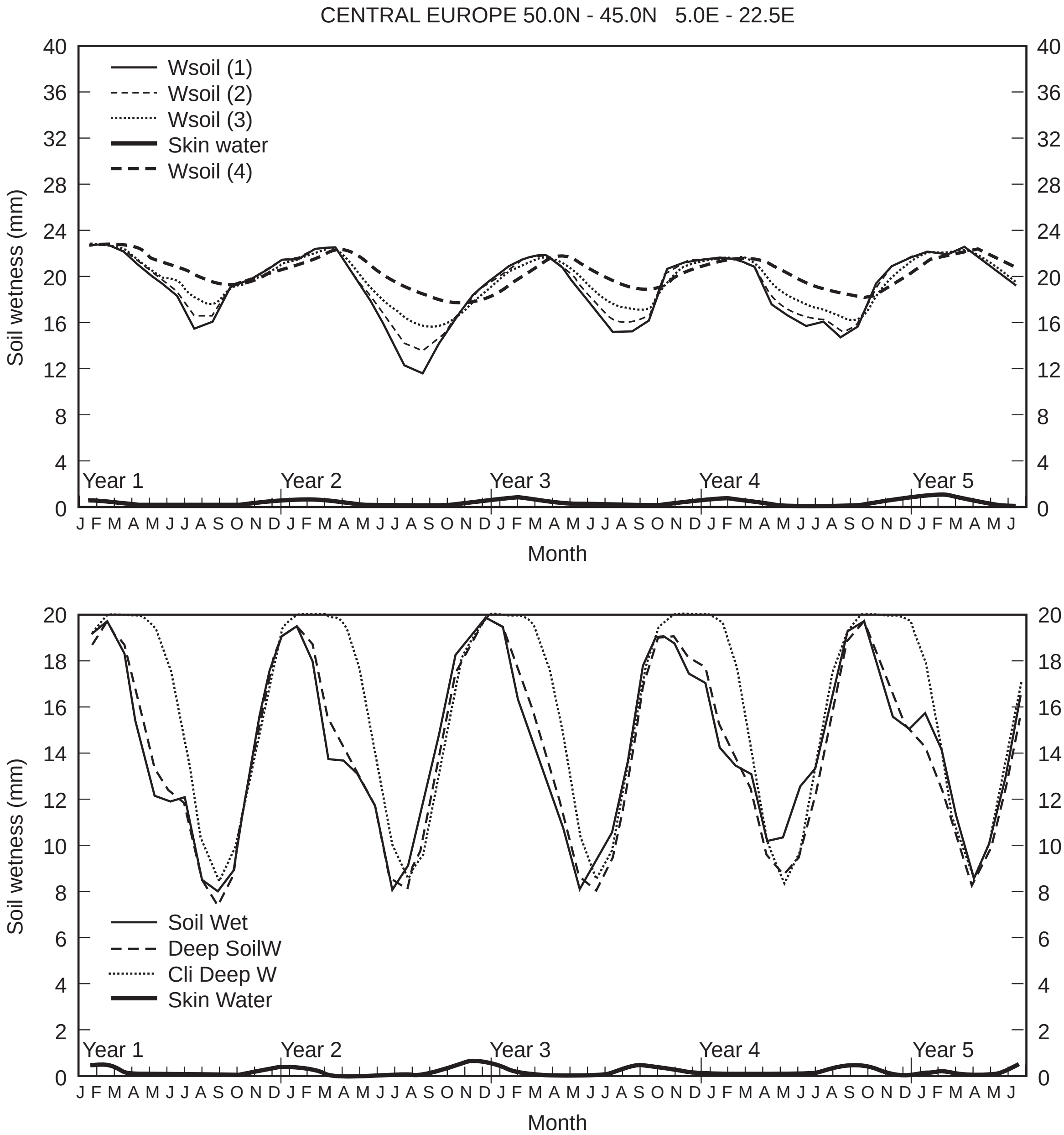

Fig. 2 Time series of monthly averages of soil wetness for Central Europe from multi-year runs with T63L31. The upper panel shows results for the NEW model, the lower panel for the control. The labels 1 to 4 in the upper panel correspond to increasingly deeper soil layers.

850 hPa as well as the positive near-surface bias between 20 and 50 degrees north are reduced. This can mainly be attributed to changes in shallow convection and to the boundary layer entrainment. The shallow convection change makes the convection less deep but enhances the mixing across the inversion and therefore moistens the levels around 850 hPa (Fig. 4a). Over land, the entrainment dries the boundary layer from above, but it competes with more moistening at the surface due to the new surface hydrology. The net effect is to reduce the moistening by vertical diffusion over land (Fig. 4b).

The effects on the tropical circulation are very similar in the NH summer integration. It is worth noting that about half of the impact (enhanced Hadley circulation, reduced easterly errors and increased tropospheric temperatures) is due to the boundary layer and air-sea interaction changes; the other half comes from the shallow convection change.

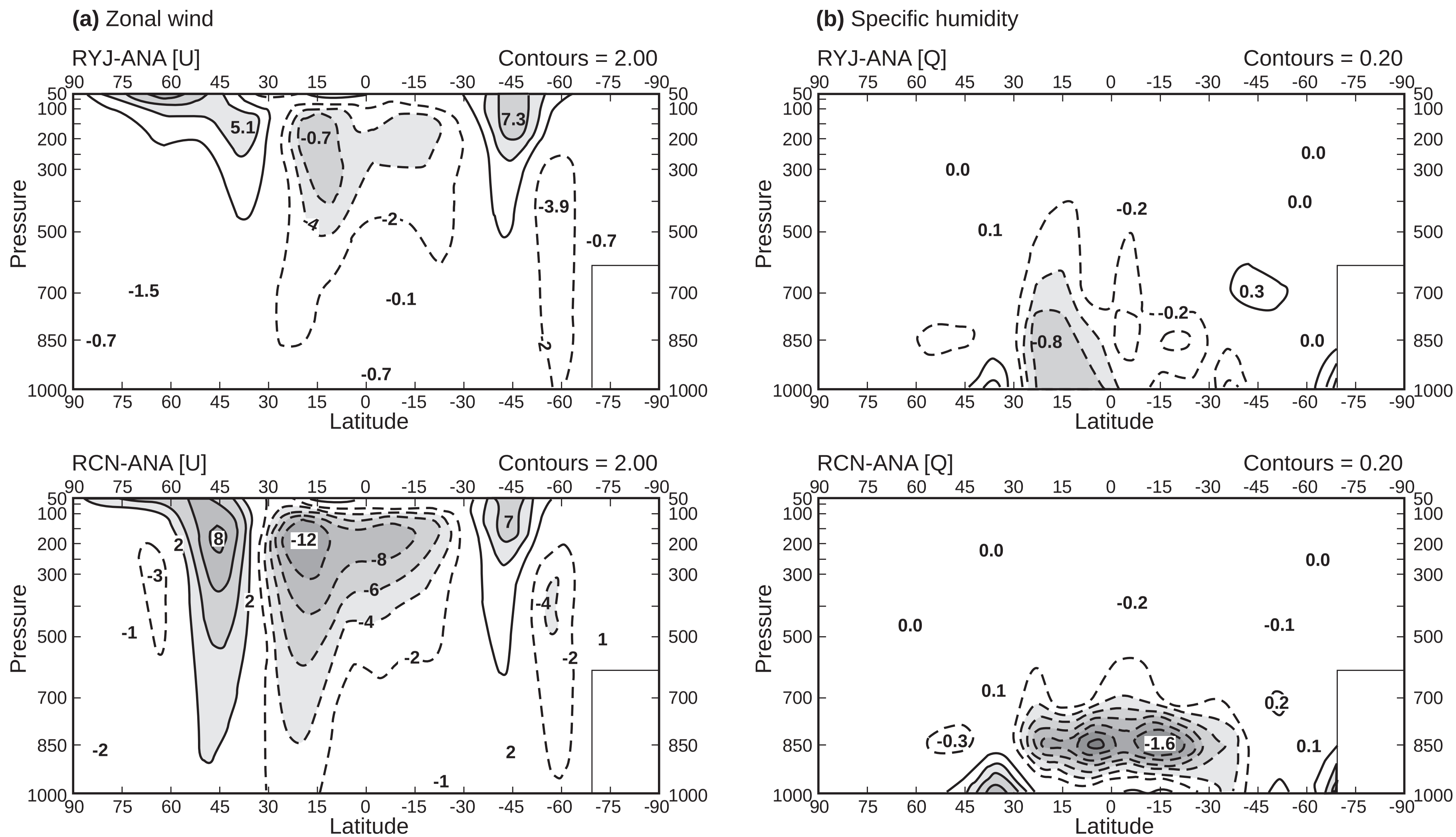

Fig. 3 Zonal mean errors averaged from day 31 to day 119 for the NH winter integrations (from 1-11-91) with NEW model (upper panels) and control (lower panels): a) zonal wind and b) specific humidity

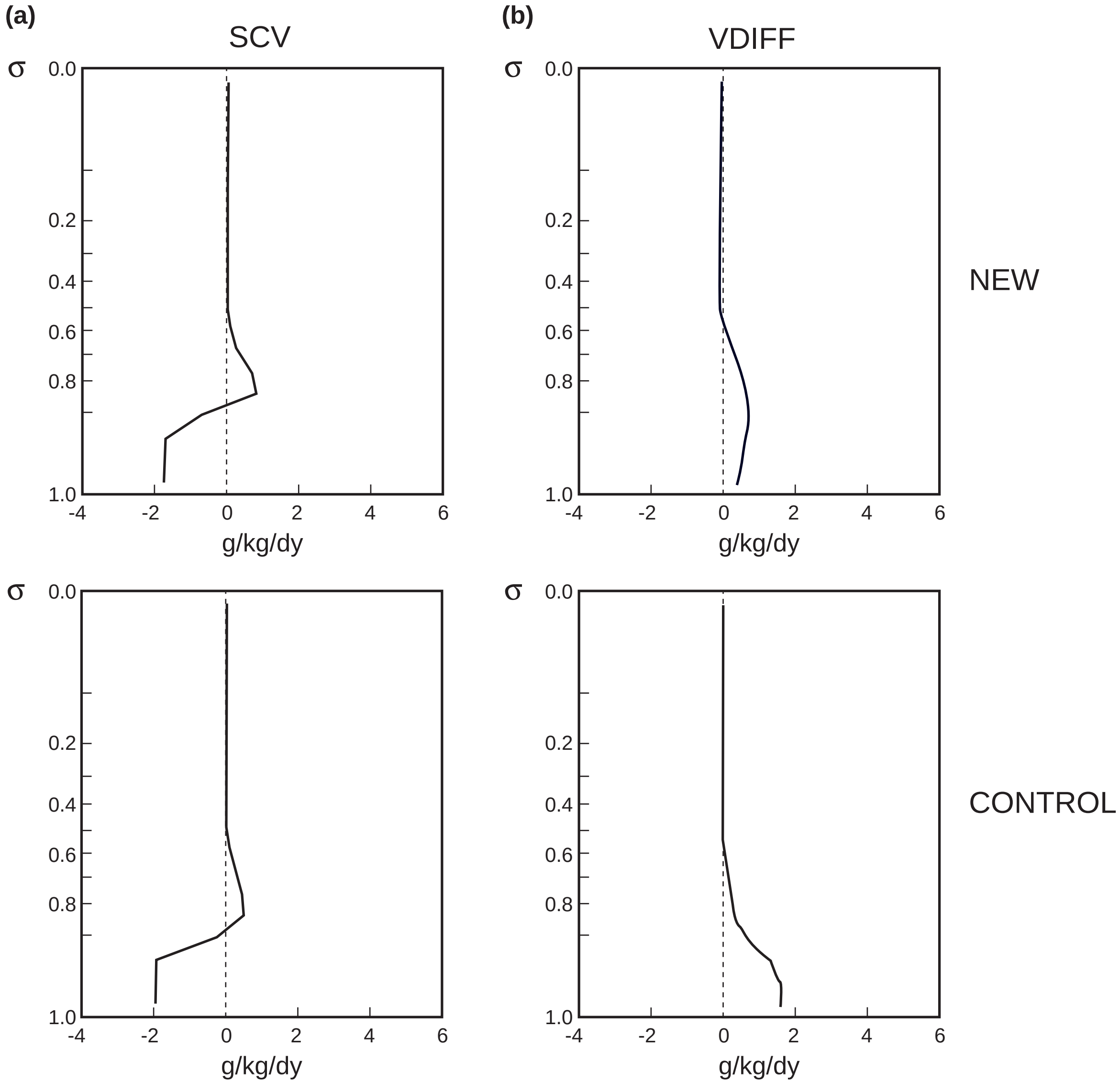

Fig. 4 a) Moisture tendency from shallow convection averaged for all sea points. b) Vertical diffusion tendency averaged for all land points. Mean results from a T63L19 NH summer from day 30 to day 119. Upper panels show results for the new model, lower panels for the control run. Units g/kg/day.

Forecast and assimilation experiments

A standard ensemble of 12 forecasts was run for the 15th of each month; later the ensemble was extended to 15 with 3 winter cases. Soil moisture and soil temperature were initialized from the monthly climate of the multi-year T63 runs. The mean impact on the scores is fairly small. Furthermore, the T213 tests confirm the earlier findings from the 120-day T63 runs: the Hadley circulation is enhanced and the systematic errors in boundary layer moisture content are reduced.

Ten days of data assimilation (with 10-day forecasts run from the last 5 days) were run for May 1993 with the NEW model. The fit to the data in this experiment is very similar to the operational suite, except for the relative humidity. The relative humidity in the NEW model boundary layer is much closer to the data. This is due to a better control of evaporation from the surface and to the introduction of boundary layer entrainment which tends to dry the boundary layer. The 500 and 1000 hPa European and Northern Hemispheric scores show a clear advantage of the NEW model over the Control run (not shown).

17 days of data assimilation and the corresponding forecasts covering the period of November 1992 were run at T106 with the NEW and Control model. The main signal is again from the boundary layer humidity. The impact of the changes on the mean scores is very small for this data assimilation experiment, although considerable day to day variability was found.

Parallel run

The NEW model was run in parallel (data assimilation and forecast) from 2 July 1993 until it entered into operation (4 August 1993). In this section we compare the Control and the NEW run.

The analysis fit to the data is very similar, except for the humidity, as before. The fit of boundary layer humidity to radiosonde data (not shown) is better over the Northern Hemisphere and worse over the Southern Hemisphere.

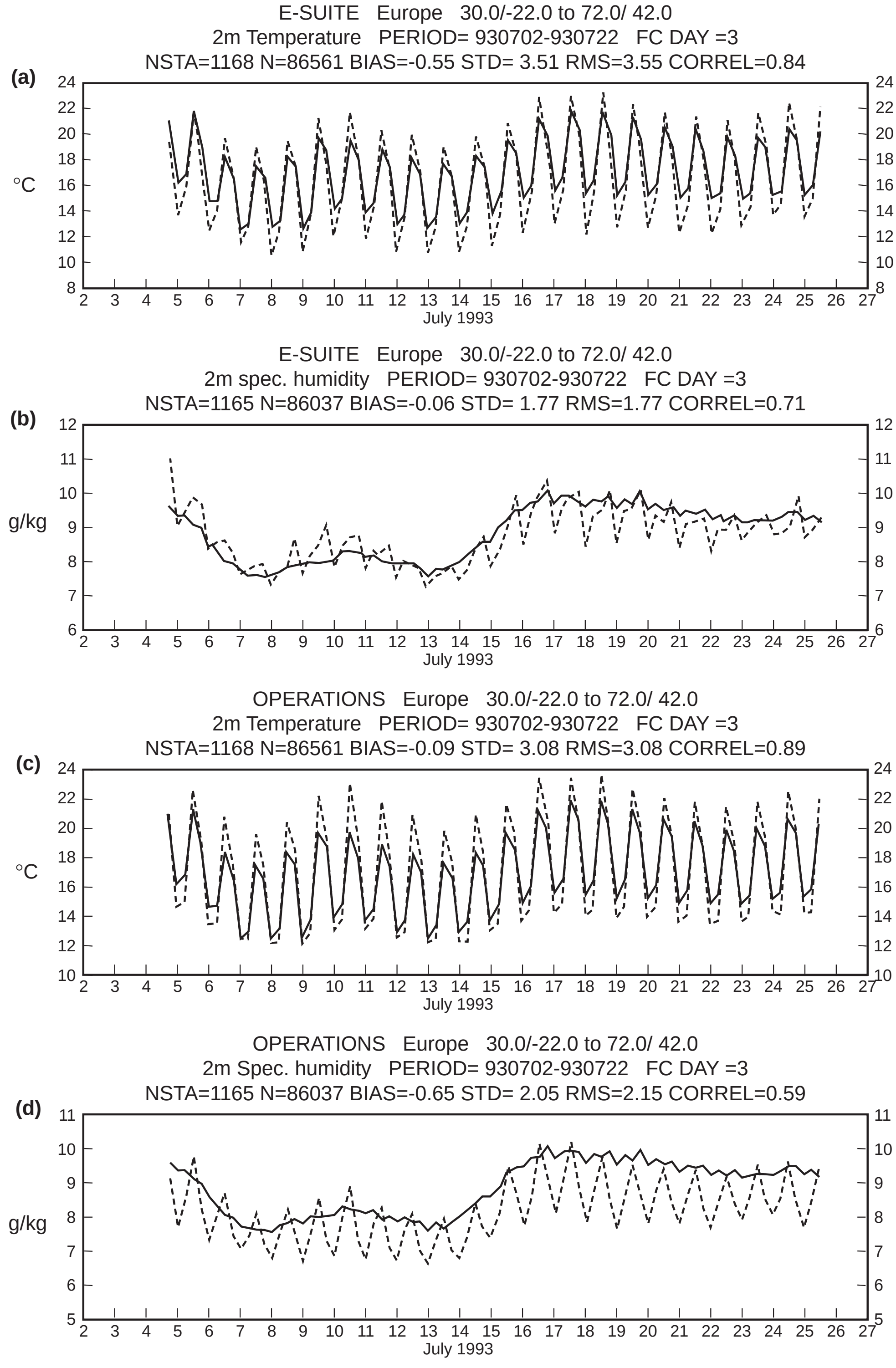

Fig. 5 Time series of temperature and specific humidity over Europe (54 to 72 hour forecasts, dashed lines) in comparison with SYNOP data (solid Line): a) Two metre temperature, NEW model; b) Two metre specific humidity, NEW model; c) Same as in a), for Control; d) Same as in b) for Control.

Comparison of near surface parameters with SYNOP data also shows an improved moisture structure in the boundary layer. The specific humidity -is always closer to the data in the NEW model and the amplitude of the diurnal cycle is reduced (see Fig. 5 for Europe). In the NEW model the moistening from the surface is partially compensated by drying from the boundary layer top. Also the reduced coupling between the atmosphere and surface makes the specific humidity drop less with the reduction in surface temperature during the night. The beneficial effect of Cycle 48 on near surface specific humidity is most pronounced over Southern Europe (not shown).

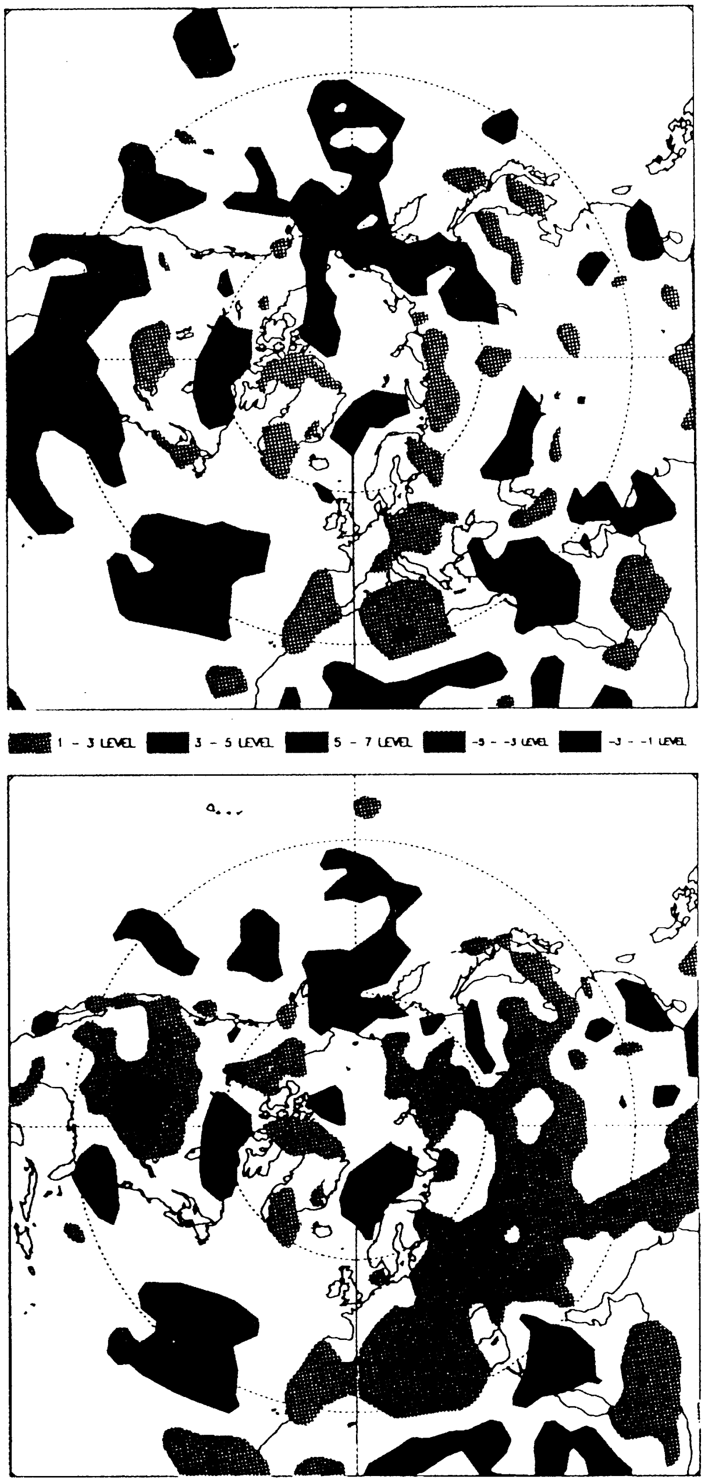

Fig. 6 Temperature error of day 5 forecasts averaged from 4 July to 3 August 1993: Top panel: New model parallel run; Bottom panel: Operations

The daytime two-metre temperature errors over Europe are reduced and they compare better with the SYNOP observations (Fig. 5). The night-time temperatures are also lower and in fact become too low. It should be realized, however, that the daytime near-surface temperatures are coupled to a deep atmospheric layer and that the night-time cooling is restricted to a very shallow boundary layer. The improved temperature structure of the atmosphere over continental areas is probably best illustrated by the 850 hPa temperature error map of the day 5 forecast with the control run and the NEW model, averaged over the entire parallel run, for a total of 31 forecasts (see Fig. 6; night-time plot is not shown, but very similar). That these temperature errors exist over deep layers becomes clear from the cross section along the latitude band 40°-50°N (Fig. 7).

Averages of the scores for Europe and Northern Hemisphere are shown in Fig. 8. Both regions show an improvement in the day 3 to day 4 range which is considered to be quite robust (the scatter from individual forecasts is very small). It is believed to be related to the reduction of the continental temperature bias.

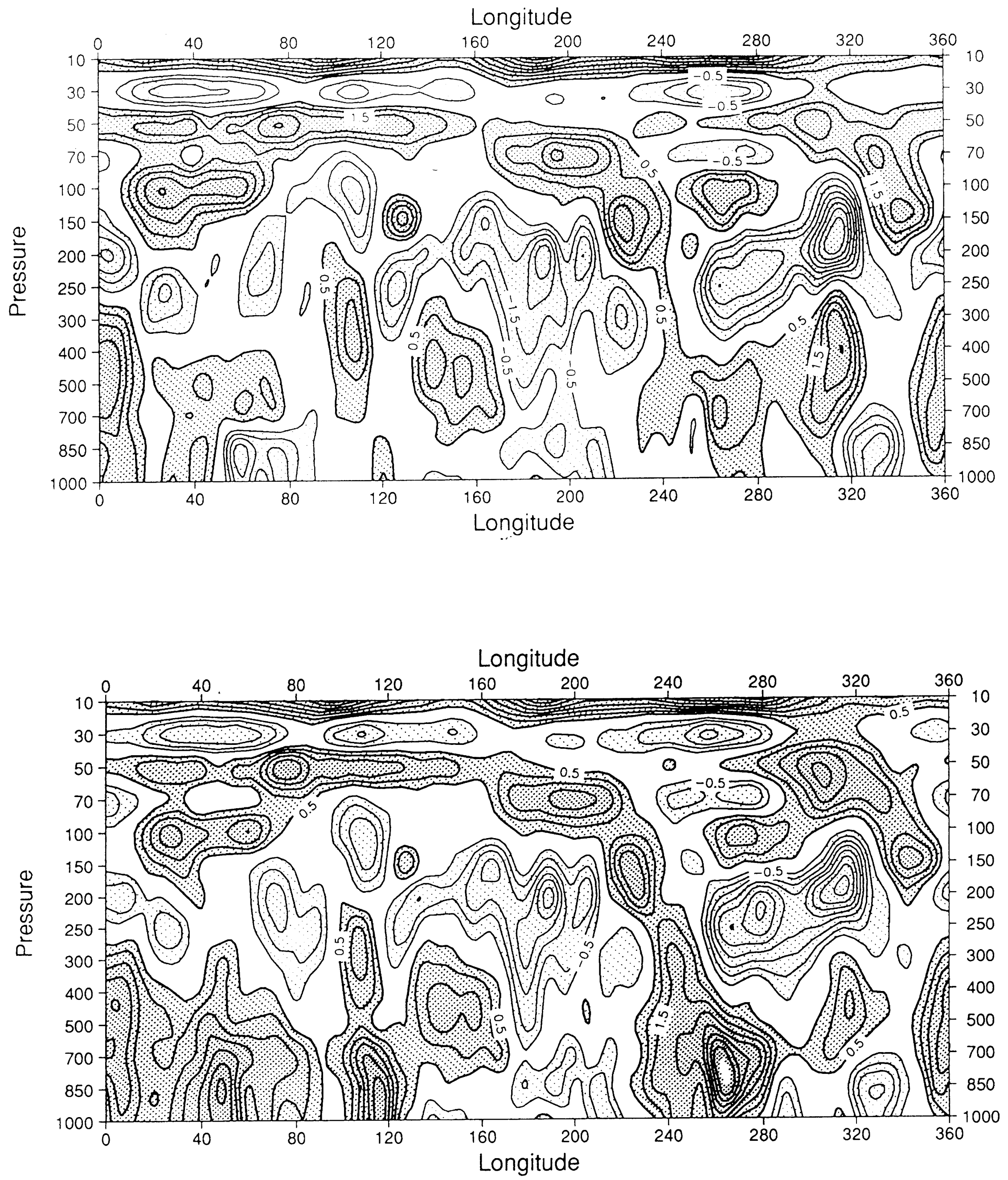

Fig. 7 Cross section of temperature error at latitude band from 40° to 50° North for NEW (top panel) and Control (bottom Panel)

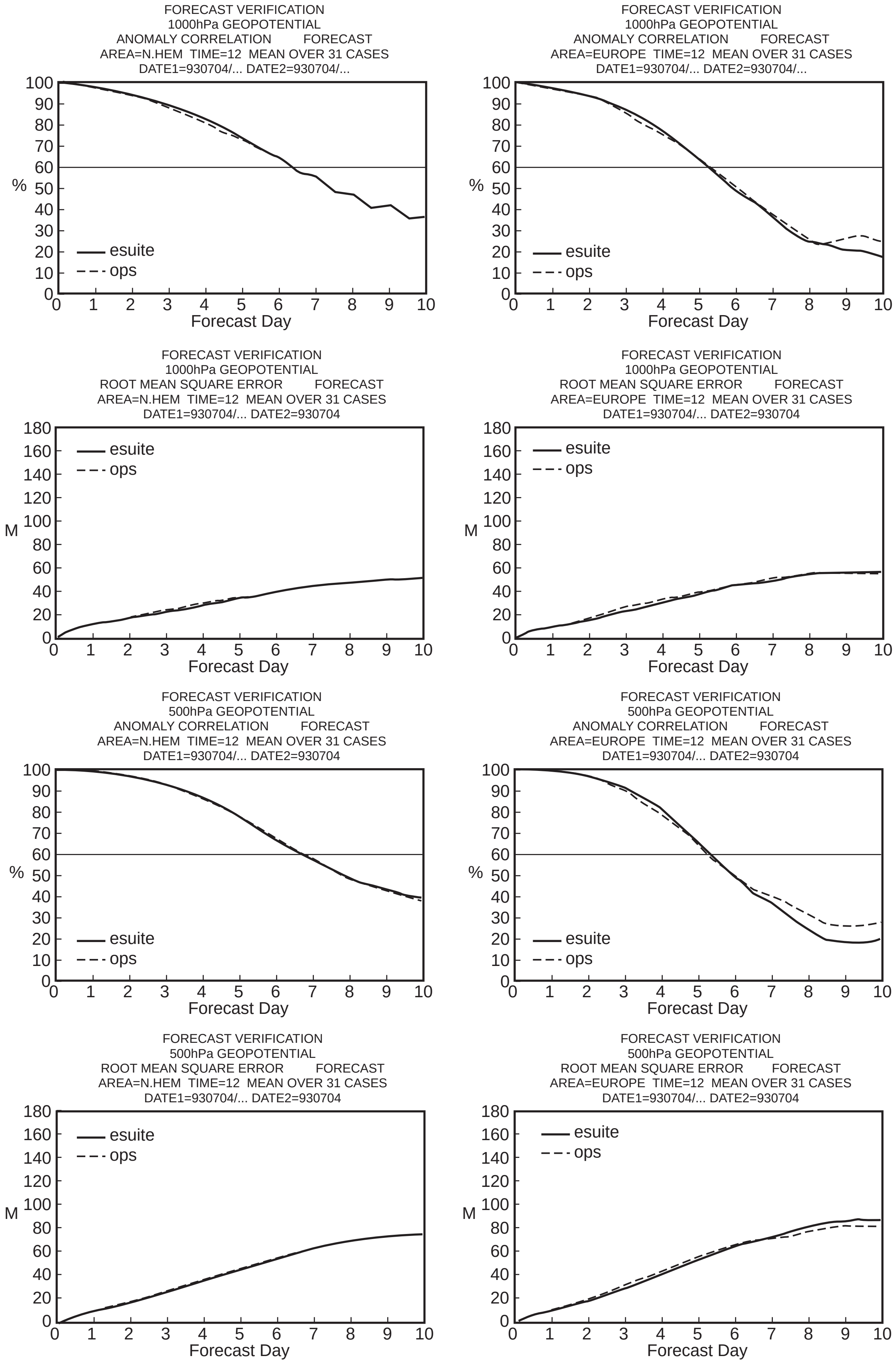

Fig. 8 Scores averaged over 31 forecasts from the parallel run with the NEW model in comparison with Control

REFERENCES

Beljaars, A.C.M., and A.K. Betts, 1992: Validation of the boundary layer representation in the ECMWF model. Proceedings of the 7-11 September 1992 Seminar on validation of Models over Europe, pp. 159 196, Vol. II. ECMWF, Reading.

Betts, A.K., J.H. Ball and A.C.M. Beljaars, 1993: Comparison between the land surface response of the European Centre Model and the FIFE-1987 data. *Quart. J. Roy. Meteor. Soc.,* **119,** 975-1002.

Beljaars, A.C.M. and A.A.M. Holtslag, 1991: On flux parametrization over land surfaces for atmospheric models. *J. Appl. Meteor.,* **30,** 327-341.

Blondin, C., 1991: Parametrization of land-surface processes in numerical weather prediction.

Land surface evaporation: measurement and parametrization, **T.J. Schmugge and J.-C. André,** Eds., Springer-Verlag, 31-54.

Brankovic, C, and J. van Maanen, 1985: The ECMWF climate system. ECMWF Tech. Memo. No. 109, 51pp. & figs.

de Cosmo, J., 1991: Air-sea exchange of momentum, heat and water vapor over whitecap sea states. *PhD thesis,* 212 pp. Dept. of Atmospheric Sciences, Univ. of Washington, Seattle.

Louis, J.-F., 1979: A parametric model of vertical eddy fluxes in the atmosphere. *Boundary-layer Meteorol.,* **17**, 187-202.

Louis, J.-F., M. Tiedtke and J.-F. Geleyn, 1982: A short history of the PBL parametrization at ECMWF. Proceedings, ECMWF workshop on planetary boundary layer parametrization, Reading, 25-27 Nov. 81, 59-80.

Mason, P., 1991: Boundary-layer parametrization in heterogeneous terrain. In Proceedings of the 16-18 September 1991 Workshop on Fines-scale modelling and the development of parametrization schemes, pp. 275-288. ECMWF, Reading.

Miller, M., A.C.M. Beljaars and T.N. Palmer, 1992: The sensivity of the ECMWF model to the parametrization of evaporation from tropical oceans., *J. Clim.,* **5,** 418-434.

Mintz, Y. and G.K. Walker, 1993: Global fields of soil moisture and land surface evapotranspiration derived from observed precipitation and surface air temperature. *J. of Appl. Meteoro.,* **32,** 1305-1334.

Troen, I. and L. Mahrt, 1986: A simple model of the atmospheric boundary layer; sensitivity to surface evaporation. *Bound.-Layer Meteor.,* **37,** 129-148.

Viterbo, P. and L. Illari, 1993. The impact of changes in the runoff formulation of a general circulation model on surface/near surface parameters. To appear in *J. of Hydrology.*