Tropical forecasting at ECMWF: The influence of physical parametrization on the mean structure of forecasts and analyses

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SUMMARY

The period 1984–1985 has seen a dramatic improvement in the quality of the ECMWF operational forecasts for tropical regions. This improvement is discussed in terms of the model changes during this period. Revisions include the introduction of a parametrization of shallow cumulus convection, modifications to the parametrization of deep cumulus convection, a new cloud scheme, and an increase in horizontal resolution with a change in spectral truncation from T63 to T106. Impact of the various model changes is assessed through a set of forecast experiments and through parallel 10-day forecasts over a 20-day period. The results show that the improvements in tropical forecasts are mainly through the reduction of systematic errors in response to a more realistic tropical diabatic forcing. The impact of these changes on the analyses is also assessed. This is shown to be substantial, particularly for the thermal state.

1. INTRODUCTION

Forecast errors in the tropics grow rapidly (Shukla 1981) and, as shown by Heckley (1985a) and Kanamitsu (1985) for the ECMWF model, tropical forecasts soon become dominated by systematic errors (defined here as errors associated with the steady component of the flow field). For operational ECMWF forecasts made during 1983/84, Heckley (1985a) estimated the wind forecasts for the tropical belt as skillful to about three days at 850 hPa and nearer ten days at 200 hPa (where ‘skill’ is defined to be the time for which the r.m.s. of the vector wind error is less than that of a corresponding persistence forecast). Kanamitsu (using the much tougher measure of a climatological forecast) found the 1983 forecasts to be skillful only to about one day at 850 hPa and 2–5 days at 200 hPa. For transient disturbances, however, Kanamitsu found predictive skill in many cases beyond three days—which is encouraging in view of the very large error growth in the steady flow. The performance of the recent ECMWF operational forecasting system in analysing and tracking easterly waves in the tropics has been evaluated by Reed et al. (1986), who also found useful predictive skill.

Although the usefulness of tropical forecasts has dramatically improved in the last few years, due to better observational data and improvements in data assimilation and forecast models (see WMO (1985), GARP special report No. 44), forecasts for the tropics remain relatively poor compared with those for the extratropics. This is despite errors in tropical forecasts being comparable with, or even less in absolute terms than, those of mid-latitude forecasts (Kanamitsu 1985). Climatological variances of fields in the tropics are so much smaller than those in the extratropics that tropical forecasts have to be considerably more accurate, in absolute terms, in order to exhibit any ‘skill’. Another problem that has to be contended with is the dominance of physical forcing over dynamical forcing. One would expect improvements in physical parametrization to have a substantial impact on the tropical simulation. Indeed, recent work (Kershaw 1985; Krishnamurti 1985) supports this view. Little is known, however, about the benefits of increased resolution on the skill of tropical forecasts.

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The object of this paper is to assess the impact of a combination of revised physics and increased horizontal resolution on systematic errors in ECMWF operational forecasts for the tropics. The formulation of the physical parametrization can have a strong influence not only on systematic errors but also on forecasts of individual tropical disturbances. Heckley et al. (1987) show an extreme example of this for forecasts of hurricane Elena.

The following changes were made to the ECMWF forecast model during May 1985:

(i) parametrization of shallow, non-precipitating, cumulus convection (appendix A);
(ii) revision of the parametrization of deep cumulus convection (appendix B);
(iii) new cloud cover scheme for calculation of radiation (Slingo 1987);
(iv) increase in horizontal resolution from a spectral truncation of T63 to T106 (associated with this was a change in horizontal diffusion and a revision of fixed surface fields, including an envelope orography based on use of 1, rather than \( \sqrt{2} \), times the standard deviation of the subgrid-scale orography) (Jarraud et al. 1986).

The new cloud cover scheme, for radiation calculations, is based on a diagnostic approach which allows for four cloud types: convective cloud, and three layer clouds — high, middle and low (Slingo 1987). Convective cloud and its associated cirrus clouds are derived from the penetrative convection scheme. Otherwise, high- and medium-level cloud is based on a function of relative humidity, whilst low-level cloud depends on relative humidity, vertical velocity and thermal stability. The scheme gives realistic cloud distributions and has substantially improved tropical cloudiness, particularly the representation of the diurnal cycle in cloud over the continents (Tiedtke and Slingo 1985). The impact of the scheme on medium range forecasts is slight but beneficial.

The increase in resolution and related boundary changes have a further impact on tropical forecast quality. However, as far as tropical systematic errors are concerned, the resolution change has less effect than the change to the physics.

It is useful to summarize forecast errors prior to these changes (Heckley 1985a; see also Heckley (1985b) for a more detailed review). The main errors were:

(i) a drying of the tropical atmosphere;
(ii) a cooling of the tropical troposphere and equatorial stratosphere and a warming immediately below the tropopause;
(iii) a weakening of the Hadley circulation;
(iv) a weakening of the subtropical anticyclones;
(v) a poleward and upward displacement of the subtropical jets.

Other model changes have occurred since the tropical systematic errors were documented by Heckley (1985a), but these did not appear to affect significantly the main characteristics of the errors. The introduction of the diurnal cycle in May 1984 improved the low-level thermodynamic structure, particularly over tropical continents (Heckley 1985b), and extensive revision of the analysis scheme (Shaw et al. 1987), introduced in late May 1984, improved the definition of ITCZs in the early stages of the forecasts. A correction to the moisture-dependence of specific heats, introduced some months earlier, also had a significant effect on temperature errors near the tropical tropopause. Other than the May 1985 changes listed above the only other major change (in December 1984) involved the treatment of long-wave radiation, which was revised to incorporate the absorber gases directly, through the technique of exponential sum fitting (Wiscombe and Evans 1977). This reduced the stratospheric cooling, particularly at the top model levels; there was also a reduction in the warming on and immediately below the equatorial tropopause and a reduction in tropospheric cooling (Ritter 1984). There was, however,
little effect on the tropical flow. Further details on this change are given in Slingo et al. (1988). Although these changes to the analysis/forecasting system resulted in beneficial reductions in the forecast errors, systematic errors prior to May 1985 remained large and still very much as characterized by Heckley (1985a).

A dramatic change in systematic errors followed the changes introduced in May 1985. These changes and their impact on tropical forecasts are the main subject of this paper.

2. IMPROVEMENT IN FORECAST SKILL 1984–1985

As a measure of the increase in predictive skill between these two years, two periods are considered: June through November 1984 and June through November 1985. The first period is after the introduction of the diurnal cycle but before the revised treatment of long-wave radiation. The second period is after both the revised treatment of long-wave radiation and the May 1985 changes.

Each period contains six months of daily forecasts. These forecasts were verified against the operational analyses over the tropical belt (20°N–20°S). The mean r.m.s. error of vector wind, as a mean over the tropical belt and as a mean of the six-month period, is shown in Fig. 1(a) for the 850 hPa level, for the 1984 and 1985 model forecasts. Also shown are the corresponding persistence forecast errors, and estimates of the error of climatological forecasts. The error of a climatological forecast can be estimated as $1/\sqrt{2}$ of the limit of the persistence error (Thompson 1961). In the example shown the climatological error has been estimated as $1/\sqrt{2}$ of the persistence forecast error at day 10. This will be an underestimate as the persistence error is still continuing to grow at that time. There are two estimates shown, one from the 1984 persistence forecasts and one from 1985.

Examining the persistence error first, it is apparent that this has increased between 1984 and 1985. This is as one would expect from increased resolution. Given two forecasts which are essentially similar but for their resolution, that of higher resolution will have larger r.m.s. errors because such scores are sensitive to phase errors associated with small-scale features. The 24-hour model forecasts are also slightly worse in 1985, but not so much as one might expect in the light of the changes in the persistence forecast errors. Beyond 48 hours the model forecasts are improved dramatically. In terms of r.m.s. error this is a substantial reduction. The 1984 forecasts at 850 hPa were better than persistence up to about three and a half days, but in 1985 equality was reached after about eight and a half days. However, if one uses the climatological forecast as a measure of skill, the improvements are more modest, which is more in accord with subjective assessment of forecast quality.

The scores for the 200 hPa level are shown in Fig. 1(b). The same general comments apply as for 850 hPa, but now one sees a large increase in forecast skill even when measured against the estimated climatological forecast error, from just under three days to about four and a half days.

Although these figures show a dramatic reduction in r.m.s. forecast error, and the importance of this should not be underestimated, much of this improvement, particularly at low levels, occurs when the model has lost most of its forecast skill. The error growth during the first few days is very fast. In order to improve these forecasts further it will be necessary not only to reduce the asymptotic limit of the errors, i.e. improve the model climate, but some way must also be found to reduce or retard the initial error growth.
rms vector wind forecast error
June to November
Tropical belt – 850 mb

Figure 1. Root mean square vector wind error as a mean over the tropical belt (20N–20S) and over the period June through November, as a function of forecast length. Curves shown are persistence forecast errors and model forecast errors for 1984 and 1985. Also shown is the error of climatological forecasts (estimated as $1/\sqrt{2}$ of the persistence forecast error at day 10). (a) 850 hPa; (b) 200 hPa.

3. FORECAST EXPERIMENTS

The impact of the parametrization changes was first studied in a set of 13 pairs of T63 forecasts and 8 pairs of T106 forecasts, each pair consisting of one forecast with the original model and one with the modified parametrization. The initial data are spread evenly over the seasons between 1983 and 1985, plus a case during FGGE (11 June 1979). The FGGE case is within the period of the monsoon onset and was chosen to study the impact of the individual parametrization changes. Results of these experiments are discussed in detail by Tiedtke and Slingo (1985) and by Slingo et al. (1988).

The impact of the complete set of changes (parametrization changes and changes associated with the increase in resolution) was then examined for a parallel run of the old and new analysis/forecasting systems for a period of twenty days beginning 11 April 1985. On each of the twenty days, a ten-day forecast was carried out from the 12 GMT analyses using both forecast models. The parallel run confirms the earlier findings and enables us to determine the systematic impact on the tropical flow by forming ensemble means. Moreover, it shows clearly the impact of the model changes on the analyses
through the use of the new model in the data assimilation. Also it is a more rigorous test as the impact of the model on the analyses can be expected to feed back on to subsequent forecasts. Ensemble means were made of the forecasts and analyses for both forecast systems. The ensemble forecasts from each model were compared against the corresponding ensemble analyses (with the caveat that only analyses from the new system were available after 31 April 1985).

Differences may occur between the analyses because the forecast model is used in a four-dimensional assimilation to provide the first-guess field for the analysis. Bengtsson et al. (1982) give a review of the ECMWF analysis system. Identical data were presented to the two analyses but it cannot be guaranteed that identical data were accepted, since short-range forecasts are used to decide on the acceptance or rejection of observations. With these provisions in mind the data set allows a 'clean' evaluation of the impact of the model changes both on the analyses and the forecasts.

In the following we refer to the 'old model' and to the 'new model' as the model version without and with the changes, respectively.

4. FORECAST IMPACT

The impact of the model changes on the tropical forecasts may be summarized as follows:

(1) The tropical diabatic heating is increased through an intensified hydrological cycle.
(2) The model's tendency to cool the tropical and subtropical troposphere is removed.
(3) The forecasts of the tropical flow are improved beyond day 2, in particular showing stronger subtropical anticyclones and stronger trade winds.

(a) The trade wind boundary layer

The trade winds play a central role in the energetics of the atmosphere as they accumulate latent heat from the ocean and transport it into the tropics (Riehl and Malkus 1957, 1979). Shallow convection is an important part of this process. It ensures that moisture is vertically transported from close to the surface into the cloud and inversion layers above, effectively increasing the downstream moisture transport.

The influence of shallow convection on the moisture field is clearly seen in Fig. 2, which shows a typical cross-section over the North Atlantic of relative humidity for a case during FGGE (11 June 1979). Without shallow convection the model produces a boundary layer which is almost saturated throughout, with little evidence of a moist cumulus cloud layer above. When the effects of shallow convection are parametrized in the model, water vapour is accumulated in the cloud layer at the top of the boundary layer. This occurs through the additional convective fluxes of moisture out of the boundary layer; these counterbalance the drying effect from the large-scale downward motion and thus maintain a moist cloud layer. The overall effect is to deepen (by 50–100 hPa) the whole moist layer while drying the lower part of the boundary layer. These effects of shallow convection are confirmed by the findings of the separate parametrization study, in which shallow convection was investigated using a special data set from a period of undisturbed trades during ATEX (see appendix A, Fig. A1).

(b) Moisture fluxes

As a result of the drying of the boundary layer more moisture is supplied from the subtropical oceans into the atmosphere through turbulent fluxes. This can be seen from evaporation fields for the ensemble of 5-day parallel run forecasts (mean evaporation
rates between 96 h and 120 h) shown in Fig. 3, (a) for the old model and (b) for the new model, together with climate estimates over the tropical and subtropical oceans for April, (c) (Esbensen and Kushnir 1981). The new model produces much higher values over the subtropical oceans; in some regions the evaporation has been almost doubled from typically 3 mm to over 5 mm a day. The values are now generally much closer to the climate estimates than with the old model. Thus the insufficient moisture supply from the subtropical oceans which was typical of the previous model (Tiedtke 1982) has largely been corrected by the introduction of the effects of shallow cumulus convection.

(c) Precipitation

The increase in moisture supply over the subtropical oceans (where there is little precipitation) has a direct effect on the moisture flux convergence into the tropical belt and thereby on the occurrence of deep convection in the tropics. The effect on the moisture convergence has not been diagnosed due to absence of fields archived in the coordinate system of the model from the parallel run, but the effect on cumulus convection
Figure 3. Accumulated evaporation 96–120 hours for ensemble of parallel run forecasts. (a) Old model; (b) new model; (c) climate estimates over tropical oceans, after Esbensen and Kushnir (1981). Contour interval: 1 mm d$^{-1}$. Values greater than 4 mm d$^{-1}$ are shaded.

is clearly evident. As the occurrence of penetrative cumulus convection depends strongly on moisture convergence, an increase in moisture convergence as a result of greater moisture supply from the subtropical oceans should directly produce an increase in tropical precipitation. This is confirmed, as can be seen from Fig. 4 which compares the precipitation of the old and new models accumulated from 96 to 120 hours for the ensemble of forecasts, together with the climatological estimate (Jaeger 1976). The most striking feature is the enhanced ITCZ in the new model. When compared with the climatological precipitation for April the rainfall pattern is generally now more reasonable. Details in this pattern also verify well. Figure 5 shows the fractional cloud cover for April 1985 produced by the Climate Analysis Center in Washington (NOAA 1985). This field shows a clear double structure in the ITCZ in the eastern equatorial Pacific. This structure is well captured by the new model (Fig. 4(b)).

In order to clarify which parametrization change contributes most to this impact, a series of test forecasts was made with various model formulations for one of the cases (FGGE, 11 June 1979): i.e. (a) for the old model; (b) for an intermediate version which contains only the change to the penetrative convection scheme; and (c) for the new
Figure 4. Accumulated precipitation 96–120 hours (mm d\(^{-1}\)) for ensemble of parallel run forecasts. (a) Old model; (b) new model; (c) April climatology (Jaeger 1976). Contour intervals: 1 mm d\(^{-1}\), 2 mm d\(^{-1}\) as light solid contours, solid contours at intervals of 4 mm d\(^{-1}\), starting at 4 mm d\(^{-1}\). Values greater than 8 mm d\(^{-1}\) are shaded.

Figure 5. Fractional cloud cover (235 K). April 1985. Mean fraction of 2.5\(^\circ\)×2.5\(^\circ\) areas covered by pixels with equivalent black-body brightness temperatures ≤235 K. Solid contours are at intervals of 0.05 with dashed contours at 0.01 below 0.05. This parameter is highly correlated with convective rainfall amounts in tropical and summer hemisphere continental regions. (Reproduced from April 1985 monthly climatic summary, published by CAC, NMC, Washington.)
model. The horizontal resolution was the same in all experiments (T63). The precipitation rates averaged over the whole 10-day forecast period are presented in Fig. 6 together with Jaeger's (1976) June climatology. The comparison shows that it is the introduction of shallow convection which has the largest effect. When only penetrative cumulus convection is modified, there is an overall increase of precipitation in the tropics (compare Fig. 6(b) with Fig. 6(a)) but the precipitation becomes too uniform and far too widespread (over the Pacific and Atlantic for example) with little indication of an ITCZ downstream of the trades. However, when shallow convection is included (Fig. 6(c)) precipitation becomes confined to smaller areas with much increased maxima indicating a more organized tropical circulation and a more realistic ITCZ. The introduction of the new cloud scheme has a much smaller effect on the precipitation than the convection changes, as is confirmed in additional experiments, not shown here, where only the cloud scheme is modified.
Figure 7. Global mean hydrological budget as a function of forecast length. Solid line—precipitation; dashed—evaporation. Units mm d\(^{-1}\). (Ensembles of parallel runs.)

(d) Global hydrological and energy balance

The more intense hydrological cycle is also evident in the global values of precipitation and evaporation calculated for the old and new models as a function of forecast length (Fig. 7). Actual values of global precipitation (even for the annual global mean) are difficult to obtain; the stippled area in the figure represents the mean ± 1 standard deviation of a number of (not all independent) estimates of this quantity, as summarized by Hoyt (1976).

A global balance between the net radiative cooling and the net heating due to convection, large-scale condensation and sensible heat flux is reached more quickly with the new model and is maintained at a higher level throughout the whole forecast period. The new model also differs from the old in that the precipitation increases sharply and then decreases sharply towards equilibrium levels. The effect is much less obvious with the old model. This 'overshoot' is exaggerated in the new model by the reduced diffusion applied to the divergence field.
Atmospheric Energy Budget.

Figure 8. Global atmospheric energy budget as a function of forecast length. (a) Old model; (b) new model. ‘Total input’ is convective plus condensation plus sensible heat. (Ensembles of parallel runs.)

The global atmospheric energy budget is illustrated as a function of forecast length in Fig. 8 for (a) the old model and (b) the new model. Large-scale condensation (which is mainly confined to mid-latitudes) is largely unaffected by the model changes. The new model shows a reduced value of sensible heat flux; this together with the increased evaporation affects the partitioning between sensible and latent heat flux at the surface (the Bowen ratio). As expected, the largest change is the increase in latent heating from convection by approximately 20 W m⁻². In the old model, the latent heating was too weak, the model cooled and tended towards an equilibrium state in which the radiative cooling was reduced. The new model starts out with higher radiative values, indicating a slightly warmer initial state, and tends towards a balance with convective heating decreasing. The energy balance shows the old model to be cooling and the new model to be warming during forecasts.

(e) Thermal structure

The zonal mean temperature errors for the ensemble forecasts at day 5 are illustrated in Fig. 9, for (a) the old model and (b) the new model. For the old model the errors are much as described by Heckley (1985a), except that the erroneous stratospheric cooling has been eliminated by the revised radiation scheme. The errors are largely characterized by a cooling in the mid and lower tropical troposphere of up to 2.5 K (0.5 K d⁻¹) and a warming on and immediately below the tropical tropopause. On the whole, there is a net cooling. For the new model, the temperature errors are much reduced with smallest errors at around 700 hPa and a tendency to cool the layer below and to warm the layer above.
Figure 9. Forecast error at day five of the zonal mean temperature. Contour interval 0.5 K. Zero line is drawn heavy; positive—solid, negative—dashed. Regions of error greater than ±2 K are shaded. (Ensembles of parallel runs.) (a) Old model; (b) new model.

Figure 10. As Fig. 9 but for zonal mean specific humidity. Contour interval 0.5 g kg⁻¹. Zero line is drawn heavy, positive—solid, negative—dashed.
The study of the hydrological cycle has already shown that both models have a tendency to dry the whole atmosphere. From Fig. 10 it is evident that this occurs predominantly in the mid and lower tropical and subtropical troposphere. It also shows that the drying is more pronounced with the new model.

Changes in temperature and moisture strongly affect the thermal stability. The vertical profile of $\theta_e$ (not shown) shows that the new model is overall more stable (less conditionally unstable) between 1000 and 850 hPa, less stable between 850 and 700 hPa, and more stable above.

(f) Mean flow

In section 2 it was shown that the new model has substantially more skill in the tropics when judged by the r.m.s. vector wind error. All experiments performed so far indicate that this increase in skill arises in the main from improvements in the very-large-scale flow rather than from scales associated with transient eddies (e.g. tropical cyclones), although there is some indication that the diabatic forcing with the new physics produces more realistic development. The revised physics led for example to a better forecast of the monsoon onset over India during FGGE (Slingo et al. 1988).

Improvement on the largest scales is already noticeable after day 2 and fully established at around day 4. We study here the 5-day forecasts. Fig. 11 shows the 850 hPa vector wind field for the old model, (a) as analysed, and (b) for the ensemble mean 120-hour forecast. The corresponding fields for the new model are shown in Fig. 12. The two analyses are similar, indicating that the model changes have relatively little effect on the wind field through data assimilation; the 120-hour forecasts, however, show large differences in the tropics. The new model retains stronger trade winds over almost all areas of the subtropical oceans. It is interesting to note from the surface pressure field (not shown) that the equatorial trough is deeper, and somewhat broader in the new model. In addition, the subtropical highs are also slightly intensified. Thus the meridional pressure gradient between the deep- and subtropics is increased. The circulation associated with the subtropical highs is enhanced and accounts for the considerable reduction of the low-level wind errors in the new model. This can be seen from Fig. 13 which shows the 850 hPa streamfunction errors for the old and new models at day 5. The weakening of the subtropical anticyclones in the old model appears as spurious cyclonic circulations in the forecast error field over the Atlantic and Pacific in both hemispheres. Except for some areas over the SE and NW Pacific these errors have largely disappeared in the new model.

Figure 11. 850 hPa vector wind field, old model (ensemble of parallel runs). (a) Analysis; (b) 120-hour forecast. Contour interval: 5 m s$^{-1}$.
Figure 12. As Fig. 11 but for new model.

Figure 13. 850 hPa streamfunction error at 120 hours for ensemble of parallel run forecasts. (a) Old model; (b) new model. Units: $10^6$ m$^2$s$^{-1}$. Contour interval: $2 \times 10^6$ m$^2$s$^{-1}$. positive—solid, negative—dashed; zero contour absent.

A reduction in streamfunction errors also occurs at 200 hPa (Fig. 14). This improvement is most apparent over the Atlantic and African regions. Problems remain over the Indian Ocean and in the region of the South Pacific convergence zone. Figures 13 and 14 illustrate the baroclinic nature of the tropical forecast errors and their large meridional scale, as discussed by Heckley (1985a). Improvements are also reflected in the zonally averaged zonal flow which shows less spurious acceleration of westerly flow in the tropical lower troposphere and less deceleration of the westerly flow in the upper troposphere.

The meridional circulation in the analyses and five-day forecasts is illustrated in Fig. 15, which shows the zonal mean meridional wind. The analyses with the old and new models are fairly similar. For the forecasts, the old model shows a weakening of the lower branch of the Hadley circulation; the new model is better in this respect. Both sets of forecasts differ from the analyses in the upper troposphere. The analyses show a well-
Figure 14. As Fig. 13 but at 200 hPa. Contour interval: $4 \times 10^4 \text{m}^2 \text{s}^{-1}$.

Figure 15. Zonal mean meridional wind. Contour interval 0.5 m s$^{-1}$. Zero line is drawn heavy solid, positive—solid, negative—dashed. Regions of wind greater than $\pm 1$ m s$^{-1}$ are shaded. (Ensemble of parallel runs.) (a) Analysis, old model; (b) five-day forecast, old model; (c) analysis, new model; (d) five-day forecast, new model.
defined return flow just below the tropical tropopause. The forecasts show a weaker flow, which, from mass conservation considerations, must now occur over much of the mid and upper troposphere. One should be cautious in describing such differences with respect to the analyses as errors as such details of the atmospheric flow in the tropics are not well defined by observations, particularly in respect of the vertical structure, and may therefore be an artifact of the analysis system.

The changes to the mean flow are largely the result of the changes to cumulus parametrization. For the eight cases for which both T63 and T106 forecasts were available with both old and new parametrizations, the increase in the horizontal resolution is seen to have a smaller effect on the ensemble-mean r.m.s. wind errors than the parametrization changes. This can be seen in Fig. 16. The large improvements obtained with the revised physics occur mainly on the largest scales (i.e. zonal mean and wavenumbers 1–3) and outweigh by far any improvements in the transient eddies. But, there are indications that the transient flow is better simulated with the higher resolution model, producing improved forecasts in a number of cases.
5. IMPACT ON ANALYSES

(a) Thermal structure

The significant differences are by and large confined to the oceanic areas. At very low levels, 1000 hPa, the new analyses are slightly colder by generally less than 1 K over the oceans in the deep tropics. At 850 hPa (Fig. 17(a)) the new analyses are slightly warmer over the oceans in the deep tropics (by up to 1 K) and cooler over the subtropical oceans (by 1–2 K). Figure 17(b) shows the difference between the 120-hour forecasts at 850 hPa; the pattern is very similar to the analysis differences shown in Fig. 17(a), but the forecast differences also indicate a more general warming in the new model. At 700 hPa (Fig. 17(c)) the new analyses are warmer throughout the tropics, over the oceans, by up to 3 K. Again, the pattern is very similar to the five-day forecast differences (Fig. 17d).

Figure 17. Temperature difference: new minus old model. Ensemble forecasts and analyses. Contour interval: 1 K. (a) Analyses at 850 hPa, differences < -1 K shaded; (b) 120-hour forecasts at 850 hPa, differences <0 K shaded; (c) analyses at 700 hPa, differences >2 K shaded; (d) 120-hour forecasts at 700 hPa, differences >4 K shaded.
The analysis differences, for the most part, reflect the differences in the forecasts from the two systems. In particular, the cooling at 850 hPa over the subtropical oceans is consistent with the presence of a cloud layer maintained below subsidence inversions in the new model and the warmer tropical troposphere is in agreement with the increased convective activity.

As for the temperature field, the analysis differences in the moisture field generally reflect the forecast differences. At 700 hPa (Fig. 18) the new analyses are drier everywhere, the largest changes occur in the deep tropics, changing from 80% to 30% in some regions. The cumulative impact of the model changes on the initial moisture field is very large. This is clearly seen in Fig. 18, which shows the 700 hPa relative humidity for the old and new analyses, and five-day forecasts. The differences between the analyses for the old and new models, or the forecasts for the old and new models, are clearly larger than the differences between the forecasts and analyses for the old model or for the new model. The large impact observed in the analysis indicates the present poor humidity data coverage and, more than for any other variable, the importance of the model for the initial humidity field through the first guess.

Figure 18. Relative humidity at 700 hPa. Ensemble forecasts and analyses. Values >80% shaded. Contour interval: 20%. (a) Analyses, old model; (b) 120-hour forecasts, old model; (c) analyses, new model; (d) 120-hour forecasts, new model.
(b) Wind fields

The analysis differences in the wind field are in general quite modest, particularly at low levels. There is an indication of a strengthening of the trade winds in the analyses at low levels and an increase in convergence along the ITCZ; but these changes are slight (−1 m s⁻¹).

Some larger changes occur in the upper troposphere particularly off the west coast of South America and over the Arabian Sea (Fig. 19), where changes of up to 6 m s⁻¹ occur. Once again, these are reflections of systematic differences in the forecasts influencing the analyses through the first-guess field. The forecast differences are, however, more widespread (Fig. 19(b)); the fact that these differences influence the analyses only in certain regions is related to the data coverage in these regions. Over the Indian Ocean for example cloud-track winds are not used by the analysis, because of problems with the quality of Inmetsat winds, and few other data are available in this region.

![Figure 19. Vector wind difference at 200 hPa: new minus old model. Ensemble forecasts and analyses. (a) Analyses, contour interval: 2 m s⁻¹; (b) 120-hour forecasts, contour interval: 5 m s⁻¹.](image-url)

6. Conclusions

The impact of revised physical parametrization and increased horizontal resolution on tropical forecasts and analyses has been assessed by forecast experiments and by ensemble means of forecasts and analyses for a twenty-day ‘parallel run’ of the two systems. Conclusions may be summarized as follows.

The parametrization of shallow cumulus convection increases the moisture flux out of subtropical boundary layers. A realistic trade wind inversion is now maintained showing the typical layered structure: boundary layer — cloud layer — inversion. The increased moisture being supplied to the boundary layer is transported into the tropics by the trade winds thus increasing the moisture source for deep cumulus convection. This increased moisture source, together with revisions to the Kuo scheme, produces much greater rainfall amounts, and a more realistic geographical distribution of precipitation. The ITCZs in particular are much improved and the equatorial trough is deepened.

The major change to the penetrative convection scheme is the redefinition of the moistening parameter. This change has the effect that more convective heating and less moistening is produced for a given amount of moisture supply through convergence. By
this means the model bias towards a too cold and too moist state is reduced. Additionally, redefinition of the cloud base resulted in a more realistic diurnal cycle of convection over the continents.

As a result of the increased latent heat release the model atmospheric energy balance is changed. Instead of a general cooling of the model atmosphere there is now a small warming, temperature errors being largely reduced in the tropical troposphere. The increased diabatic heating produces, primarily, a more intense Hadley circulation and, also, changes in the large-scale rotational flow; in particular more intense circulation associated with the subtropical highs.

Improvements to the large-scale flow become noticeable beyond day 2 and remain evident at least up to day 10. Substantial improvements in objective measures of forecast quality are also evident. However, the overall error growth is still very fast during the first day or so and this rapid initial growth is little affected by the model changes.

The model forecast differences were seen to strongly influence the analyses, particularly the temperature and moisture fields. In the case of the latter, the analysis structure appeared to be dominated by the influence of the forecast model. Analysis differences in respect of wind were confined largely to data-void regions, and in these regions reflected model differences.

In both the old and the new forecasts the return flow of the Hadley circulation is relatively weak and consequently occurs over a deep layer of the tropical upper troposphere, whereas the analyses show a well-defined return flow confined to just below the tropical tropopause.

APPENDIX A

Parametrization of shallow cumulus convection

Most forecast models use parametrization schemes which represent the mixing by boundary layer turbulence and possibly some aspects of clear air turbulence above the boundary layer. Subgrid-scale moist processes are then only considered in the deep convection scheme as a mechanism for the formation of precipitation. Systematic deficiencies were found in the simulation of the boundary layer thermodynamic structure and in the intensity of the tropical heat sources, with an earlier version of the ECMWF model (Tiedtke 1982). This finding motivated the development of a parametrization scheme that could better represent the vertical exchanges by non-precipitating cumulus clouds.

Basically, shallow convection can be parametrized in two ways. In the first method the cloud layer is represented as a whole and the bulk effects of convection are parametrized. This involves predicting the cloud layer depth and the net convective heating and moistening. Parametrization schemes along this line have been proposed by Augstein and Wendel (1980) and by Albrecht et al. (1979), in the framework of a mass flux formulation similar to the one used in the Arakawa–Schubert scheme (Arakawa and Schubert 1974). However, whereas Augstein and Wendel impose the closure assumption that cumulus convection is forced by boundary layer turbulence, Albrecht et al. (1979) use an adjustment approach. Augstein and Wendel's scheme has been successfully used for simulating the time evolution of the trade wind boundary layer during ATEX in a one-column model (Augstein and Wendel 1980), but has not, so far, been implemented in a large-scale model. The parametrization by bulk methods appears particularly suitable for models where the boundary layer processes below the cloud layer are similarly parametrized through a bulk approach. The adjustment scheme proposed by Betts (1986)
is an alternative method which does not require a full knowledge of the cloud layer physics and assumes a relaxation of model thermal states towards realistic quasi-equilibrium states. The scheme has been tested in a single-column model (Betts and Miller 1986) and in a global model (Miller 1986).

In models with sufficiently high vertical resolution a multi-layer approach which attempts to resolve explicitly the subcloud and cloud layer structure is justified. Vertical transports between layers can be computed by an eddy diffusion scheme or by a higher-order closure of the turbulence equations (Sommeria et al. 1986). The mass flux concept is also a convenient approach.

The results presented here are obtained with a simple eddy diffusion scheme. The scheme considers convective transports of sensible heat and moisture within the cloud layer, through cloud base and through the level of zero buoyancy, where cloud base and zero-buoyancy level refer to the lifting of near-surface air.

The effects of non-precipitating cumulus convection on the grid-area-mean heat and moisture budget are due to turbulent fluxes of sensible heat, moisture and condensation processes (Betts 1975):

\[
\left( \frac{\partial \bar{s}}{\partial t} \right)_{cu} = -\frac{1}{\rho} \frac{\partial}{\partial z} \left( \bar{\rho} \bar{w}' \bar{s}' \right) + L(C - E)
\]

\[
\left( \frac{\partial \bar{q}}{\partial t} \right)_{cu} = -\frac{1}{\rho} \frac{\partial}{\partial z} \left( \bar{\rho} \bar{w}' \bar{q}' \right) - (C - E)
\]

\[
\left( \frac{\partial \bar{I}}{\partial t} \right)_{cu} = -\frac{1}{\rho} \frac{\partial}{\partial z} \left( \bar{\rho} \bar{w}' \bar{I}' \right) + C - E
\]

where \( s \) = dry static energy, \( q \) = specific humidity, \( I \) = liquid cloud water, \( C \) = condensation rate, \( E \) = evaporation rate, \( L \) = latent heat of condensation, \( w \) = vertical velocity, \( \rho \) = density, \( t \) = time, \( z \) = height. Bar and prime denote grid-area-mean values and convective-scale fluctuations, respectively, and \( \left( \frac{\partial}{\partial t} \right)_{cu} \) is the local tendency due to convection.

For non-precipitating cumulus convection the time change and advection of liquid cloud water \( I \) by the large-scale flow are typically small; consequently the balance is between the turbulent transport of liquid cloud water and the result of the condensation-evaporation process

\[
C - E = \frac{1}{\rho} \frac{\partial}{\partial z} \left( \bar{\rho} \bar{w}' \bar{I}' \right).
\]

Thus, the effect of shallow cumulus convection can be described by the turbulent fluxes of sensible heat, moisture and liquid cloud water, which in the case of an eddy diffusion approach are parametrized as

\[
\left( \frac{\partial \bar{s}}{\partial t} \right)_{cu} = \frac{1}{\rho} \frac{\partial}{\partial z} \left[ \bar{\rho} K \frac{\partial}{\partial z} (\bar{s} - L \bar{I}) \right]
\]

\[
\left( \frac{\partial \bar{q}}{\partial t} \right)_{cu} = \frac{1}{\rho} \frac{\partial}{\partial z} \left[ \bar{\rho} K \frac{\partial}{\partial z} (\bar{q} + \bar{I}) \right].
\]

In the present framework the difficulties lie in the specification of the eddy diffusion coefficient \( K \) and of the grid-area-mean liquid water content.
The formation of cumuli is more effectively controlled by sub-cloud layer turbulence than by the thermal state above the cloud base. A parametrization along these lines \((K)\) determined from sub-cloud layer turbulence has been proposed by Tiedtke (1988), but has not yet been tested in a global model. Here a scheme is applied which contains further simplifications, i.e. the liquid cloud water transport has been neglected and the eddy diffusion coefficient is prescribed. Neglecting the liquid cloud water transport is partly justified in view of results obtained by Soong and Ogura (1980) which indicate that it is much smaller than the turbulent transport of moisture. The eddy diffusion coefficient is assumed to be constant throughout the active (=buoyant) cloud layer and is dependent on the moisture distribution in the passive cloud layer (the layer above the zero-buoyancy level) which contains the inversion layer. It is zero elsewhere. Thus

\[
K = \begin{cases} 
K_o & \text{active cloud layer} \\
K_o \left( \frac{RH - 0.8}{0.2} \right) & \text{inversion layer} \\
0 & \text{elsewhere.} 
\end{cases}
\]

The value of \(K_o = 10 \, m^2/s\) has been determined from BOMEX and ATEX data; \(RH\) is the relative humidity (0–1) below the inversion and \(\Delta RH\) the relative humidity jump across the inversion. Although convective transport of sensible heat and moisture across the inversion due to overshooting of cumuli above the zero-buoyancy level has been found to be important for a realistic simulation of the trade wind lower troposphere, it is not considered when the air just below the inversion is dry as it is assumed that cloud air then evaporates effectively into the environment.

The scheme effectively transports moisture from the cloud-free part of the boundary layer into the cloud layer and inversion layer above. This is most pronounced in the trades, where shallow convection counteracts the drying and warming effects of the large-scale subsidence, as is clearly evident in one-column simulations using BOMEX and ATEX data, the latter being shown here as an example. The ATEX data refer to an undisturbed period of the Atlantic trade wind (7–12 February 1969) where the synoptic-scale flow was steady throughout the period and the thermal structure of the lower troposphere was essentially maintained by a balance of the mean atmospheric downward motion and the turbulent and convective mixing from below.

The parametrization scheme is verified in a one-dimensional model, where all but turbulent and convective processes are prescribed. The prescribed adiabatic processes and radiative cooling, which refer to the whole period, are taken from Wagner (1975). Turbulent processes are parametrized by the boundary layer scheme of the ECMWF forecast model. The initial data correspond to the observed time mean flow and the forecasts are also verified against this mean flow (steady-state condition!). The vertical ascent (Fig. A1(c)) shows the typical structure in the trades, with a well-mixed boundary layer capped by a conditionally unstable cloud layer and a strong inversion above.

The experiments show that when the shallow convection scheme is not included (Fig. A1(a)) the inversion becomes much too strong and descends within two days from 850 to 925 hPa. Also, as the turbulent moisture transport is confined to the layer below the inversion, the boundary layer becomes totally saturated within two days. When shallow convection is included (Fig. A1(b)) the thermal structure of the lower troposphere is well maintained, showing the typical 4-layer structure of the trades (i.e. boundary layer—cloud layer—inversion layer—free atmosphere). The cloud layer is maintained
by the additional convective transports of heat and moisture, giving a cooling and moistening of the upper part of the cloud layer of 4 K d\(^{-1}\) and 4 g kg\(^{-1}\) d\(^{-1}\) respectively, which nearly compensates the effects of the mean downward motion.

Although this scheme is rather simple and does not describe the various processes involved in non-precipitating cumulus convection, it produces realistic net diabatic heating and moistening rates for the large-scale environment. This is because the effect of non-precipitating cumulus convection on the environment is through redistributing heat and moisture and so it can be parametrized by a diffusion process.

**APPENDIX B**

*Parametrization of penetrative cumulus convection*

Penetrative convection is parametrized by a scheme (Kuo 1974), in which cumulus convective heating and moistening are related to the local moisture supply due to large-scale convergence and surface evaporation.

Two modifications have been made to the Kuo scheme. Firstly, the cloud base is redefined as the condensation level for surface air (lowest model level) rather than that for air with the mean characteristics of the sub-cloud layer. This change enhances the occurrence of cumulus convection which was previously underestimated, and gives a more realistic response, by the convection, to the diurnal cycle in surface heating over the continents. The second important change involves the moistening parameter \(\beta\), which determines the partitioning between convective heating and moistening. Various methods of specifying \(\beta\) have been proposed in the past but we follow the method proposed by
Anthes (1977) where the partitioning of convective heating and moistening is assumed to depend on the mean saturation deficit of the whole layer:

\[
\beta = \begin{cases} 
\left( \frac{1 - \text{RH}}{1 - \text{RH}_{\text{crit}}} \right)^n & \text{if } \langle \text{RH} \rangle > \text{RH}_{\text{crit}} \\
1 & \text{if } \langle \text{RH} \rangle \leq \text{RH}_{\text{crit}}
\end{cases}
\]

where \( \text{RH}_{\text{crit}} \) = threshold relative humidity below which only moistening of the air occurs; \( \langle \text{RH} \rangle \) = mean environmental relative humidity for the cloud layer. \( n \) and \( \text{RH}_{\text{crit}} \) are disposable parameters.

In the previous model a linear dependence on the moisture deficit (\( n = 1 \)) was used. This formulation tended to overestimate the moistening and underestimate the convective heating. This deficiency was particularly noticeable in a simulation of intense cumulus convection during GATE (using the composite data for phase III, Thompson et al. (1979)). The simulation was much improved when a stronger moisture dependence was used (\( n = 3, \text{RH}_{\text{crit}} = 0 \)) which provides less moistening and more heating and as a result a more realistic thermal state. Although the assumption about the moistening parameter, and in particular the special choice about the moisture dependence (\( n = 3 \)), appear rather arbitrary, there is some evidence from observational data that values of \( n \) between 2 and 3, and \( \text{RH}_{\text{crit}} \) between 0·25 and 0·50 provide the best agreement between simulated and observed rainfall rates (Kuo and Anthes 1984). The value \( \text{RH}_{\text{crit}} = 0 \) appears rather small compared with the value diagnosed by Kuo and Anthes, whose results came to our attention after our experiments had been completed. More recent test forecasts carried out with a larger value of \( \text{RH}_{\text{crit}} \) (0·5) show a reduced spin-up in precipitation during the early stages of the forecasts.

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