DATA ASSIMILATION IN THE TROPICS

William A. Heckley
European Centre for Medium-range Weather Forecasts
Shinfield Park, Reading, U.K.

Abstract

The role of the data assimilation system and its component parts is reviewed with emphasis on its application to the tropics. A number of examples are chosen to illustrate the special difficulties associated with analysis in these regions, and ways to address these problems are discussed. The overall performance of the system is reviewed. Whilst there are fundamental difficulties (e.g. poor first-guess, lack of data, weak mass-wind coupling), the assimilation is found to behave reasonably well. New techniques and data sources offer much hope for improvement.

1. INTRODUCTION

Within a Centre devoted to medium-range weather forecasting for Europe, it is reasonable to ask the question 'Why analyze the tropics?' Indeed, such a question has been posed in the past. Today, with an entire seminar series devoted to tropical-extratropical interaction, the answer is clear, at least on medium to long time scales. Overwhelming evidence exists for teleconnection patterns emanating out of the tropics and influencing the weather of the northern latitudes. Many phenomena, such as the 30-60 day wave, are now seen to be truly global in character. This interaction is discussed extensively by other authors in this volume and the reader is referred to these works. Tropical-extratropical interaction is not just confined to the large space and time scales. A glance at the tracks of tropical storms over the Atlantic during the last twenty years (Fig.1, from Neumann, 1988) shows immediately that many of these storms move northwards out of the tropics into the sub-tropics, and often into the extratropics; turning into, or modifying existing, extratropical systems in the process. Thus there is often a fairly direct and short time-scale interaction between the tropics and extratropics. A notable and fairly recent example of this type of interaction is the so-called 'great storm' of 15-16 October 1987 (see for example, Burt and Mansfield, 1988), which caused very severe and widespread damage to southern England and northern France. This system could be traced back to its origins as a tropical storm in the Atlantic. If the models are to stand a chance of correctly predicting such phenomena they must start out with an accurate analysis.

The main objective of the data assimilation scheme is to provide reliable initial states for the forecast model. Other objectives are verification of forecasts, and diagnostic studies. The analyses must be global, in numerical form, and use all appropriate types of observation available. An additional requirement, at ECMWF, is that the data assimilation system must be able to run efficiently, with minimal human intervention. All these requirements impose constraints on the design of the assimilation scheme.

1.1 The Assimilation System

Information is available in the form of observations, short range forecasts, a priori knowledge of the characteristic structures of atmospheric flow (e.g. near-hydrostatic balance, quasi-geostrophy etc.), climatological behaviour, and so on. One also knows, or may estimate, the accuracy (relevant information content) of each of these different types of information. One is trying to produce the *optimum* initial state for the forecast model. The model is unable to represent (due to resolution, or other deficiencies in formulation) certain real atmospheric phenomena. Measurement accuracy of, say, a radiosonde wind measurement may be quite high, but part of the measurement is of structures that the forecast model is unable to assimilate.

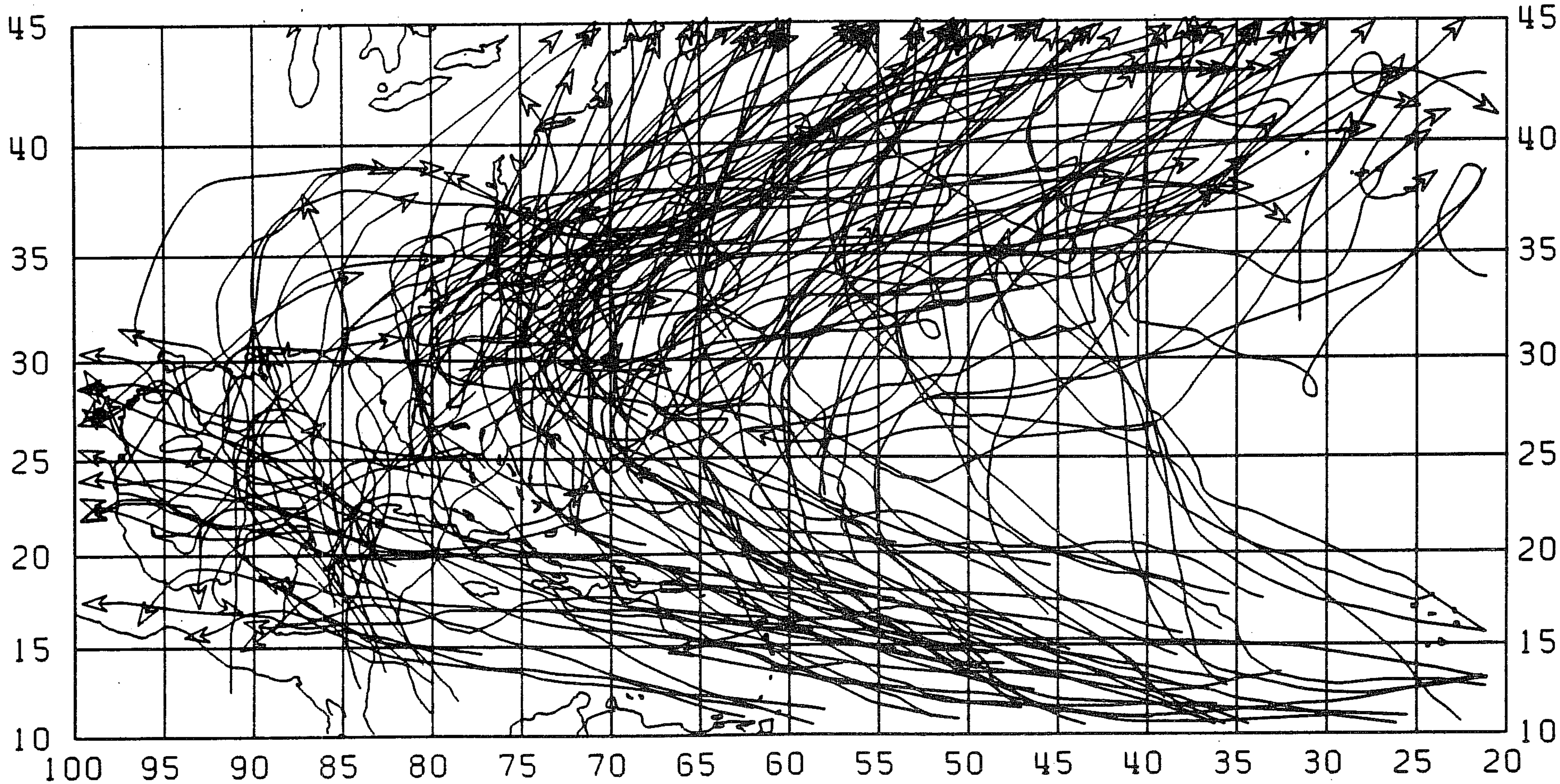


Fig.1 Tracks of 141 tropical storms and hurricanes, 1962-1981. (From Neumann (1988))

Thus the observation, though accurate, may be 'unrepresentative'. The error assigned to each observation must take this into account. Such error assignments are based largely on experience, assimilation studies, and to some extent, guesswork. The data assimilation scheme must combine all these different types of knowledge in such a way as to extract the maximum information on the state of the atmosphere, consistent with the limitations of the model. A common operational approach is to use a form of four-dimensional assimilation. A forecast model is run continuously and model variables are 'corrected' from time to time as observations become available. In this way the model fields are 'nudged' towards the observations. Observations can be inserted at, or around, the observation time, perhaps repeatedly; or gathered together and inserted intermittently (for example every six hours). The latter, *intermittent data assimilation*, is the type currently employed at ECMWF. The former, *continuous data assimilation*, is carried out at the UK Met Office, for example (Lorenc et al., 1991). A natural extension to these techniques is four-dimensional variational assimilation. In this procedure a repeated integration is carried out with modified initial states to find the model trajectory (forecast) which best fits the observations and/or other constraints. The technique of modifying the initial state involves the use of the adjoint of the forecast model, followed by sophisticated (and expensive) minimization procedures. Many groups, as well as ECMWF, are beginning to experiment with this type of approach. A review of assimilation methods is given by Hollingsworth (1986), and also Pailleux (1990).

The type of intermittent data assimilation currently employed at ECMWF is illustrated in Fig.2. There are three main components: a six-hour forecast, an analysis, and an initialization. These three contiguous steps are repeated continuously. It is assumed that most of the evolution of the flow in a six-hour period is captured by the short forecast. It is the role of the analysis step to 'correct' any errors in this forecast. This is accomplished through interpolation of the forecast to the observation points and examination of the differences between the observations and the forecast values. These differences or 'observational increments' are then used to modify the neighbouring values on the forecast grid, both in the horizontal and in the vertical. In order to accomplish this modification, empirical functions are used to describe the characteristic scale and shape of the forecast errors. These functions have been empirically determined through statistical comparison of forecast errors against dense radiosonde networks (e.g. Hollingsworth and Lönnberg, 1986; Lönnberg and Hollingsworth, 1986). At the same time knowledge of the balance conditions of the atmosphere are used to relate the wind changes to mass changes, and visa versa. Thus if the observations are being used to correct the wind field, geostrophic balance is invoked (at least within the extra-tropics), in order to make a corresponding change to the mass field. As well as making the analysis problem less 'under determined', it helps to ensure that the model retains the information (although if the atmospheric flow is far from geostrophic, this approach can cause severe instability in the assimilation scheme (Heckley, 1989)). After the analysis an initialization step is carried out. Its purpose is to control 'noise' in the short-range forecasts used in the assimilation. This is necessary as the short forecast is also used as a check for bad observations; a noisy short-range forecast can therefore lead to poor quality control decisions.

For more details of the ECMWF assimilation system the reader is referred to Lorenc (1981), Bengtsson et al. (1982), Shaw et al. (1987), Wergen (1988,1989) and Undén (1989).

This general framework of data assimilation is valid globally, but the tropics pose many special and peculiar difficulties. In the rest of this paper a broad cross-section of these problems are discussed in order to give some flavour of the difficulties. Some current attempts to circumvent the problems are described, as are hopes for the future. It is not all doom and gloom in the tropics. Although there are many problems much progress has been made and there is considerable hope for the future. In many ways the tropics are inherently more predictable than the extra-tropics, responding as they do to the relatively slowly varying boundary forcing.

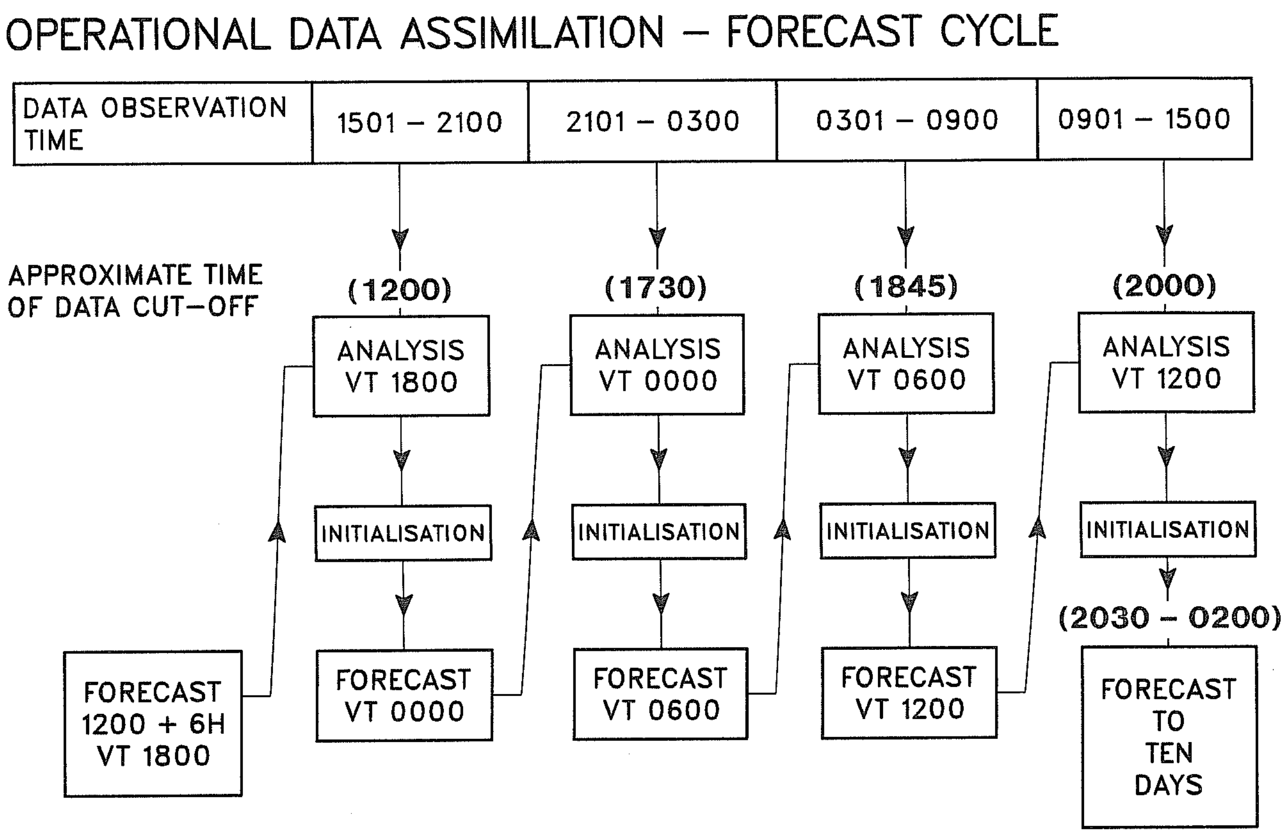


Fig.2 The ECMWF data assimilation scheme.

2. SPECIAL DIFFICULTIES WITH TROPICAL ANALYSIS

2.1 Data Availability

This is without doubt the most serious problem as far as data assimilation is concerned. It is not that there are no data, or even a lack of it. Satellite temperature sounding data, for example, are widespread. However, as we shall see later, such data are of little use in the tropics. A truer statement might be 'lack of useful data'.

Radiosondes are the only operational instruments which give us detailed and reliable information on the vertical structure of the atmosphere. Fig.3 shows the distribution and reception rates of radiosonde reports, as received at ECMWF, during April 1984. The northern hemisphere continents are well covered, and reception rate is usually good. Within the tropics, coverage, with certain notable exceptions, is minimal and reception rates of, even these, is low. Central America/Caribbean, India, Australasia are relatively well covered. However, most of Africa, South America and virtually all of the oceanic areas are very thinly covered. The situation today is much the same.

An example of the total observations received at ECMWF at around 00 UTC on 28th April 1984 (fairly typical), is shown in Fig.4. This shows a) Surface observations from land stations (SYNOP) and ships, b) drifting buoys, c) radiosondes (TEMP), d) satellite temperature sounding data (SATEM), e) wind sounding data from PILOT balloons, f) satellite cloud-tracked wind estimates (SATOBs), g) aircraft reports (AIREP). The SYNOP and TEMP network is less dense than in the northern hemisphere extratropics. PILOT reports are only available in a few regions (mainly west Africa, Australasia), AIREPs are almost totally absent over most of the tropical belt, SATEMs provide good coverage (but these are of too low a quality to be of much use in the tropics). However, there is a fairly good coverage of SATOBs, unfortunately these only define the wind at one level, and due to quality control problems there is usually a complete absence over the Indian Ocean.

With the exception of a few areas mentioned above, which are relatively well served with radiosondes, there are relatively few data available in the tropics - and what there are tend to provide datum at a single height. This general lack of data leads to large analysis uncertainty. SATOBs, which have a relatively wide coverage, are very important to tropical analysis. However, there are few problems in data selection, as sometime occur in northern latitudes. A corollary of this is that there are probably more gains to be had, in terms of the final quality of the analysis, by improving the analysis technique than would be the case in a data dense configuration.

New observing techniques and new instruments offer vastly increased quantities of data, for example scatterometer wind estimates yield surface wind data over the oceans with a horizontal resolution of 100kms (Anderson et al., 1987). Some of these will be discussed in section 7.

2.2 Quality Control

2.2.1 Nature of the tropical flow

An inherent difficulty in data assimilation in the tropics, that is not often mentioned, relates to the nature of tropical flow itself. The climatological variance of the flow is much smaller than in mid-latitudes. There exist very large-scale quasi-stationary systems with a time-scale of 30-40 days. Embedded on these are small-scale transient disturbances with life-cycles of 3-4 days. Both are important for tropical 'weather'. This combination of quasi-stationary flow and infrequent small-scale perturbations (often of large amplitude) plus a lack of data leads to severe difficulties in quality control of the data. The low climatological variance of the tropical atmosphere means that errors that might be considered as small in an extra-tropical context are *large* in the tropics.

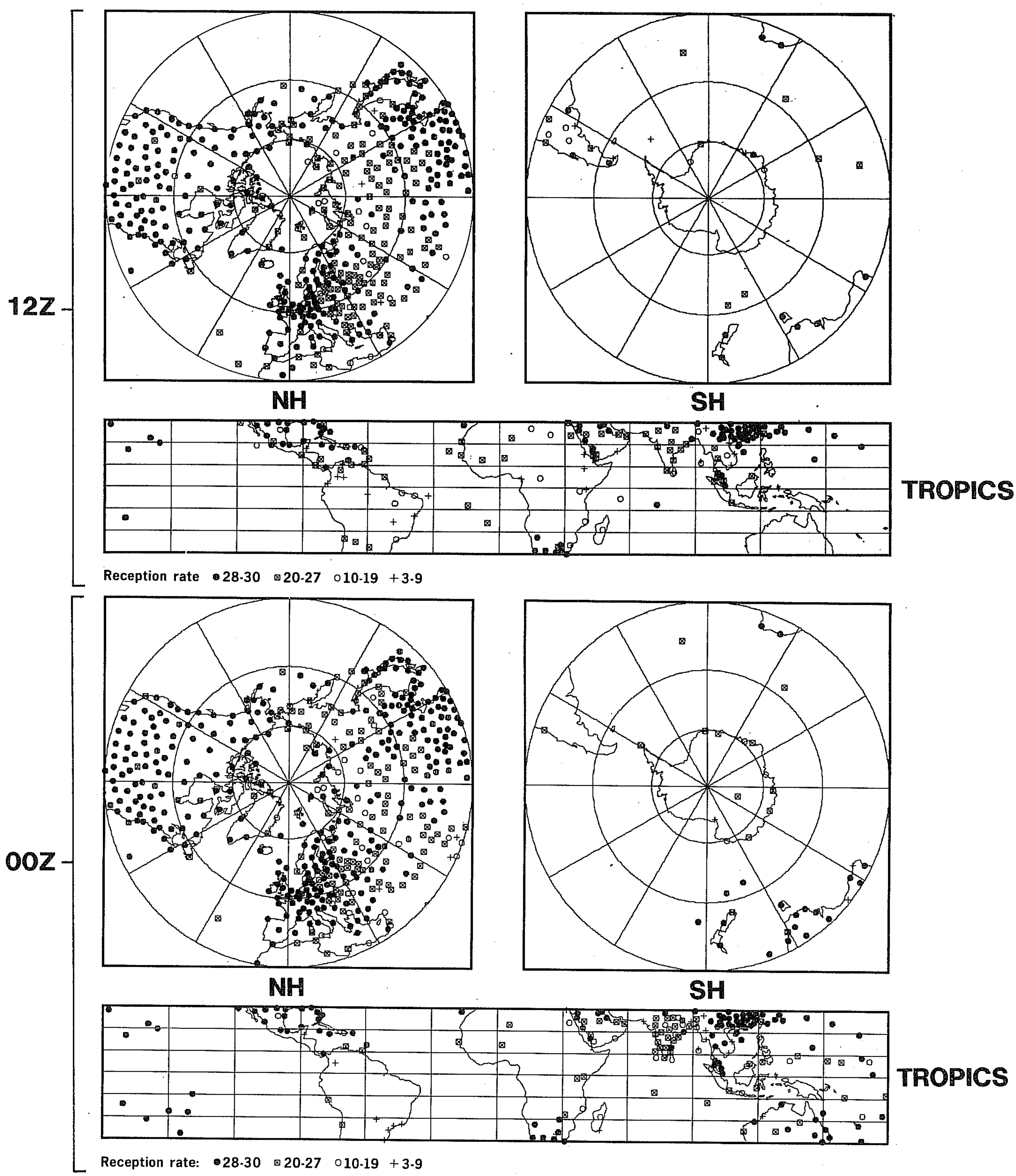


Fig.3 Distribution and reception rate of radiosonde ascents, from land stations, at ECMWF during April 1984. Upper panels 12 UTC; lower panel 00 UTC.

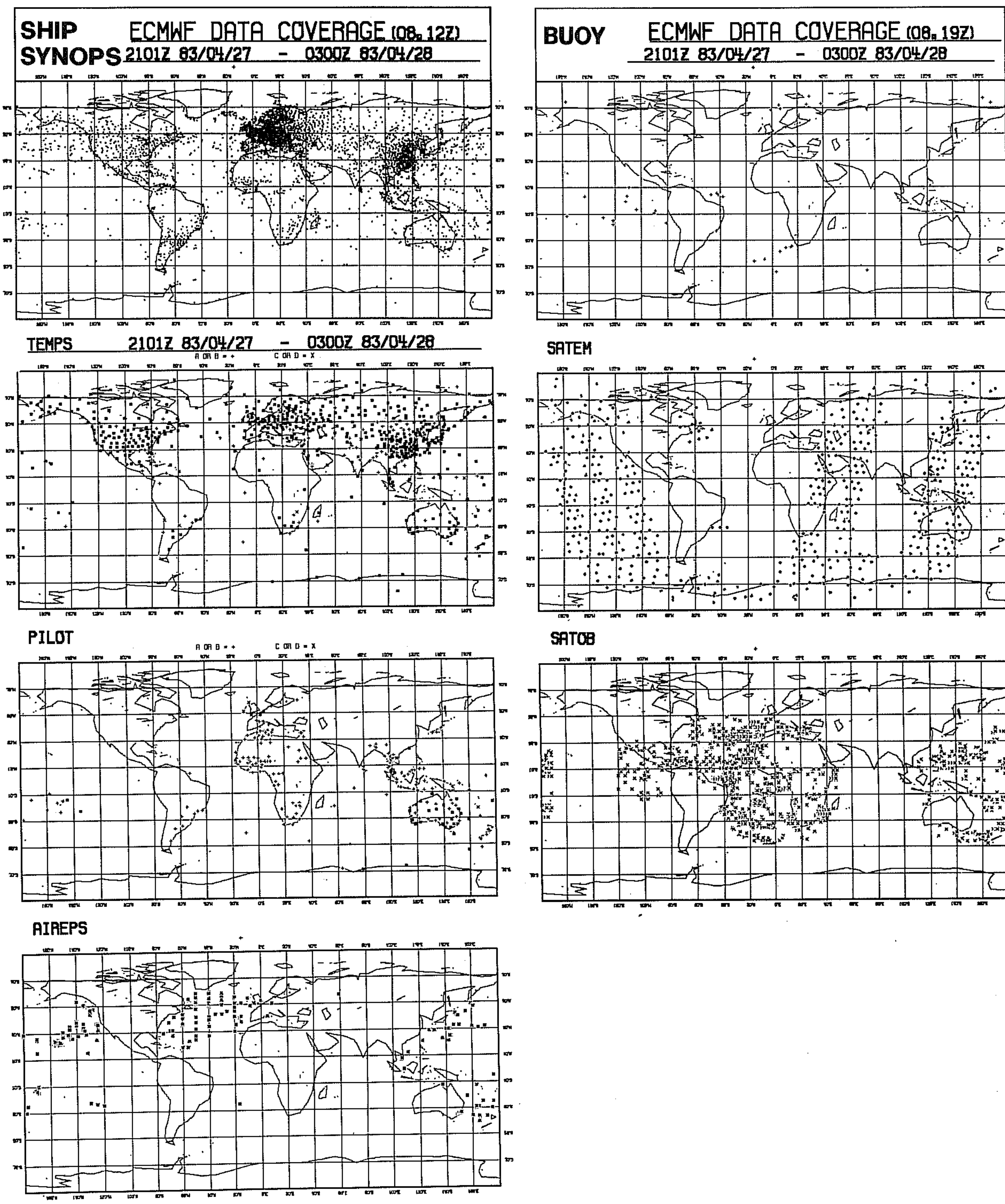


Fig.4 Available observations at ECMWF on 28 April 1984, 00 UTC.

Within the tropics there is a predominance of diabatic forcing. These processes are not well understood, and even less well modelled. This leads to a relative poor performance of the forecast model compared to the extratropics, where such processes exert less of an influence. These factors mean that the short-range forecast is a relatively poor 'first-guess' in the tropics. Compounding the quality control problems discussed above.

2.2.2 Isolated observations

When there are many, independent, observations in close vicinity one may use them as quality control checks against each other. Often, within the tropics, one is faced with the situation in which one has either isolated observations, or many observations with shared errors (e.g. satellite derived fields).

Occasional large departures from the first guess are to be expected, but repeatedly large, similar, departures over a long period are unlikely. For this reason it is useful to monitor the averaged departures of observations, or the analysis, from the first-guess over a long period of time (say a month). If this is done then one occasionally observes large departures, which have often been traceable to biases in the measuring instruments at particular stations. Such monitoring allows a feedback with the data producers and improvement in data quality at source.

2.2.3 Biased observations

Isolated, consistently biased observations tend to be picked up in the long term monitoring statistics, as mentioned above. Large quantities of similarly biased observations are less easy to deal with.

The importance of SATOBs for tropical analysis was emphasized in 2.1. They are the principal data source over much of the tropics. Fig.5 (from Eriksson, 1991) shows an analysis (an average over an eight day period) a) using SATOBs, and b) without. The flow over the equatorial Atlantic is completely different. However, these data are not without their problems. Fig.6 (also from Eriksson, 1991) (a) shows the average departures of observations from the first-guess (during December, 1989; 500-100hPa, 150hPa for TEMPs) according to speed classes, for the northern hemisphere and also for the tropics. The TEMP data are the most accurate. Within the northern hemisphere, one can see little bias against TEMP data, except for very high wind speeds - in which the first guess seems a little slow. Comparison against AIREPs reveals much the same story. SATOB data, on the other hand, are generally too slow with respect to the first guess. This suggests that the SATOBs are biased towards slow wind speeds. A conclusion consistent with earlier results of Källberg (1985), who pointed out that clouds are not simply advected along with the environmental flow at the cloud level. Within the northern hemisphere extratropics (where the first guess is fairly accurate) one may control the use of these data by rejecting any which depart too far from the first-guess.

Within the tropics, SATOBs are the dominant data source. Looking at Fig.6 (b) one sees that the first-guess, within the tropics, is biased low against SATOBs, AIREPs and TEMP data. With respect to SATOBs this bias remains fairly constant, and small throughout the whole speed range; whereas it dramatically increases for higher speeds compared to AIREPs and TEMPs - being biased too slow. This admits of two possible explanations a) given an unbiased initial state the model itself tends to produce a systematic bias within the tropics towards low wind speeds (model climate drift); b) the large amount of SATOB data biases the analysis, and the forecast model faithfully retains this bias during the six-hour forecast used to produce the first-guess. Further assimilation studies by Eriksson (pers comm) indicate forecast-model bias to be the dominant contributor in this respect. This could cause quality control difficulties because both first-guess and SATOBs are biased in the same sense and to a similar degree; thus the SATOBs 'look' relatively good to the quality control, whereas the more accurate TEMP and AIREP data tend to look suspicious. In practice, the first-guess field is assigned such a large error that most

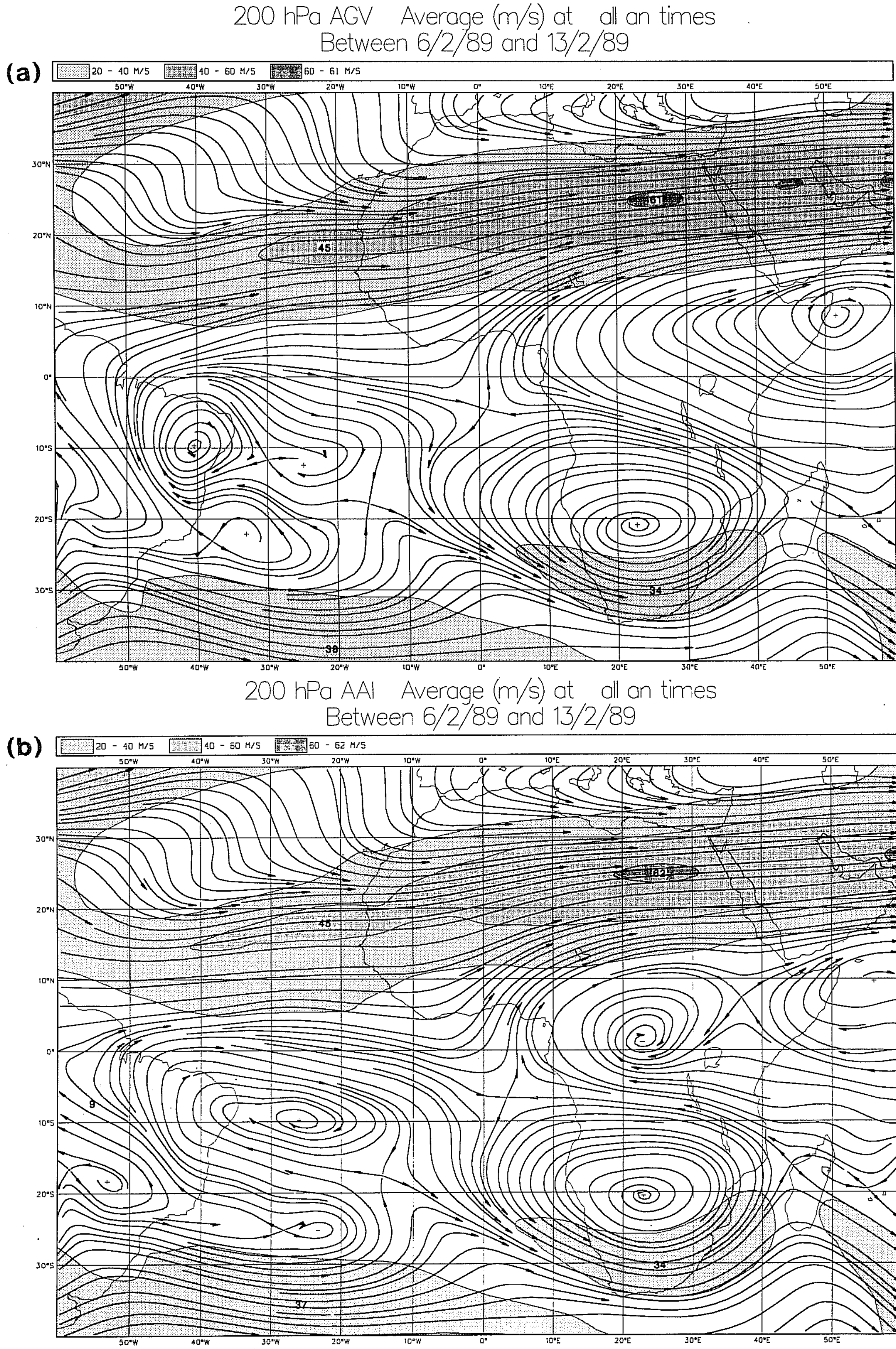


Fig.5 Impact of SATOB data on tropical analyses. (From Eriksson, 1991)
a) with SATOB data; b) without SATOB data.

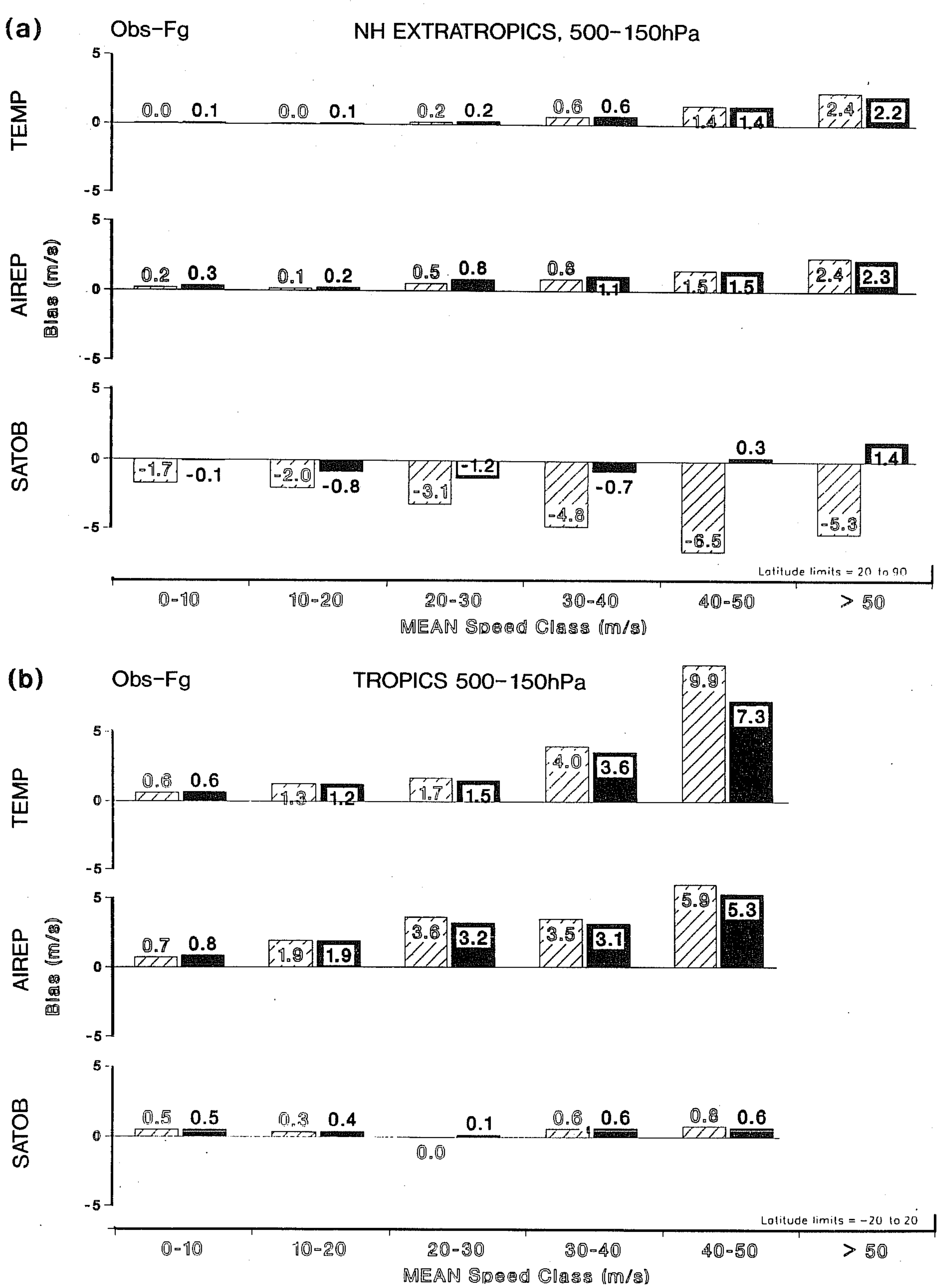


Fig.6 Observation/first-guess differences in speed classes.
December 1989, 500-100hPa (150hPa for TEMPs).
a) northern hemisphere; b) tropics.
Hatching refers to all data, shading to data presented to the analysis after quality control. (From Eriksson, 1991)

of the data is accepted.

2.2.4 Subjective intervention

Subjective intervention is sometimes used to supplement objective checks (Krishnamurti et al., 1981). Such intervention can be helpful in improving on general objective quality control judgements, and specifically in accommodating the problems associated with severe weather events discussed above. It can also be an effective means of incorporating additional information; based on either data not exploited by the objective analysis (data such as satellite imagery) or on data which cannot be exploited by the analysis because of limitations on the analysis technique. Such techniques are not used at ECMWF. This subject is taken up further in section 4.1.

2.3 Influence of Model Errors

Model errors influence the analyses through the first-guess forecasts. Errors in these may be broadly thought of as arising from two sources a) initial adjustments during the first few days associated with subtle imbalances in the initial thermodynamic fields; and b) a rapid drift of the forecast fields towards the model's own preferred climate. Sections 2.3.1-2.3.3 discuss the former, and 2.3.4 the latter.

When a major change to the operational forecasting system is planned, it is the practice at ECMWF, and other Centres, to carry out an extended 'parallel run' of the new and old systems. This leads to two completely independent sets of analyses and forecasts having as input, exactly the same data. This was done, for example, before the May 1985 change to the ECMWF forecast model. The change to the model included: introduction of a parametrization of shallow convection; a modification to the parametrization of deep cumulus convection; a new cloud scheme for radiation; and an increase in horizontal spectral resolution from T63 to T106. These changes and their impact are discussed in Tiedtke et al. (1988). The analysis and initialization components of the assimilation remained unchanged at this time. Thus any changes in the analysis fields themselves must be directly attributable to the influence of the forecast model through providing the first-guess. Tiedtke et al. (1988) studied the impact of the changes on the ensemble of model forecasts, and then compared this with the impact on the ensemble of analyses (20 days).

Forecast impact was found to be very large in the tropics. Parametrization of shallow convection increases the moisture flux out of the subtropical boundary layers. A realistic trade wind inversion is obtained with the new model. Increased moisture being supplied to the boundary layer is transported into the deep tropics by the trade winds thus increasing the moisture source for deep cumulus convection. This increased moisture source, together with revisions to the deep convection scheme, produces much greater rainfall amounts, and a more realistic geographical distribution of precipitation. The ITCZ's in particular are much improved and the equatorial trough is deepened. The major change to the deep convection scheme is in the definition 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 previous model bias towards a too cold and too moist state is reduced. As a result of the increased latent heat release the model energy balance is changed. Instead of a general cooling of the model atmosphere, the new model exhibited a small warming, temperature errors being largely reduced in the tropical troposphere. The increased diabatic heating produced, primarily, a more intense Hadley circulation and also, changes in the large-scale rotational flow; in particular more intense circulation associated with the sub-tropical highs.

With these effects on the forecasts in mind let us turn to the analyses.

2.3.1 Thermal structure

The significant differences are by and large confined to the oceanic areas. At very low levels, 1000hPa, the new analyses are slightly colder by generally less than 1K over the oceans in the deep tropics. At 850hPa (Fig.7a) the new analyses were slightly warmer over the oceans in the deep tropics (by up to 1K) and cooler over the sub-tropical oceans (by 1-2K). Fig.7b shows the difference between the 120-hour forecasts at 850hPa; the pattern is very similar to the analysis differences shown in Fig.7a, but the forecast differences also indicate a more general warming in the new model. At 700hPa (Fig.7c) the new analyses are warmer throughout the tropics, over the oceans, by up to 3K. Again, the pattern is very similar to the five-day forecast differences (Fig.7d).

The analysis differences, for the most part, reflect the differences in the forecasts from the two systems. In particular, the cooling at 850hPa over the subtropical oceans is consistent with the presence of a cloud layer maintained below subsidence inversions in the new model and the warmer troposphere is in agreement with the increased convective activity.

2.3.2 Moisture field

As for the temperature field, the analysis differences in the moisture field generally reflect the forecast differences. At 700hPa (Fig.8) 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 in the initial moisture field is very large. This is clearly seen in Fig.8, which shows the 700hPa relative humidity for the old and new analyses, and five-day forecasts. The differences between the analyses for the old and new models, are clearly larger than the forecasts and analyses for the old model, or for the new model. The large impact observed in the analyses indicates the present poor coverage of humidity data, and more than for any other variable, the importance of the model for the initial humidity field through the first-guess.

2.3.3 Wind field

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 ($\sim 1ms^{-1}$).

Some larger changes occur in the upper troposphere, particularly off the west coast of South America and over the Arabian Sea (Fig.9), where changes of up to $6ms^{-1}$ occur. Once again, these are reflections of systematics differences in the forecasts influencing the analyses through the first-guess field. The forecast differences are more widespread (Fig.9b); the fact that these differences influence the only in certain regions is related to the data coverage in these regions. Over the Indian Ocean, for example, SATOBs are not used by the analysis, because of quality control problems with the INMETSAT winds, and few other data are available in this region.

2.3.4 Spin-up

It has been observed from the very earliest forecasts that there is a tendency for model precipitation to undergo large changes during the first few days of the forecasts (Girard and Jarraud, 1982; Heckley, 1985). This is accompanied by a dynamical adjustment (Heckley, 1982). Kasahara et al. (1988) suggests that this is a feature common to all forecast models. It is largely a tropical problem (Heckley, 1985) and within the tropics is largely confined to the oceanic areas (Wergen, 1988). Typically it has the form that within these regions precipitation is initially very weak, almost absent, and continues to be weak at six hours, by 24 hours it has become over-intense, then by day two or three it reaches fairly stable values. It is sensitive in form to changes in the parametrization of deep cumulus convection (Illari, 1987). For example the May 1989 changes to the ECMWF forecast model resulted in the opposite behaviour of initial precipitation too strong and then 'spinning-down' (Arpe, 1989). Clearly it is indicative of inconsistency

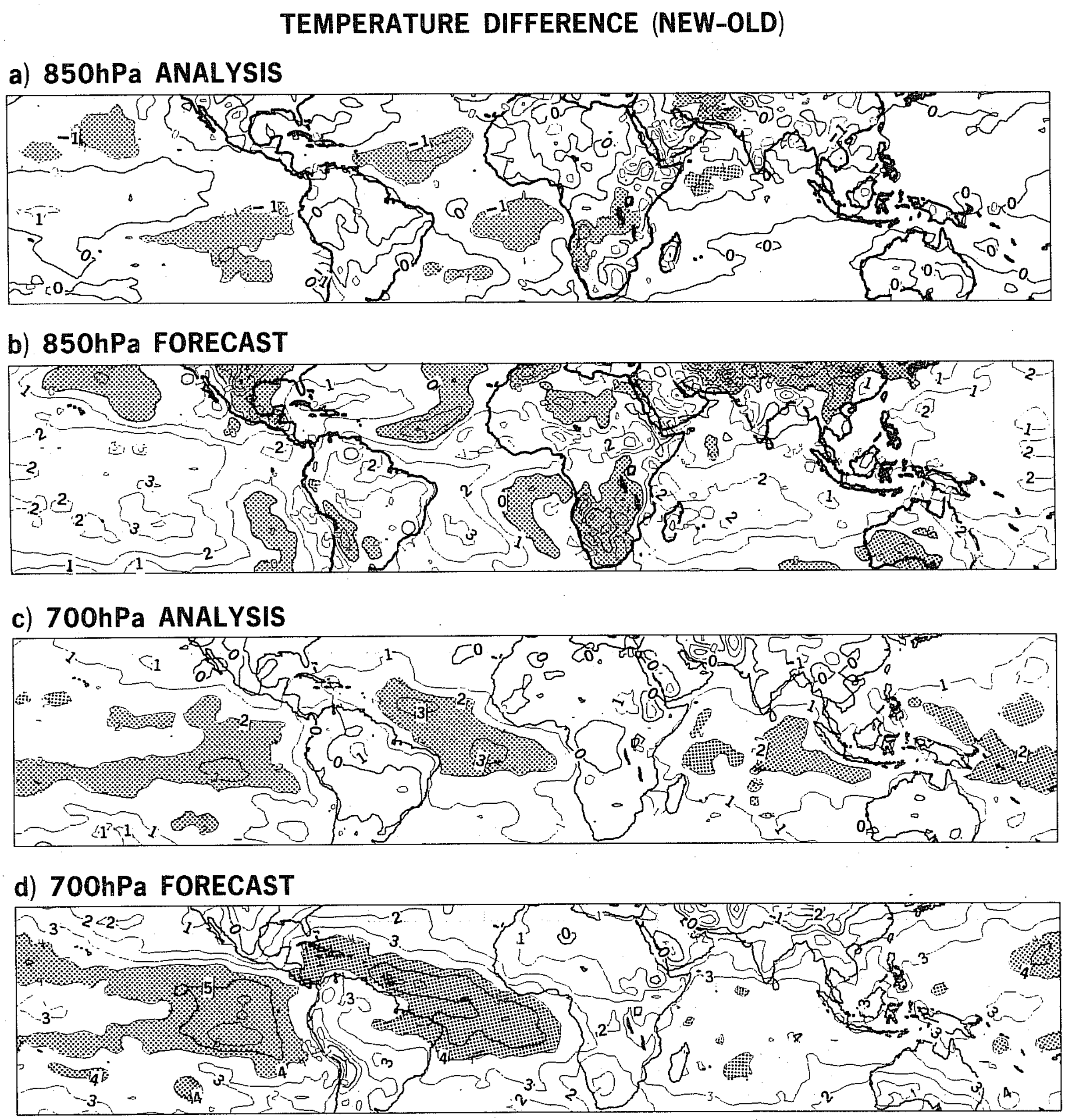


Fig.7 Temperature differences: new minus old model. Ensemble forecasts and analyses. Contour interval: 1K. (a) Analyses at 850Hpa, differences <-1K shaded; (b) 120-hour forecasts at 850hPa, differences <0K shaded; (c) analyses at 700hPa, differences >2K shaded; (d) 120-hour forecasts at 700hPa, differences >4K shaded. (From Tiedtke et al., 1988)

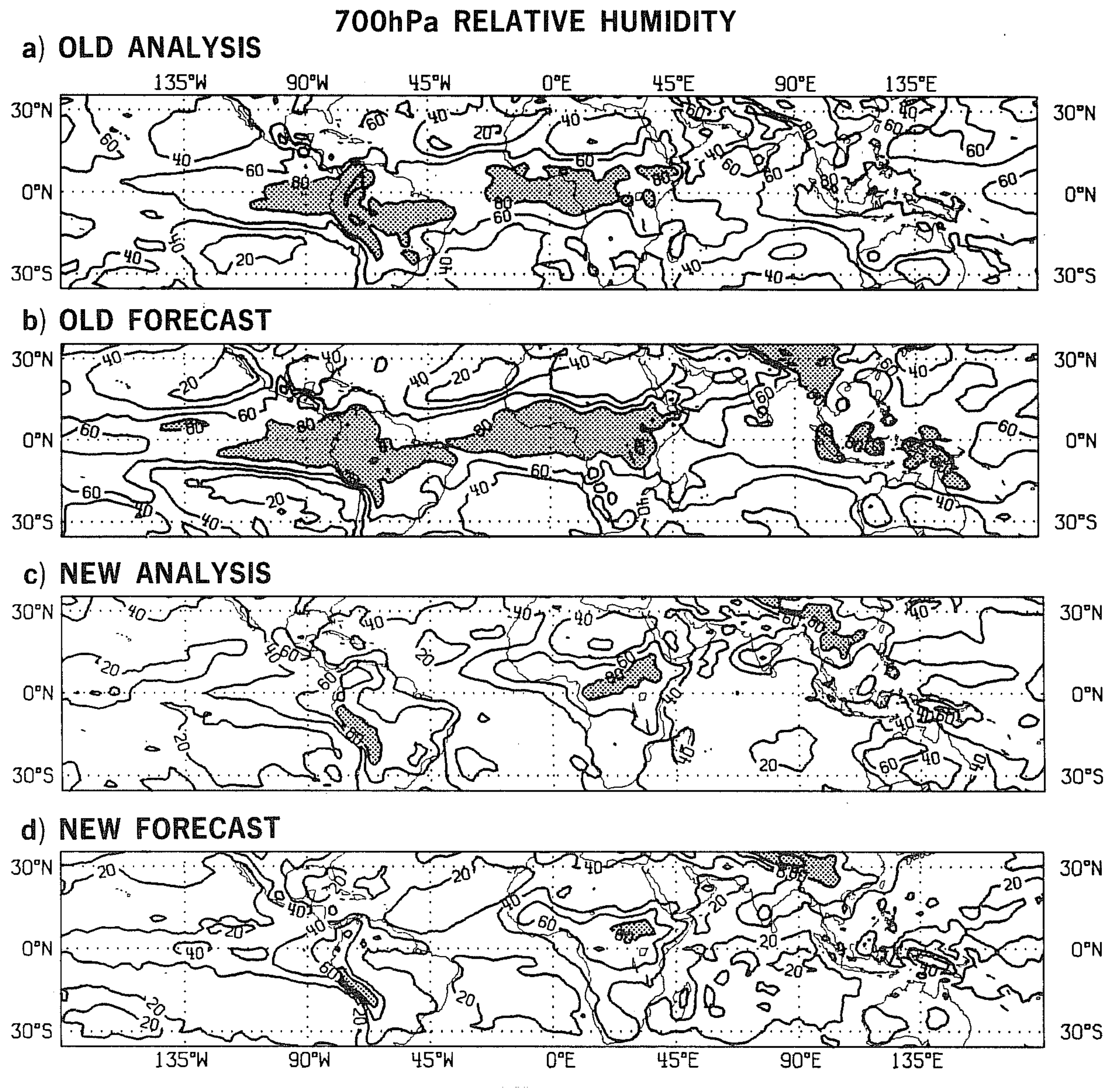


Fig.8 Relative humidity at 700hPa. Ensemble forecasts and analyses. Values >80% shaded. Contour interval: 20%. (a) Analyses, old model; (b) 120-hour forecasts, old model; (c) analyses, new model; (d) 120-hour forecasts, new model.
(From Tiedtke et al., 1988)

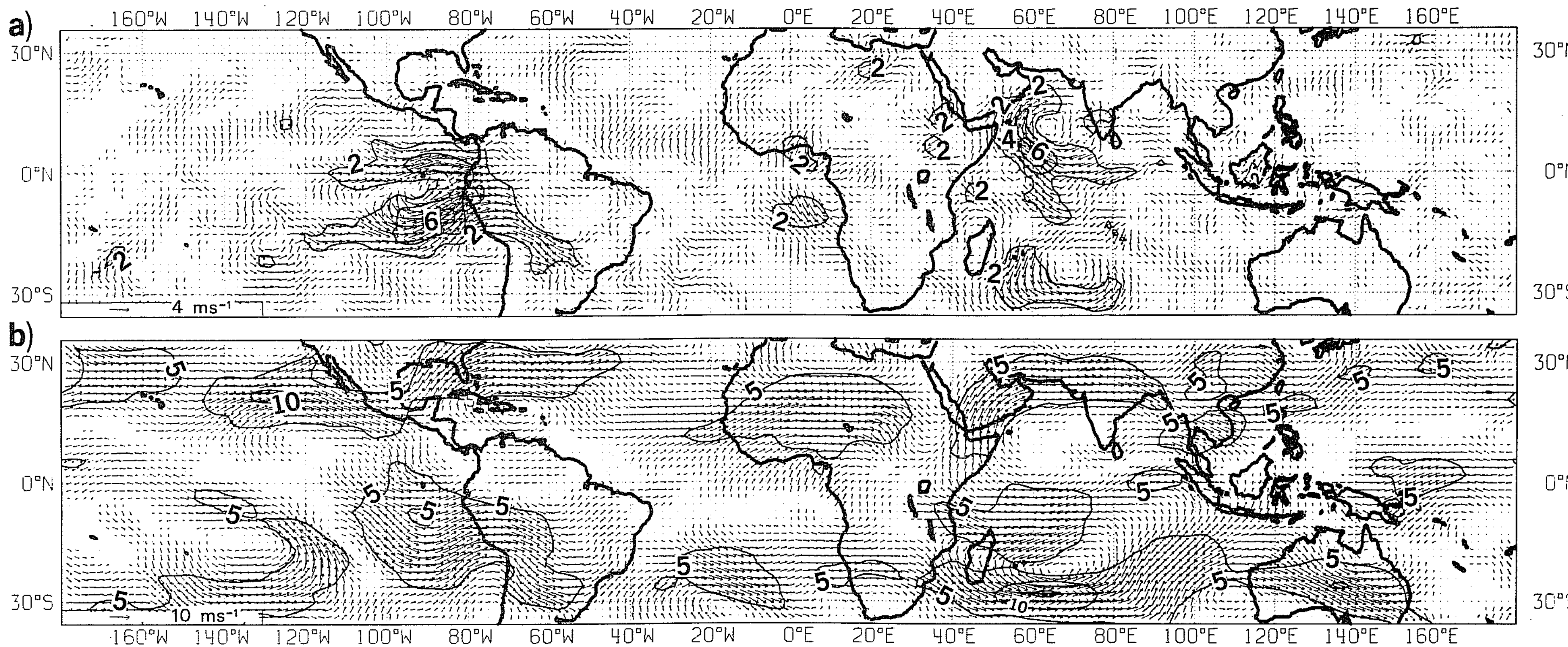


Fig.9 Vector wind difference at 200hPa: new minus old model. Ensemble forecasts and analyses. (a) Analyses, contour interval: $2ms^{-1}$; (b) 120-hour forecasts, contour interval: $5ms^{-1}$. (From Tiedtke et al., 1988)

between the analyzed state and the model's parametrizations. Krishnamurti et al. (1988) have shown that it can be eliminated through adjusting the initial state through a form of nudging. Gradual improvements in the model parametrizations seem to be progressively reducing the severity of the effect. But the general problem of ensuring a consistency between the initial state and the physics of the forecast model is a very difficult one. A clean solution may perhaps only be found through variational assimilation (section 6), but other techniques such as 'physical initialization' (section 5) have been tried, and found to be useful, at least in a research context.

2.4 Lack of mass-wind coupling

In middle and high latitudes a general state of near balance (geostrophy) exists between the horizontal wind field and the horizontal gradient of pressure. This relationship allows a *multivariate* analysis of both mass and wind, given observations of either. Within the analysis this knowledge is incorporated through the definition of correlation functions of the prediction errors relating streamfunction and geopotential (see for example, Lorenc, 1981).

As the equator is approached the coupling becomes weaker and horizontal variations of pressure and temperature become progressively smaller. This lack of a simple relationship between pressure and wind fields in the tropics is one of the reasons why techniques in objective analysis of the tropical atmosphere lagged behind those used for the extratropics. The problem is addressed in current systems either by the use of multivariate schemes which are decoupled as one approaches the equator (equatorward of 30° at ECMWF, e.g. Lorenc, 1981); by univariate schemes which may be combined with a subsequent variational adjustment (e.g. Jones, 1976); or by direct insertion assimilation techniques in which the achievement of an appropriate balance is left, in part at least, to the assimilating model (e.g. Lyne et al., 1982; Lorenc et al., 1991). Variational techniques, discussed in section 6, are currently seen by many to be the way ahead.

Quite apart from the problem of carrying out a multivariate analysis (lack of a simple relationship between mass and wind), the weak gradients of pressure and temperature present problems in that observing techniques frequently lack sufficient accuracy to resolve the gradients. Inaccurate observations may often degrade, rather than enhance, an objective analysis of quantities such as geopotential height.

2.5 Large-scale Analysis

Comparing the fit of the first-guess to observations over a long period of time, and dividing this up in to large-scale and synoptic-scale (total wavenumber ≥ 9) one finds that the 'error' is dominated by synoptic-scales in the northern hemisphere extratropics, and by large-scales within the tropics. An example of this is shown in Fig.10 (Hollingsworth and Lönnberg, 1986).

An obvious large-scale analysis problem is caused by the tendency of the forecast model to rapidly drift towards its own preferred climate. This introduces very large-scale biases into the assimilation (discussed in section 2.3) of which there is insufficient observational data available to fully correct.

It is not possible to distinguish between Rossby and gravity modes on the basis of wind or mass information alone. Cats and Wergen (1983) found that there is a serious aliasing between Rossby and Kelvin modes within the analysis. Solving this problem would require high quality mass data, which, as discussed in 2.4, is not available.

Other investigations by Cats and Wergen (1983) indicated a general difficulty within the tropics in resolving large-scale structures not present in the first guess field, even when a comprehensive set of observational data (synthesized) were available to the analysis. Their experiments consisted of both idealized and realistic observation networks, which were used in the analysis

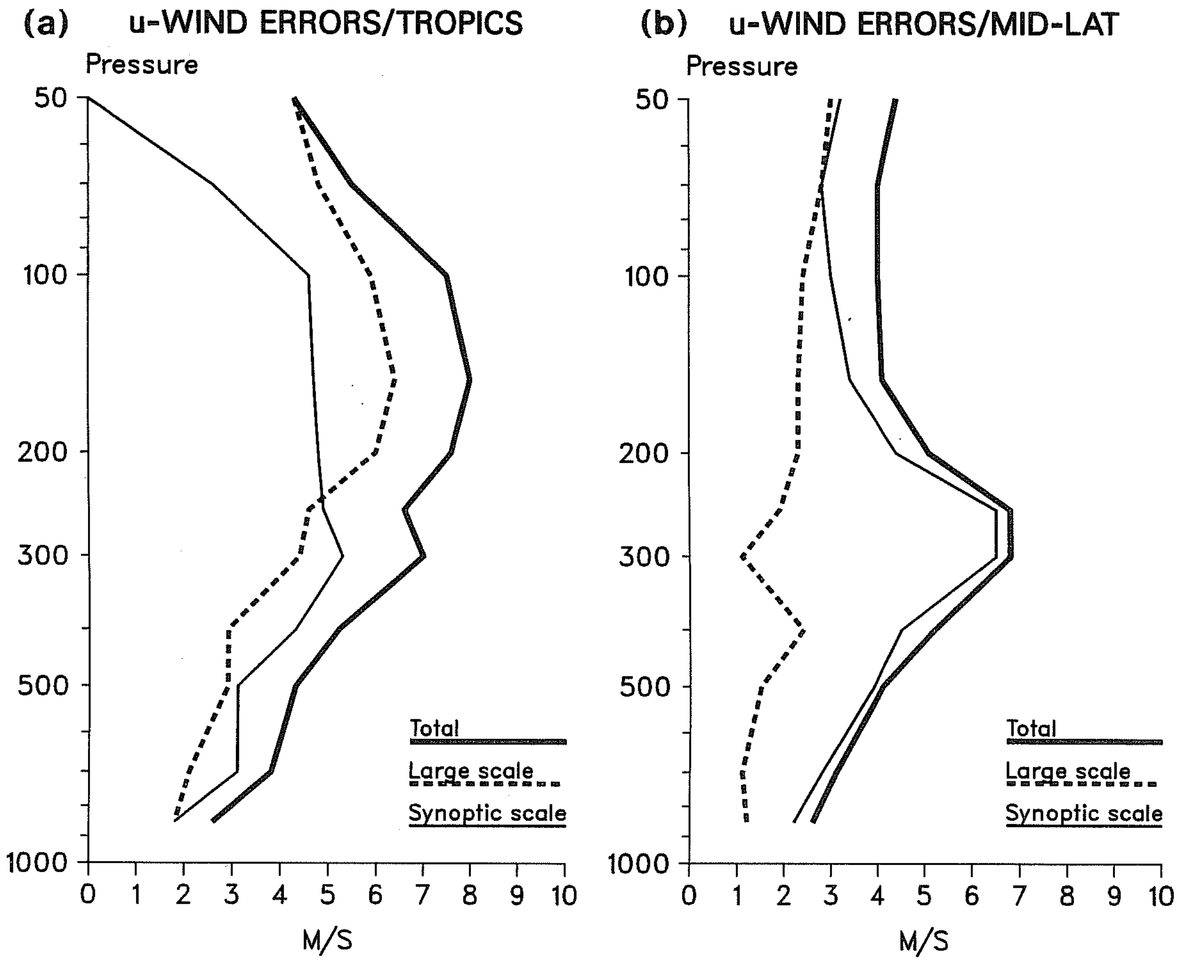


Fig.10 Total error of the first-guess forecasts of zonal wind, as measured against radiosonde observations (heavy solid line); and the contributions by the large-scale (n<9, solid line), and synoptic-scale (n ≥9, dashed line); for (a) the tropical belt, and (b) the northern hemisphere extratropics.
(10b from Hollingsworth and Lönnberg, 1986).

of fields for which the first-guess had had imposed upon it prescribed errors - specifically a multiple of a selected normal mode. The 'truth' to which the analysis should advance was known, being the analysis from a first-guess free of the prescribed error. In this way they were able to measure the response of the analysis to various scales of error, using different observational networks. With the idealized data coverage a serious source of error stemmed from aliasing, both in the horizontal and in the vertical, which occurs through the interaction of the network density and the scale and form of the prediction error structure functions. With the operational network of observations the analysis error was substantially greater than for the idealised network. These problems stem from the inherent local character of the optimum interpolation (OI) technique. A possible solution would be to perform a multi-pass analysis in which the scale of the structure functions change on each pass (perhaps starting at large scales and reducing). In practice, at ECMWF, a (small) constant term is added to the functions to allow data an influence on the largest scales. Whilst, probably not the ideal solution, it does result in an improvement (Lönnberg, pers comm).

Again, variational analysis may help to solve the aliasing problems, but it may not help with the model bias problem. Indeed, Wergen (1990) suggests that the latter could be a serious problem with variational analysis.

2.6 Analysis of small-scale features

Within the tropics it is desirable to analyze not only the rotational but also the irrotational (divergent) component of motion. The ITCZ is an obvious example of a divergent circulation on the large-scale, but irrotational circulations on smaller scales are also likely to be important - for the instigation of convective activity for example. In earlier formulations of OI, the error in the velocity potential of the first-guess was assumed to be zero, which leads to a purely rotational analysis on the scale of the analysis volume. However, the analysis is carried out locally by gathering data into (slightly overlapping) boxes of about 660km by 660km, and of a vertical depth depending on the amount of data available (typically the whole atmosphere within the tropics). The analysis within each of these box volumes is non-divergent. Divergence is 'analyzed' on scales larger than the box size because a different selection of data is used within each of the boxes. An analysis of $f\underline{v}$, rather than $\underline{v}$, would provide an implicit analysis of the large-scale divergence field, at least in the extra-tropics (Philips, 1963).

Fig.11, from Reed et al. (1988), illustrates a characteristic problem the analysis (using non-divergent structure functions) has in drawing to small-scale divergent features in the observations. Other examples may be seen in Julian (1984). Fig.11a shows the operational analysis, and Fig.11b shows a hand analysis. The latter suggests that quite a reasonable interpretation of the flow is possible, drawing to all the available data. The operational analysis is totally unable to fit many of the observations. By contrast the operational analysis of a sharp, but largely rotational, feature (Fig.12) is well captured by the analysis.

The effective resolution of the analysis is determined not only by the observation density; an upper bound is defined by the model grid, which is T106 with nineteen levels in the vertical at present. The scale of the response to the observations is also determined by the statistical structure functions. The structure functions used at ECMWF were revised by Lönnberg (1989) and the new higher resolution was implemented in July 1988. The new functions have a horizontal scale at the equator of 200km (compared to 1000km previously) decreasing to 400km at 30 degrees north/south.

Daley (1983) developed a general formulation of the OI system which provides for both rotational and irrotational elements within the analysis volume. The prediction error covariances are modelled as: -

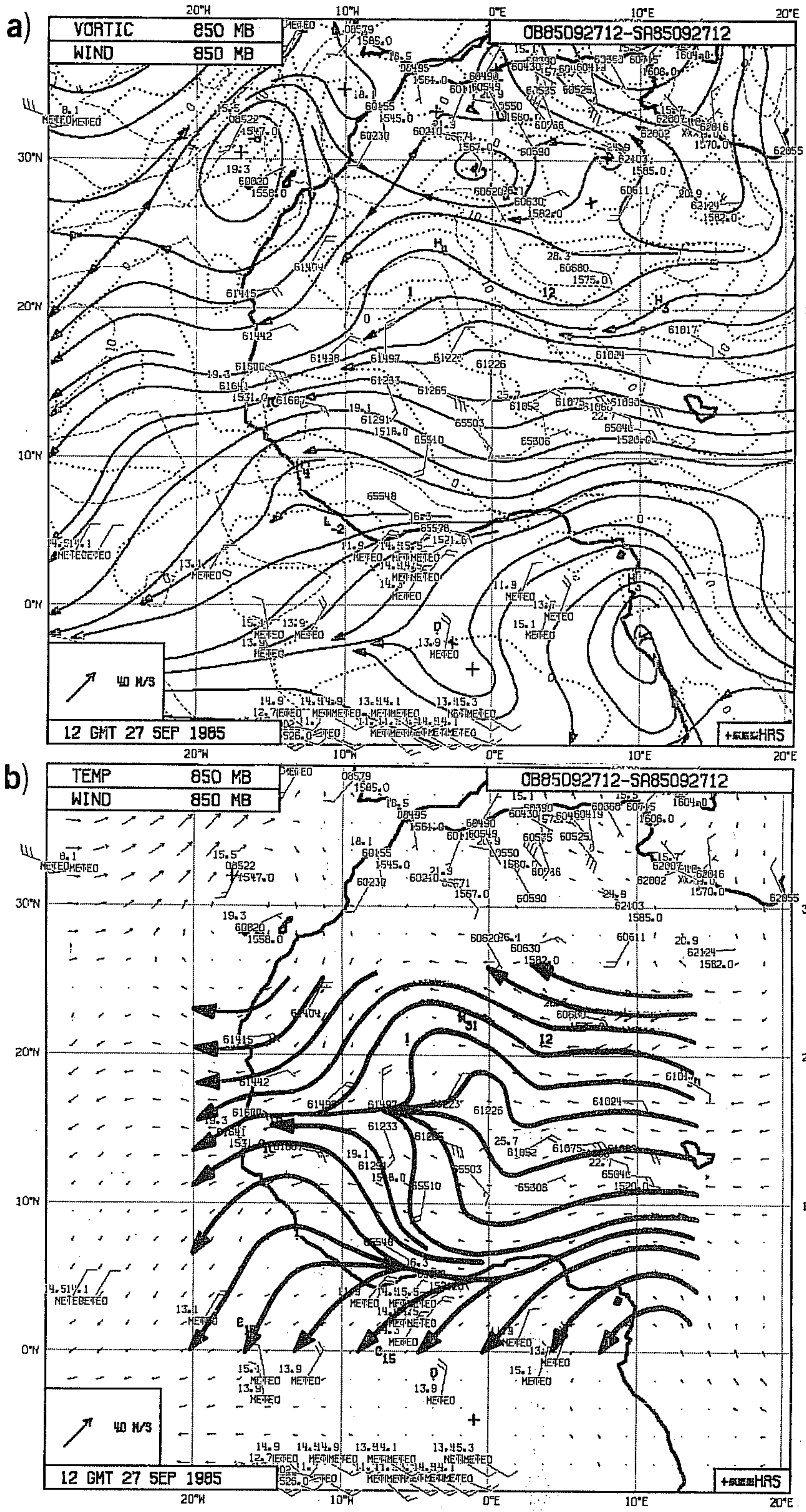


Fig.11 a) 850hPa ECMWF analysis, 1200 UTC 27 September 1985.
Solid lines, streamlines. Dotted lines, vorticity isopleths ($10^{-5}s^{-1}$), dashed lines isotachs (ms^{-1}). H's indicate positions of vorticity maxima and L's positions of vorticity minima. Winds are plotted according to the usual convention: full barb = $5ms^{-1}$.
b) Manual streamline analysis. (From Reed et al., 1988)

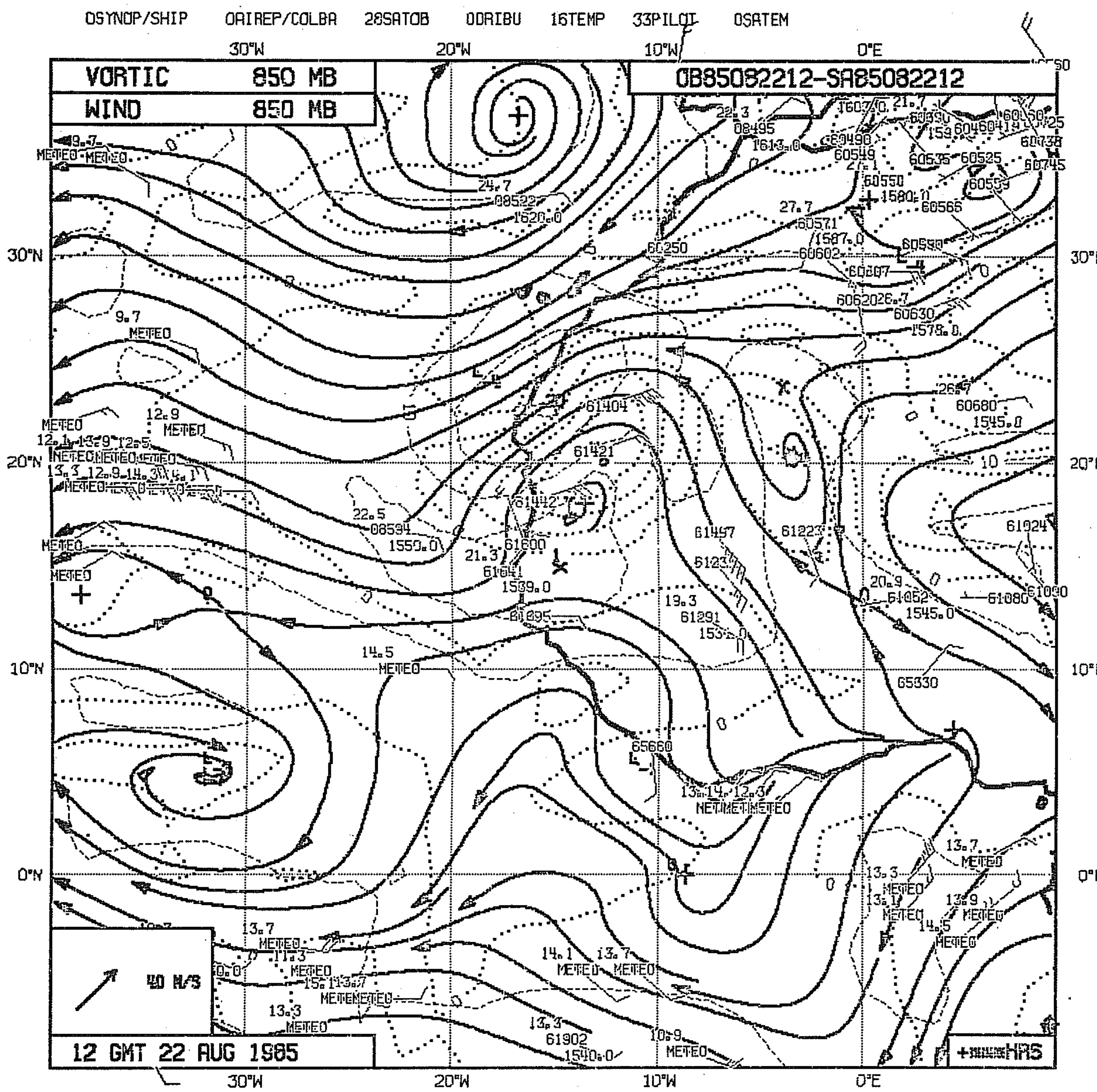


Fig.12 The 850hPa ECMWF analysis 1200 UTC 22 August 1985. For further explanation see Fig 11. (From Reed et al., 1988)

$$<\phi\phi> = E_\phi^2 V_\phi F(r), \qquad <\psi\psi> = E_\psi^2 V_\psi F(r), \qquad <\chi\chi> = E_\chi^2 V_\chi G(r),$$

$$<\phi\psi> = \mu E_\phi E_\psi V_{\phi\psi} F(r), \qquad <\chi\psi> = 0, \qquad <\chi\phi> = 0,$$

where ϕ is geopotential, ψ is streamfunction, χ is velocity potential and μ is a geostrophic coupling constant. The functions $F \equiv F(r)$, $G \equiv G(r)$, are assumed to be homogeneous and isotropic, and are modelled by a series of Bessel Functions. Coefficients of the latter are statistically determined from forecast error studies (see for example, Lönnberg and Hollingsworth, 1986). In practice, $G(r)$ is taken $\equiv F(r)$. The vertical coupling, $V \equiv V(\eta)$, are also empirical functions statistically determined from forecast error studies. $E(x,y,\eta)$: forecast errors are determined from analysis errors at previous analysis time (a product of the OI scheme), empirically enhanced to reflect error growth during first guess forecast.

Horizontal wind covariances are derived from the above by differentiation: -
$\underline{v} = \nabla\chi + \underline{k}\wedge\nabla\psi$.

The above formulation is fairly general and allows for errors in both velocity potential and stream function. One may express the total wind prediction error variance E_v^2 as:

$E_v^2 = E_\psi^2 + E_\chi^2$,

and in general $E_\psi^2 \neq E_\chi^2 \neq 0$.

Suppose $E_\chi^2 = \upsilon E_v^2$,

then $E_\psi^2 = (1 - \upsilon)E_v^2$.

In terms of forecast errors, $\upsilon = 0$ implies non-divergence, $\upsilon = 1$ irrotational flow. It is interesting to note that $\upsilon = 0.5$ implies an analysis univariate in u and v (Daley, 1983). Theoretical arguments (Daley, 1983), empirical investigations (Hollingsworth and Lönnberg, 1986; Lönnberg and Hollingsworth, 1986) and assimilation studies (Undén, 1989) suggest υ=0.1 to be a reasonable value.

This framework allows explicit analysis of divergence within the scale of the analysis volume. Such an approach has been tested at ECMWF by Undén (1989) and was implemented operationally in January 1988. Fig.13 (from Undén ,1989) shows an example of the analysis impact at 200hPa of assuming that ten percent of the error variance is in the irrotational component of the flow (υ=0.1). Quite large changes (5 to 10ms^{-1} being fairly typical) are apparent, and overall fit to the data is generally improved. It is interesting to compare typical analysis increments in the divergent component of the flow (analysis minus first-guess) when rotational (υ=0, Fig.14a), and partly irrotational (υ=0.1, Fig.14b) are used. In the purely rotational case these analysis increments are coming from contributions due to different data selections between neighbouring analysis volumes and occur on scales larger than the analysis volumes (i.e. greater than 660km). In the partly divergent case, the irrotational flow is explicitly analyzed on smaller scales. The most striking impact of the change is that introducing a limited degree of divergence into the structure functions has removed a great deal of 'noise' from the increments. Divergence increments on the larger scales are much the same.

2.7 Analysis of humidity

Humidity is a difficult field to analyze within the tropics, it has large spatial variability both in the vertical and horizontal, and there are very little data available, other than satellite measurements of upper tropospheric humidity (which are not currently used at ECMWF). A consequence of the latter that the analysis is strongly influenced by the forecast model through

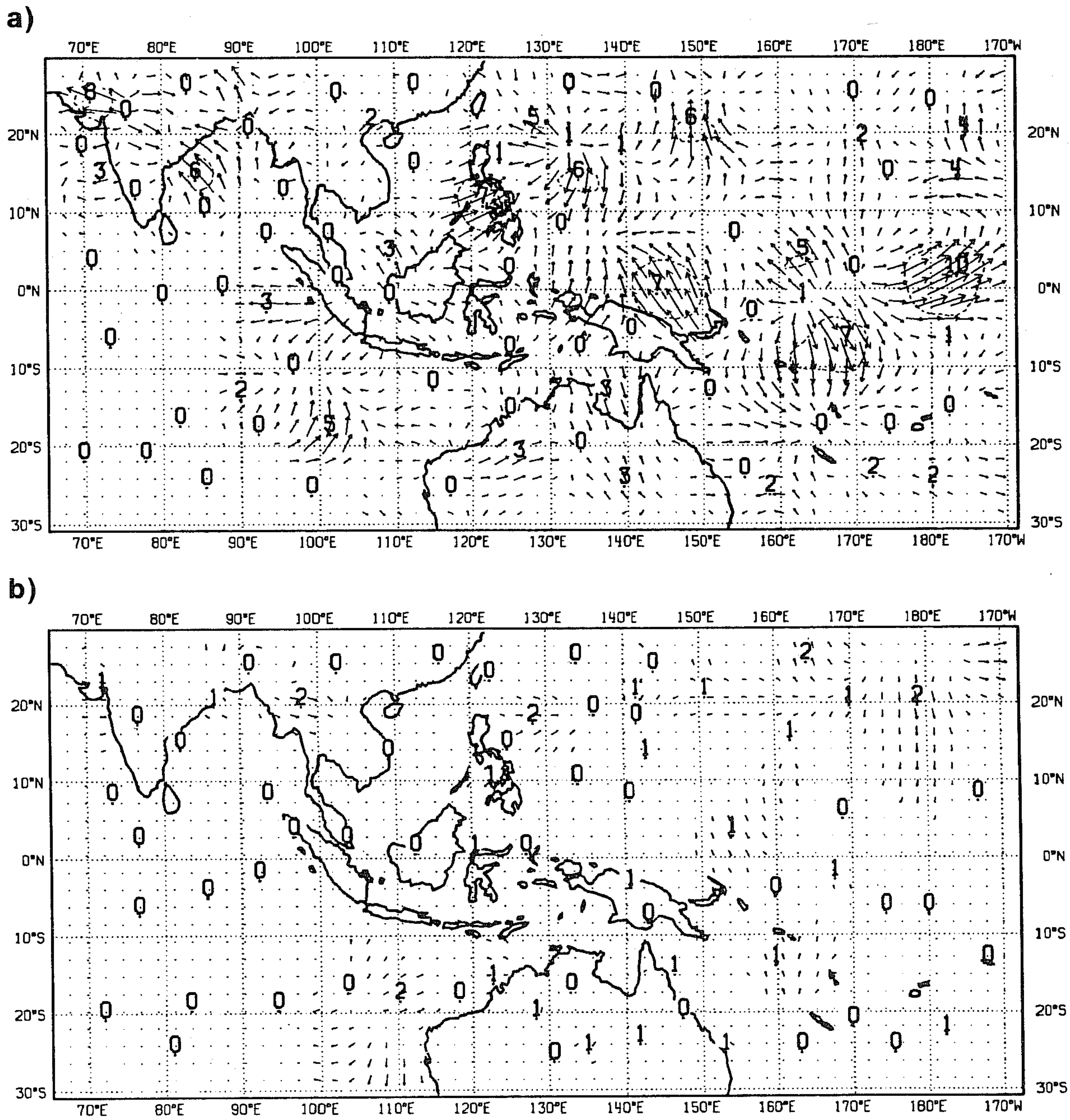


Fig.13 Wind differences at 0000 UTC 8 December 1986 between analysis with divergent structure functions and analysis with non-divergent structure functions at (a) 200hPa and (b) 850hPa. Isotachs for every 5ms^{-1} and numbers indicate maxima (and minima) in ms^{-1}. (From Undén, 1989).

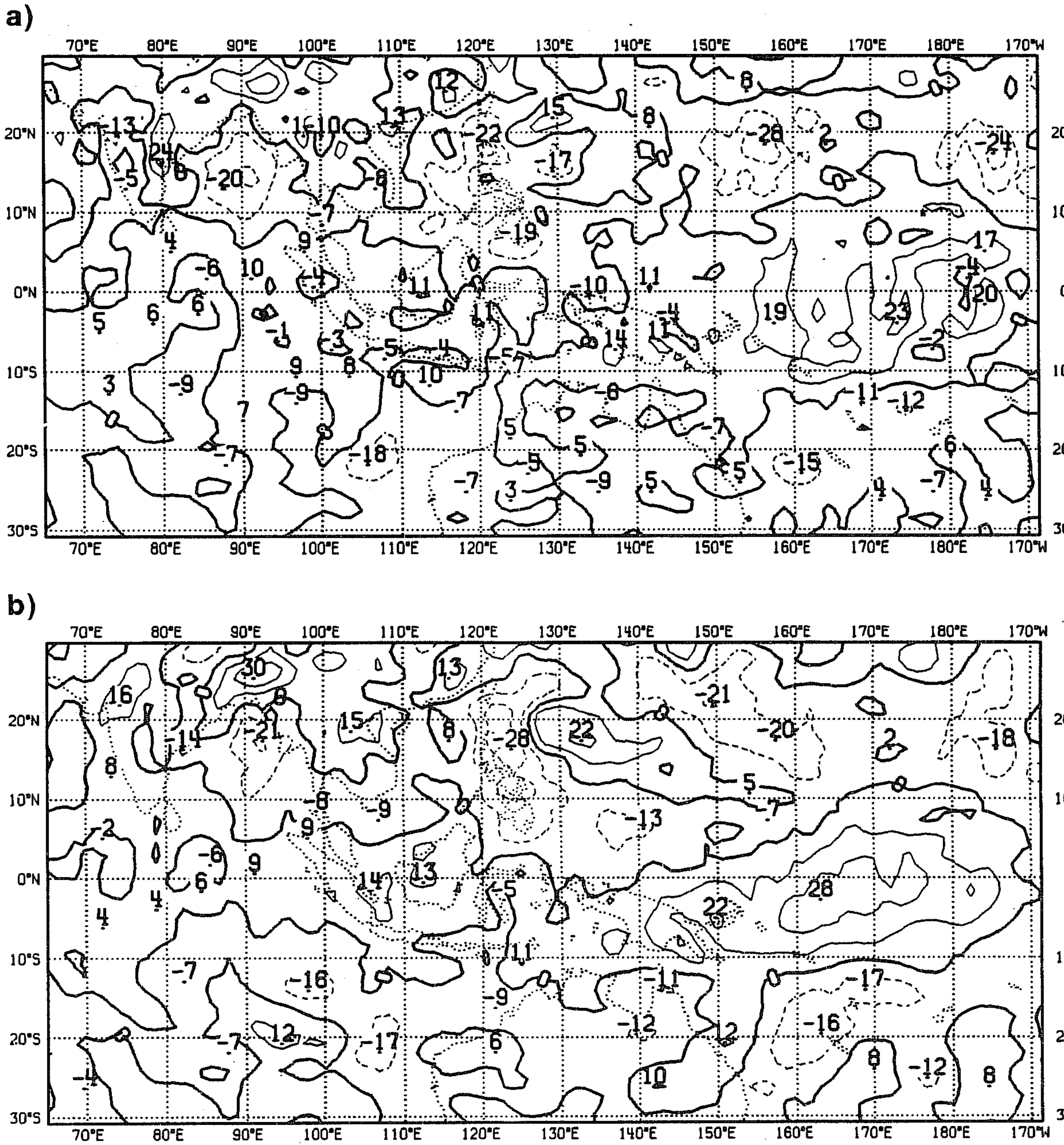


Fig.14 Analysis increments of divergence at 200hPa 0000 UTC 8 December 1986 when analyzing with (a) the non-divergent structure functions and (b) the divergent ones. Unit is $10^{-6}ms^{-1}$ and isolines are for every $10^{-5}s^{-1}$. (From Undén, 1989)

the first-guess, as discussed in section 2.3. However, as illustrative of problems that can arise, it is interesting to consider how the analysis of other variables can influence the initial moisture field.

At ECMWF the analyzed variable is relative humidity. In the absence of observations of moisture (often the case) the relative humidity of the analysis will be that of the first-guess, as seen in section 2.3. Specific humidity depends on relative humidity and temperature. If the analysis changes the temperature but not the relative humidity, the specific humidity is changed. Suppose that the temperature observations have a warm bias, this will increase the temperature in the analyzed state - the specific humidity will be increased. The rate at which evaporation from the surface takes place within the model depends on the difference between the low level specific humidity in the model (initially the analysis) and the saturation specific humidity of air in contact with the surface (e.g. with the sea surface, the temperature - which is analyzed independently). Thus the effect of a warm temperature bias is to reduce this difference in specific humidity - and therefore to reduce the initial evaporation in the model - directly causing a *spin-up* problem in the model evaporation. Such an effect has been found by Illari (1990) in data assimilations at ECMWF, where NOAA-10 MSU sounding data were found to have a warm bias over the subtropical oceans.

2.8 Initialization

As discussed earlier, initialization has the role of controlling noise in the data assimilation. It was found very early on in NWP that very small imbalances, nearly always present in the initial data, would give rise to spuriously large rapid oscillations (mainly in the surface-pressure field and divergent wind). These oscillations died out over a period of days, as the various dissipative mechanisms (present within all NWP models) took their effect. The forecasts beyond this time are unaffected by this initial disturbance, but during the first few hours the oscillations are particularly severe (amplitudes of several mb's are quite common). The six-hour forecast used to provide the first-guess is thus severely contaminated. Its use in quality control of observations is suspect and erroneous decisions are made. A form of non-linear normal-mode initialization (NLNMI) was found to be totally successful in controlling this noise (Machenhauer, 1977; Baer, 1977). But it was found also that if physics were included in the initialization the procedure did not converge; all early NLNMI schemes were therefore adiabatic. Wergen (1988) overcame the convergence problem of the Machenhauer scheme by obtaining an estimate of the *slow* physical forcing, and held this as a constant forcing during the iterations. Wergen's technique proved very successful and has been used operationally at ECMWF since September 1982. Using this scheme the large-scale divergence within the analysis is retained by the initialization, these fields now contain very plausible divergent circulations.

The initialization constraint is to set the time tendencies of the gravity mode terms initially to zero, in this way every Rossby mode component of the initial state may be thought of as having a 'balancing' gravity mode component. This is the so called 'slow manifold' described by Leith (1980). In setting their time tendency to zero, the modes are initially held stationary. This is fine for free modes, but for forced modes this is totally inappropriate. The latter will slowly respond to the forcing within the model but will suffer initially from phase errors. Candidates for this readily spring to mind: the diurnal and semi-diurnal atmospheric tides. Gravity waves are essential to the correct description of these. The semi-diurnal pressure tide has a surface pressure amplitude of 1 to 2mb (Chapman and Lindzen, 1969). A pre-requisite for the correct assimilation of such a feature is a diurnal forcing of radiation in the forecast model, this was introduced in May 1984. The appropriateness of the initialization condition was addressed through operational changes in March 1986 (Wergen, 1989). Wergen excludes the diurnal and semi-diurnal tides from the initialization by estimating their amplitudes based on a time history of the last ten days (which is updated every six hours). In this way the tides are not initialised and are free to propagate. This was found to considerable improve the analysis of tidal phenomena.

Modelling studies carried out by Errico (1984), Errico and Williamson (1988) and Errico (1989) suggest that the Machenhauer condition is inappropriate for many of the diabaticaly forced modes. However, a practical alternative approach has yet to be found.

2.9 Surface Analysis

Many modelling studies have indicated the sensitivity of tropical forecasts to boundary conditions particularly, sea surface temperature (SST) (WMO, 1986), soil moisture (e.g., Shukla and Mintz, 1982), albedo (Rowntree et al., 1985). Analysis of the former is in a reasonable state (partly because it is a fairly slowly varying field) usually being based on a combination of remotely sensed data and ship observations (see e.g., Heckley, 1983). Soil moisture analysis is more difficult, often involving a prior precipitation analysis (which, to say the least, is difficult within the tropics), such a scheme has been implemented at ECMWF by Vasilijevic (1989). It is likely that for improvements in the longer term forecasts within the tropics (timescales of perhaps a week onwards) improvements in the analysis of surface conditions will be crucial.

3. PERFORMANCE OF THE ASSIMILATION SCHEME

3.1 Relative roles of forecast, analysis and initialization

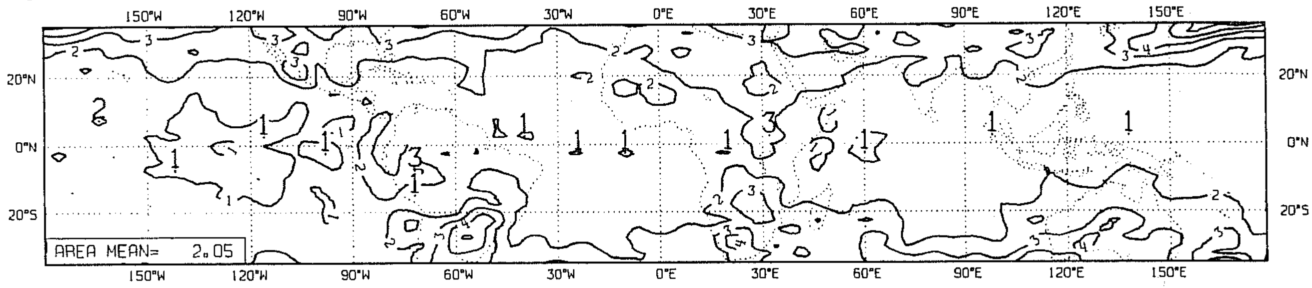
As discussed in section 1, the type of data assimilation currently employed at the ECMWF, is *four dimensional intermittent.* An analysis is carried out every six hours using data over an interval of ±3 hours of the analysis time. The first-guess for each analysis is a six-hour forecast from the previous analysis after initialization. Thus the assimilation contains three main components: forecast, analysis, and initialization. According to Hollingsworth et al. (1986), it is envisaged that within each assimilation 'cycle'(forecast + analysis + initialization), the bulk of the evolution of the atmospheric state will be described by the forecast increment (F); the analysis will make small modifications to this prediction (A), in accordance with the observational data; and the initialization will make even smaller adjustments, in order to accomplish balance (I). The requirement may be expressed as F > A > I. The greater the inequalities, the better the assimilation is working. By and large, in the northern-hemisphere extratropics this condition is well satisfied (Hollingsworth et al., 1986). It is interesting to examine this behaviour in the tropics.

Fig.15 shows the RMS contributions of the three terms to the 850hPa vector wind field. Values are obtained for the first ten days of March 1990, and are shown for the latitude band 35°S to 35°N. Area mean values, $F=2.05ms^{-1}$, $A=1.56ms^{-1}$, $I=0.72ms^{-1}$, suggest that overall the assimilation is working as expected; but much of the F contribution comes from the poleward extremes of the latitude band; that of A from the few areas in which there are data. I is more uniform. Within the deep tropics A is typically the same order as F, and often larger (Indonesia, for example). There are many areas where the I contribution is similar in size to A, or F.
At 200hPa (not shown) the picture is very much the same.

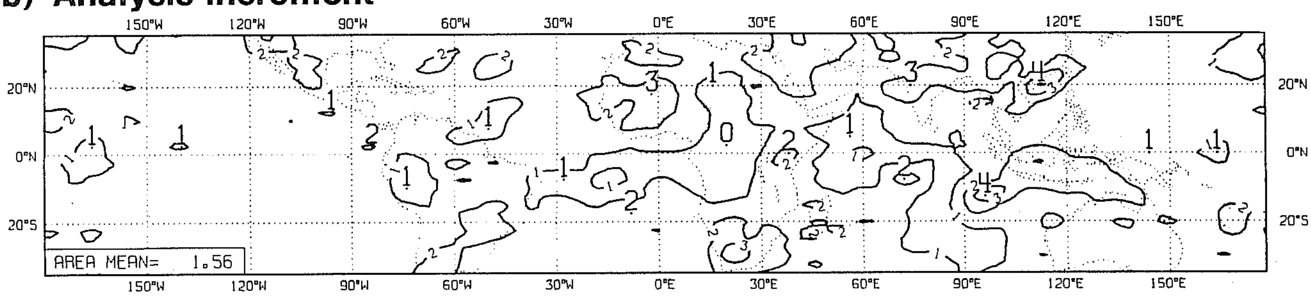
The overall impression one gets is of forecast errors continually advected into a few, sparse, relatively data rich areas where they are continuously being removed by the analysis (e.g. Indonesia, west Africa etc.); elsewhere, where there are relatively little observational data, the analysis increment is small and the fields reflect systematic biases (climate drift) in the model first-guess - since, in these regions, the analyses already biased by the model climate drift the forecast increments also small. The initialization increment is also small, the fact that it is the same order as F and A in some regions, is more a reflected of the fact that the latter are small (hardly any data) rather than I being 'large'.

Within the deep tropics, at least, one may question the efficacy of a model based four-dimensional assimilation such as this. A principal role of the forecast component is to carry

a) Forecast increment



b) Analysis increment



c) Initialization increment

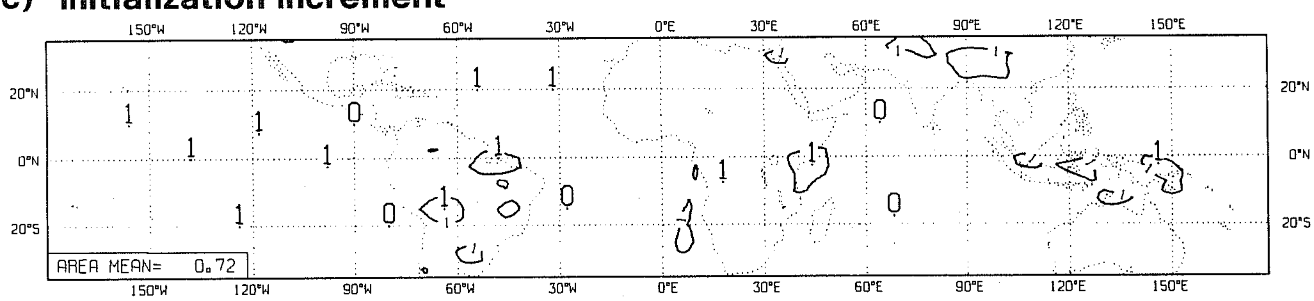


Fig.15 a) mean forecast increment. b) mean analysis increment. c) mean initialization increment.
RMS change to v (ms^{-1}) at 850hPa for first 10 days of March, 1990.

information obtained from observations at the current analysis time forward into the next cycle in the background field. In this way time continuity is introduced and information is advected from data dense to data sparse regions. It is clear that to some extent this must be going on, and we shall see evidence for this in section 3.2. However, because of the large model biases in the tropics, a significant effect, in most regions, is the introduction of these biases into the analysis - where they remain.

3.2 Analysis of Easterly Waves

3.2.1 Evaluation of the analyses

Reed et al. (1988) have evaluated the ability of the forecasting system to analyze and forecast easterly waves, and their related tropical storms, in the African and Atlantic areas, during a two month period during the summer of 1985. Their evaluation was based on examination of analyses at six-hour intervals for the entire period and on forecasts produced once daily. Frequent use was also made of METEOSAT images (European Space Agency, 1985) in carrying out the evaluation. These were found to be particularly useful in oceanic analysis.

They found analyses for the 850hPa level to be of paramount importance in identifying and tracking the waves. For this level radiosonde and pilot winds are plentiful over west Africa, except for a small region over the Sahara. A much larger data gap exists over parts of central and eastern Africa in the region enclosed by 10°S, 20°N, 10°E and 35°E. Within this area, no reliable identification of the waves was possible. Over the Atlantic, data are sparse at 850hPa, consisting of only, sporadic, satellite cloud-track winds. These are relatively plentiful in the vicinity of the equator but not at the latitude of the storm track. West of 60°W, observations from the Caribbean islands and from coastal South America allowed the waves to be easily identified and tracked. Due to various operational reasons, low-level reconnaissance aircraft reports in the vicinity of hurricanes were not available to the analyses.

There are a large number of cloud-tracked winds and aircraft reports available at 200hPa, considerably more data are available at this level than at 850hPa. These data were, however, found to be of little value in analyzing the African waves.

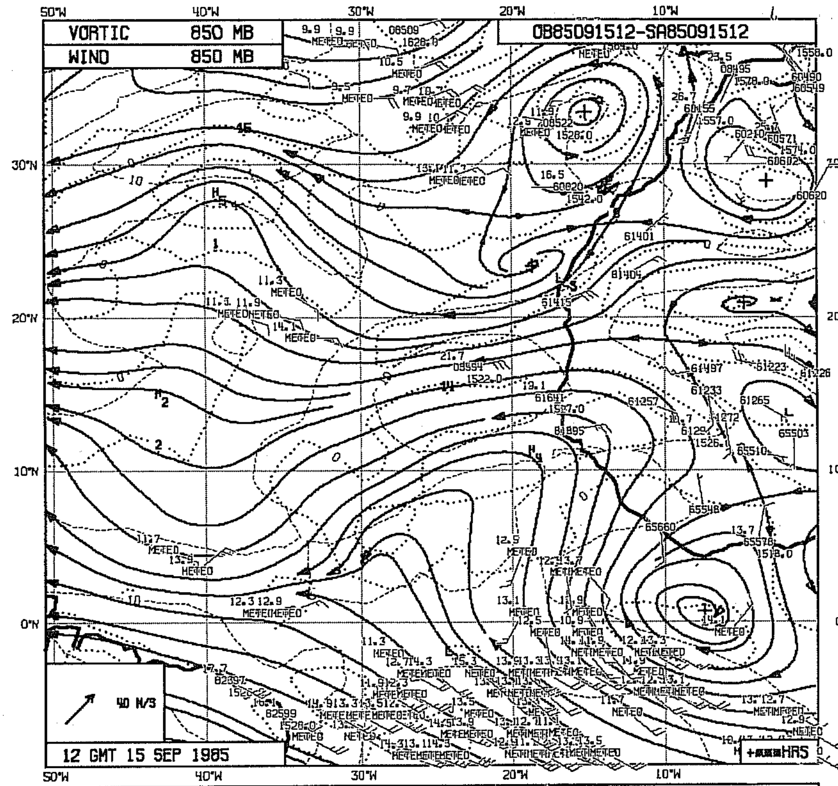
Study of the 850hPa charts revealed the presence of a more or less regular succession of waves over west Africa and the Atlantic. However, even looking at charts at six-hour intervals it was sometimes difficult to determine consistent and unambiguous positions throughout the lifetimes of the disturbances. This problem was particularly acute over data-sparse oceanic regions and stemmed from a tendency of the analyses to develop multiple vorticity centres (but often within a common synoptic scale trough), or to lack a well defined centre in the typical case where the westward propagating vorticity centre weakened in mid-ocean. This uncertainty could usually be resolved by recourse to the 700hPa vorticity to determine the longitudinal position of the wave, and also through use of satellite visible images. Wave histories thus obtained were found to be, for the most part, in good agreement with those prepared by the U.S. National Hurricane Center.

3.2.2 An example - the mid-Atlantic passage of Gloria

A facet of the analysis system that is of considerable interest is its performance in regions of deficient data. An example is chosen to demonstrate the performance of the analysis system in locating easterly waves or vortices where data were lacking, or nearly lacking, over a broad area. This example has been chosen because the analyses can at least partly be verified by independent observations, in this case by satellite observation of cloud pattern.

The operational analyses are shown in Fig.16. They depict at 48-hour intervals the progression of the disturbance and its offspring, Hurricane Gloria, across the Atlantic. The sequence begins with the analyzed 850hPa chart for 1200 UTC 15 September 1985, when the disturbance was

(a)



(b)

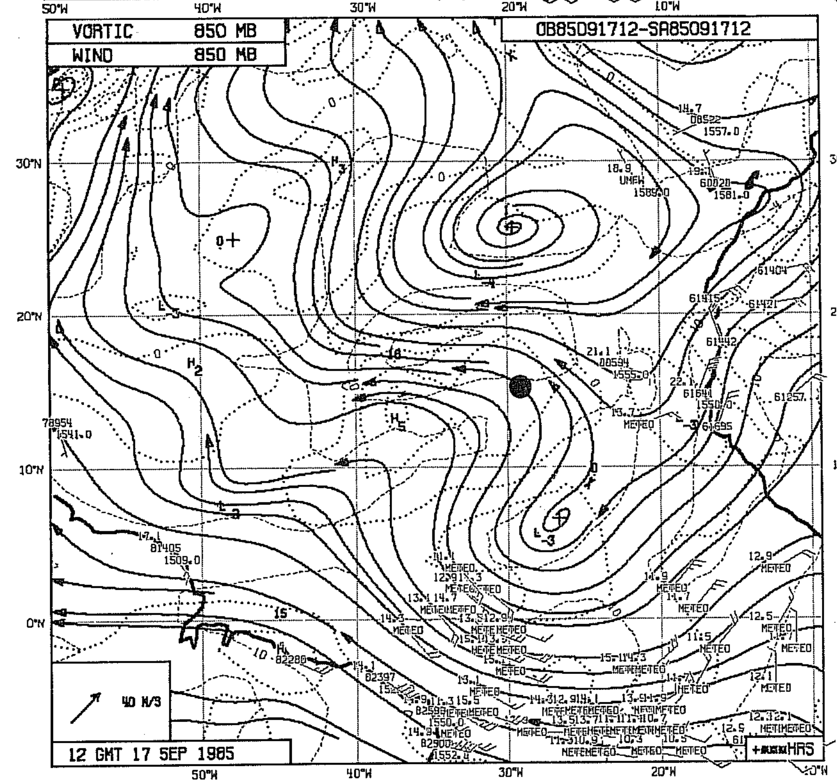
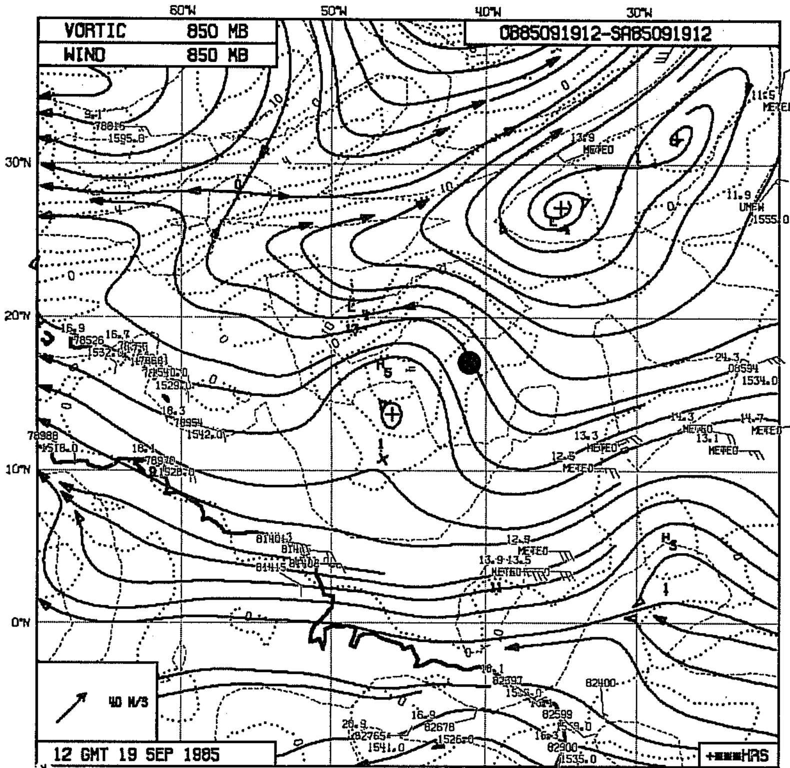
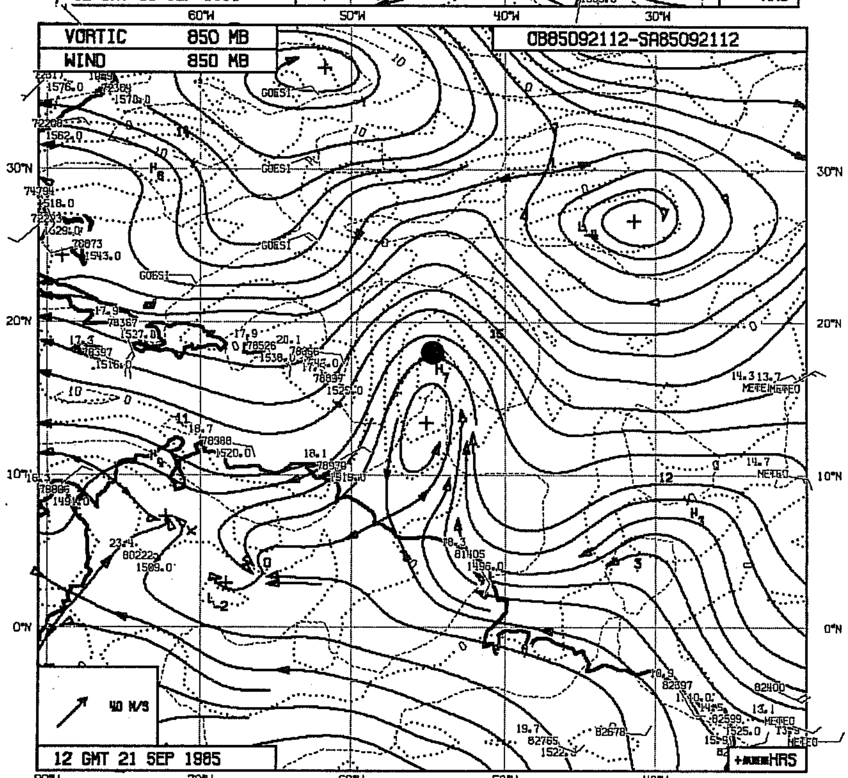


Fig.16 850hPa analysis. For further explanation see Fig 11. Large dot indicates position of wave (later hurricane Gloria), as determined from meteorological satellite observation.
a) 15th September, 1985. b) 17th.

(c)



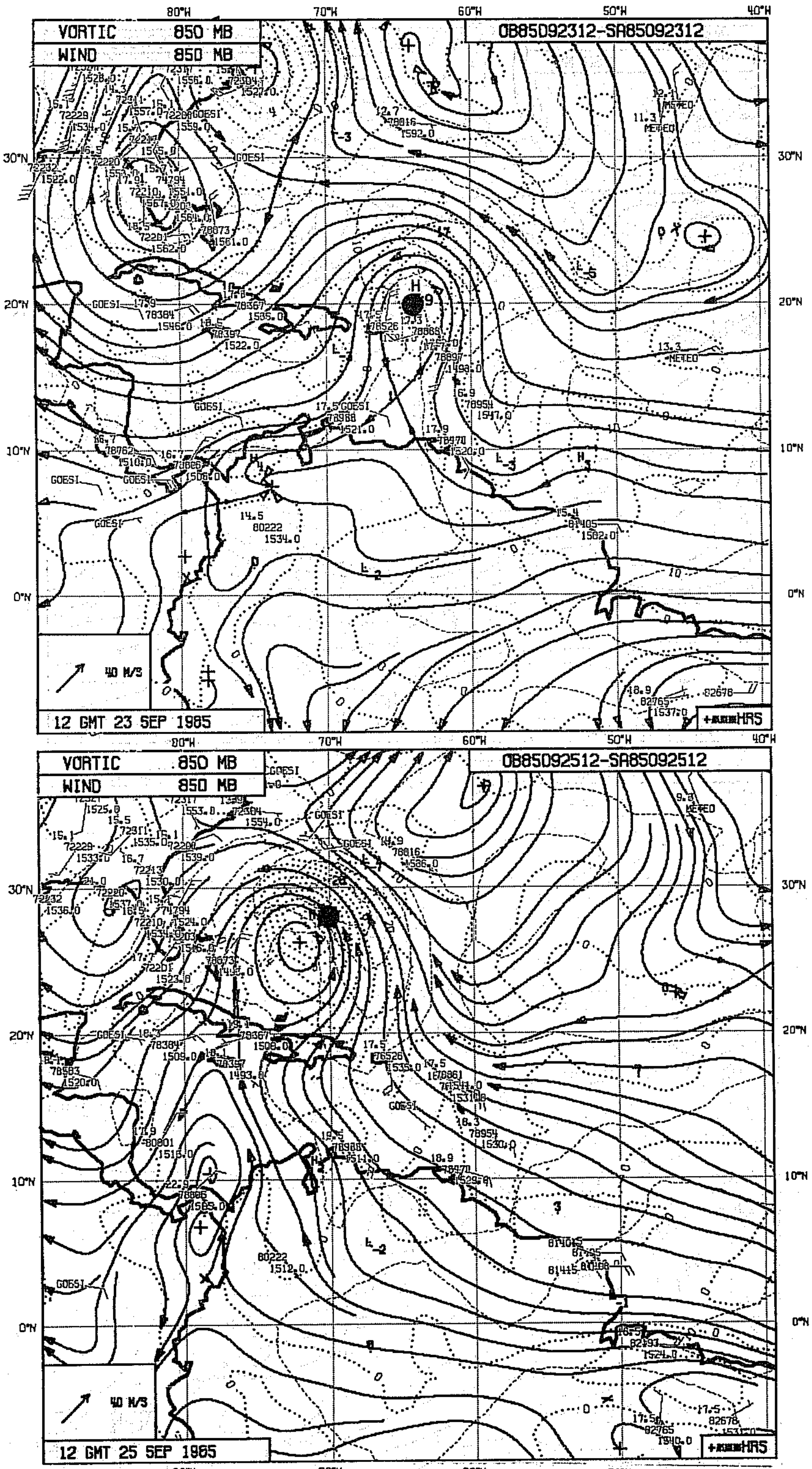
(d)



c) 19th. d) 21st.

(e)

(f)



e) 23rd. f) 25th.

located at 11°N, 19°W, a short distance off the African coast, and ends with the chart for 1200 UTC 25 September, when the storm was located east of Florida and had attained hurricane strength. From the charts for 1200 UTC 17, 19 and 21 September (Figs.16a-c), it is apparent that the analysis system maintained, and even strengthened, the wave during its passage across the Atlantic - despite the virtual absence of data. Intermediate charts (not shown) were marked by a similar deficiency of data. Surface ship reports gave some hint of the disturbance but could hardly have been responsible for the sharply defined system at 850hPa. Cloud tracked-winds were absent at 700hPa and scanty at 200hPa. No aircraft observations were received within this area.

Satellite images (e.g. Figs.17a,b) clearly reveal the presence of an unusually well-organized disturbance in the general vicinity of the analyzed wave. However, from close inspection of the images it is apparent that quite sizeable errors exist in the analyzed positions for 17 and 19 September, if the 850hPa vorticity maximum (denoted by the letter H) is chosen to define the analyzed position. The main maximum is located roughly 1000km west-southwest of the satellite-inferred position on the 17th (Fig.16a) and over 500km to the west of the observed position on the nineteenth (Fig.16b). On the 17th, a secondary maximum is evident within the same trough at a position closer to the satellite position.

The evolution of these analysis errors can be traced with the help of the available 6-hour time continuity. This reveals that the primary vorticity maximum near 13°N, 38°W on the chart for the seventeenth developed suddenly between 0000 and 1200 UTC 17 September, as an offshoot of the somewhat larger but weaker region of vorticity to the east. The latter region, in turn, represented a new development to the rear or upstream of the maximum seen on the beginning chart (1200 UTC 15 September). The original maximum moved rapidly west-southwestward and disappeared between 0000 and 1200 UTC on 16 September.

From this description it is evident that during the period, 15-19 September, the analyses of Gloria were more successful in depicting the synoptic scale wave trough than the position of the disturbance itself. However, by 1200 UTC 21 September (Fig.16c) the analyzed position and the position obtained from reconnaissance aircraft were in close agreement, though the disturbance, now a tropical storm, had not yet emerged from the data sparse region. The correction occurred gradually as the analyzed storm moved systematically slower than the observed. Once in agreement, the positions remained more or less so.

3.3 Analysis uncertainty

Fig.18 shows a comparison between two totally independent, operational, data assimilation systems, for the 850hPa wind; a) that of the ECMWF, and b) that of The U.K. Meteorological office (Ruml, Simmons an Vasilijevic, 1986). The difference is shown in Fig.18c, and the observations in (d). Similar observations were available to both systems. It is clear that over much of the Indian Ocean, differences between the two analyses are of the same order as the winds themselves. Overall, the picture is similar to that of the impact of the forecast model, discussed in section 2.3. Table 1 gives the RMS differences by region, and describes the fit of the two analyses to the available radiosonde data. Both systems fit the data to similar extent. In terms of differences between analyses, these are largest over the Indian Ocean, which, as we have seen many times already, is almost totally void of data.

A more recent comparison of the two analysis systems, for 1990, indicates similarly large differences in these data sparse regions (Arpe, pers comm).

4. USE OF "ALTERNATIVE" DATA

4.1 Bogussing of tropical storms

Tropical storms are often badly represented, even in their mature stage, by operational data

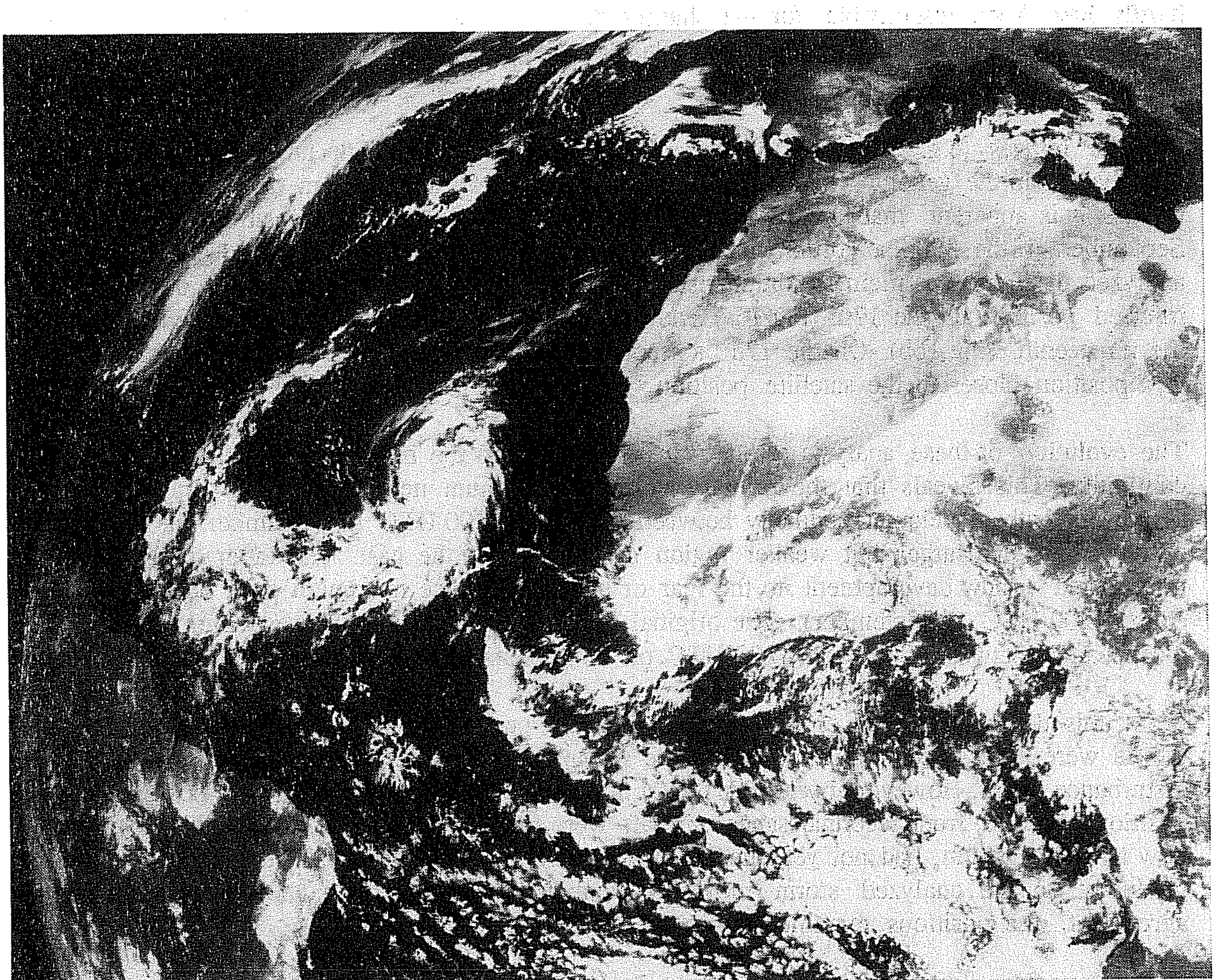
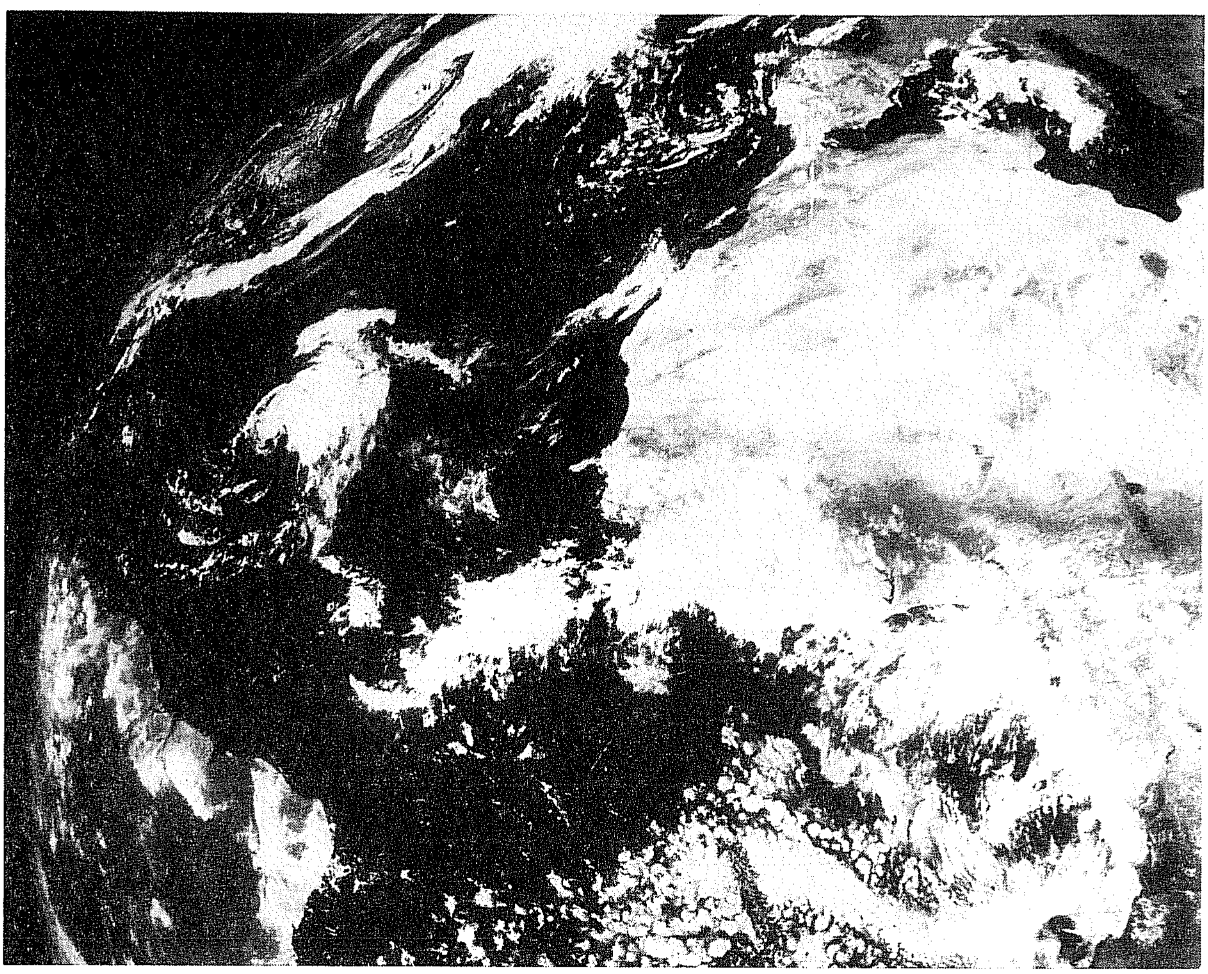


Fig.17 Meteosat visible image of Gloria.
a) 1155 UTC 17th September, 1985.



b) 1155 UTC 19th September.

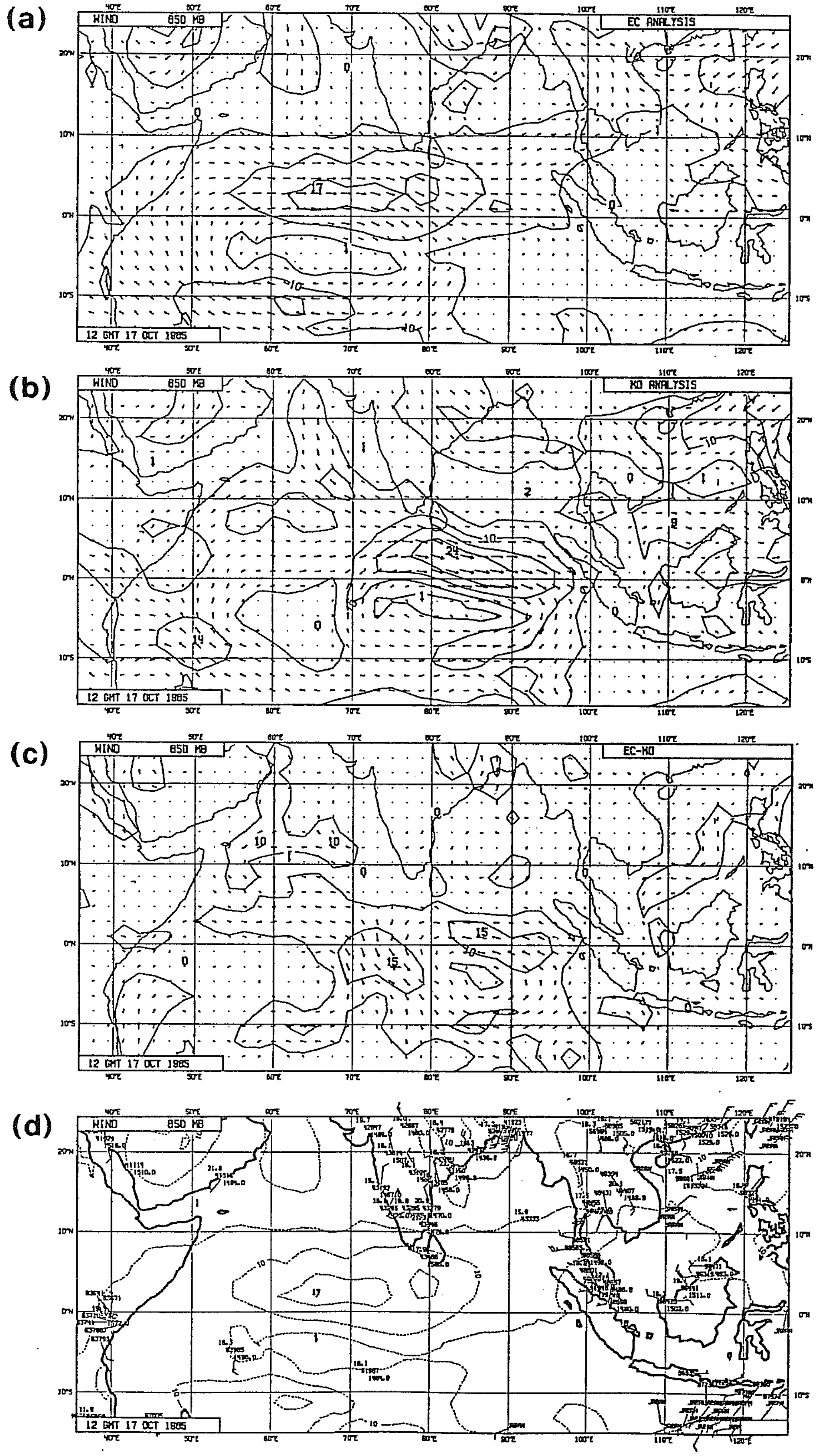


Fig.18 Analyses of 850hPa wind over the Indian Ocean for the 17th October, 1985, produced by a) ECMWF, and b) the UK Meteorological Office. The difference between the two is shown in c). d) shows the 850hPa observations received over the GTS and archived at ECMWF. Isotachs of the ECMWF analyses are also shown on the bottom plot. The contour interval is $5 ms^{-1}$.

REGION	LEVEL	RMS WIND DIFFERENCE
TROPICAL	850hPa	3.4 ms^{-1}
ATLANTIC	200hPa	6.1ms^{-1}
TROPICAL	850hPa	3.7ms^{-1}
PACIFIC	200hPa	6.8ms^{-1}
INDIAN	850hPa	4.6ms^{-1}
OCEAN	200hPa	8.4ms^{-1}

Table 1. Root-mean-square differences between analyses from ECMWF and the UK Meteorological Office, computed over three tropical oceanic regions, based on a set of five analyses. (From Ruml et al., 1986)

	Radiosondeat 850hPa	Radiosondeat 200hPa	Low-level cloud-track winds	Aireps at 200hPa
ECMWF	2.5ms^{-1}	3.6ms^{-1}	2.0ms^{-1}	4.0ms^{-1}
UKMO	2.5ms^{-1}	4.4ms^{-1}	2.1ms^{-1}	4.0ms^{-1}

Table 2. Root-mean-square fits to radiosonde observations at 850hPa and 200hPa, to low-level cloud-tracked winds and aircraft reports at 200hPa, of wind analyses from ECMWF and the UK Meteorological Office, computed over the tropics, and based on a set of five analyses. (From Ruml et al., 1986)

assimilation systems. The observation network is in many areas not dense enough to capture these systems. Over much of the oceans, where the storms develop, they can be missed completely by the conventional observing network. The resolution of the analysis itself is also a limiting factor. The statistical structure functions used in the analysis at ECMWF do not acknowledge the existence of such small-scale features as tropical cyclones; the analysis can only represent the outer environment of a storm. Hence it rejects data on typhoon cores. That is one reason why the analyzed tropical cyclones often appear too large and weak. But, as shown by Reed et al. (1988), a good track forecast for a hurricane is possible if the initial position is accurate.

Satellite visible imaging can provide information not provided by other observations. In manual operational forecasting they provide the most useful information for pin-pointing the location of tropical cyclones and for estimating their intensity and size (Dvorak, 1984). This information is however not available on the global telecommunications system in a standardized form. It is distributed as a worded message - the *Tropical Cyclone Bulletin.*

The Japanese Meteorological Agency (JMA) have incorporated this type of additional information in their global analysis system (Kanamitsu, 1985). From three input parameters (central position; central pressure; and diameter), idealized typhoon sounding data are created based on the climatology of typhoons - so called typhoon bogus observations. To represent a large typhoon JMA use 19 soundings, all within 450km of the centre of the storm.

There have been several attempts to use bogus observations in numerical weather prediction, especially for initialization of high resolution limited area models used for tropical cyclone forecasting. Knowing that the storm track is influenced by the outer storm environment the usual approach has been to superimpose an idealized vortex on a large-scale analysis followed by a balancing initialization procedure.

At the US Naval Environmental prediction Research Facility (NEPRF) Hodur used the full model equations on a very fine grid to spin up a set of tropical cyclones of different intensities and latitudes from idealized initial states. The set of tropical cyclones were then stored and used as a data set from which bogus observations were extracted and blended with a background field before entering the analysis.

A different approach has been taken in experiments at the National Meterological Centre Washington (NMC), where Mathur (1986) proposed a scheme to use bogus observations in the high resolution forecast model. From a prescribed pressure field, controlled by three input parameters (central pressure, maximum pressure, and size of the storm), wind fields are calculated in gradient-wind balance and with an ad-hoc vertical shear. Potential temperature bogus data are then found from the hydrostatic equation. The bogus data are entered in the NMC's OI analysis on a 1.1 degree grid. Integrations have been carried out on a grid spacing of 40km. The intensity and the track were found to be well predicted. Marks (1986), also at NMC, has used a simple symmetric vortex to form the so called Pocket Hurricane Model used to generate bogus vortices for spinning up the NMC Moveable Fine Mesh Model (MFM). The bogus wind fields were blended with the MFM fields. In order to support the wind increments, geostrophic increments were then added to the MFM height fields. The reported result was a substantial improvement in the 12 hour predicted motion, at the expense of degraded forecasts in the 36 to 72 hour range.

The use of bogus data has long been advocated by the UK Met Office. Their technique is based on manual intervention and it has proven useful in supporting the analysis of tropical cyclones (Hall, 1987). Its routine use has led to an improvement of the forecast movement. Normally the forecaster provides four wind observations which are inserted no less than 300km from the centre, at each level below 500hPa with equal strength.

Bogus observations have also been used in research to investigate the role of vortex structure on tropical cyclone motion. Fiorino and Elsberry (1986) used a barotropic model to show that the motion was largely governed by the large scale flow surrounding the hurricane. The strength of the cyclone outside the 15 ms^{-1} radius proved to be more important than the winds near the centre of the cyclone. deMaria (1986) found that by changing the distribution of winds greater than 100km from the centre, large differences in predicted cyclone tracks and speeds were obtained.

At ECMWF bogusing of tropical storms has been tried in research mode by Andersson and Hollingsworth (1988). Bogus observations are generated by a idealized tropical cyclone model which consists of two parts, which are superimposed: an idealized symmetric vortex, and a background field. The symmetric wind field at all levels is modelled by a *Rankine Vortex* as described by Milne-Thompson (1968) and used by Holland (1983) and others. For the background field the first-guess field is truncated to T20, which is assumed to be so smooth so as to contain no information on the cyclone itself, only on its environment. The bogus observations are added as follows: geopotential at 1000hPa, winds at 850hPa, 700hPa, 500hPa, 400hPa and 300hPa. The winds are reduced in speed in height attaining at 300hPa 35% of their value at 850hPa, but direction is unchanged. Twenty nine bogus observations are introduced within a radius of 600km of the storm centre. After the analysis, the initialization is found to draw fairly well to the bogus wind observations and at the same time introducing a balancing pressure distribution. In general the short range forecasts (up to one or two days) are improved, but beyond that impact is negligible. In these later stages the evolution seems to be dominated by the model's physics; experiments by Heckley et al. (1987) indicate the kind of extreme sensitivity that can occur in this respect.

4.2 Use of diagnosed heating rates within the initialization

Satellite radiance data are 'conventionally' used to provide data on the thermal state of the atmosphere through empirically derived temperature profiles. Such information is then used directly in the analysis scheme. Such data have far more potential than this rather limited usage (see e.g., Heckley et al., 1991). There exists a strong relationship, within the deep tropics, and to some extent the sub-tropics, between cold clouds and precipitation. This relation has been used by many authors to estimate precipitation rates in these regions (e.g. see review by Barrett and Martin, 1981). On the assumption that deep convection is associated with low-level inflow and upper-level outflow, it has also been used to infer divergence fields in these regions (e.g. Julian, 1984; Krishnamurti and Low-Nam, 1986; Kasahara et al., 1988).

Such techniques offer, presently, virtually the only hope for obtaining precipitation estimates over most of the tropics and sub-tropics. Satellites carrying active microwave instruments such as SSM/I promise more accurate precipitation measurements for the future, see section 7.

In view of the dominant role of diabatic processes within the tropics, it makes sense to use estimates of the diabatic forcing itself within the data assimilation. The principal contributor to which is precipitation, which, as mentioned above, may be estimated from OLR. Another motive is that the ECMWF model, like with many others, suffers from spin-up (Heckley, 1985) in which during the first few hours of a forecast the predicted values of condensed water and, therefore of latent heating, are rapidly varying, particularly in the tropics. Klinker (1990) demonstrates that use of model generated heating within the initialization leads to significant errors in the initial momentum tendencies within the tropical regions.

4.2.1 Precipitation Analysis

One would like precipitation estimates covering all of the tropics and sub-tropics every six hours.

However, there are significant gaps in the coverage by geostationary satellites. For this reason Heckley et al. (1990) opted to use data from polar orbiting platforms.

The NOAA series of orbiting satellites carry, as part of their instrument package, a High resolution Infrared Radiation Sounder (HIRS) sampling 20 channels from .7 to 15-micron. Each measurement resolves a circular area of 30km diameter at the satellite sub-point. At any one analysis time between five and ten orbits are available from a combination of two available polar-orbiting NOAA space craft. Typically, coverage is such that at each of the main synoptic hours there are few data voids.

Regressions are used by Heckley et al. (1990) to estimate OLR at each HIRS spot. Values of OLR are converted to brightness temperature. Observations are collected into boxes of 7*7 spots, and the fractional coverage of cold cloud is calculated within each box by counting the number of spots colder than 235K. The result is an estimate of the fractional coverage of cold cloud at a reduced resolution of about 2.5 degrees. The value 235K is the optimum threshold for estimating convective rainfall within 2.5 degree boxes (Arkin, 1979).

A Cressman analysis is then carried out of both the fractional coverage of cold cloud and of the observation times. A by product of this analysis is a field of weights which can be regarded as a point by point measure of the data density. These are used in order to provide confidence limits for the analyses. The analysis of observation times is used to modify this field of confidence weights in such a way that the weights are progressively reduced according to the degree of asynopticity of the data. Finally, the weights are subjected to a latitudinal weighting such that they become zero poleward of 35 degrees and can only reach their maximum value equatorward of 25 degrees. Thus the field of weights now reflects the data distribution, as well as the asynopticity of the observations and latitude.

Arkin and Meisner (1987) define a GOES Precipitation Index (GPI) as the product of the mean fractional coverage of cloud colder than 235K in a 2.5 x 2.5-degree box (F_c), the length of the averaging period in hours and a constant of 3mm/hour. Following the work of Arkin and Meisner (1987), the precipitation **rate** is defined by Heckley et al. (1990) as being $3F_c$.

4.2.2 Use within the assimilation

As discussed in section 2.8, a diabatic non-linear normal-mode initialization scheme is employed. The diabatic forcing is obtained from an integration of the forecast model from the uninitialized analyses. The physical forcing is time averaged over two hours of integration and projected onto the normal-modes. This forcing is then filtered by retaining only the large-scale component of the projection onto the 'slow' gravity modes. Slow in this case means modes whose free period is five hours or larger. Because of the nature of the gravity modes, the frequency constraint also acts as a strong spatial filter on the forcing. The physical forcing thus calculated will be under estimated within the tropics due to the dominance of convective heating in those regions, and the spin-up of the model's convective heating Heckley (1985), which (in the version of the ECMWF model used in the Heckley et al. (1990) study) is, characteristically, initially too weak. However, inclusion of diabatic processes in this way considerably improves analysis of the tropical divergent wind field, in the tropics and sub-tropics, over that obtained with an adiabatic initialization; and should also reduce the initial momentum tendency errors (Klinker, 1990).

Rainfall rate gives only a measure of the integrated heating rate; the vertical heating profile must be deduced in some way. This is best done through the cumulus parametrization applied in the forecast model - so as to ensure consistency between analyses and subsequent forecast (Fig.19). The necessary calculations are rather straightforward with the Kuo scheme used in the ECMWF model until May 1989, where the diagnosed precipitation rates are incorporated as follows (Heckley et al.,1990; see also Puri (1987), who takes a similar, although much simpler,

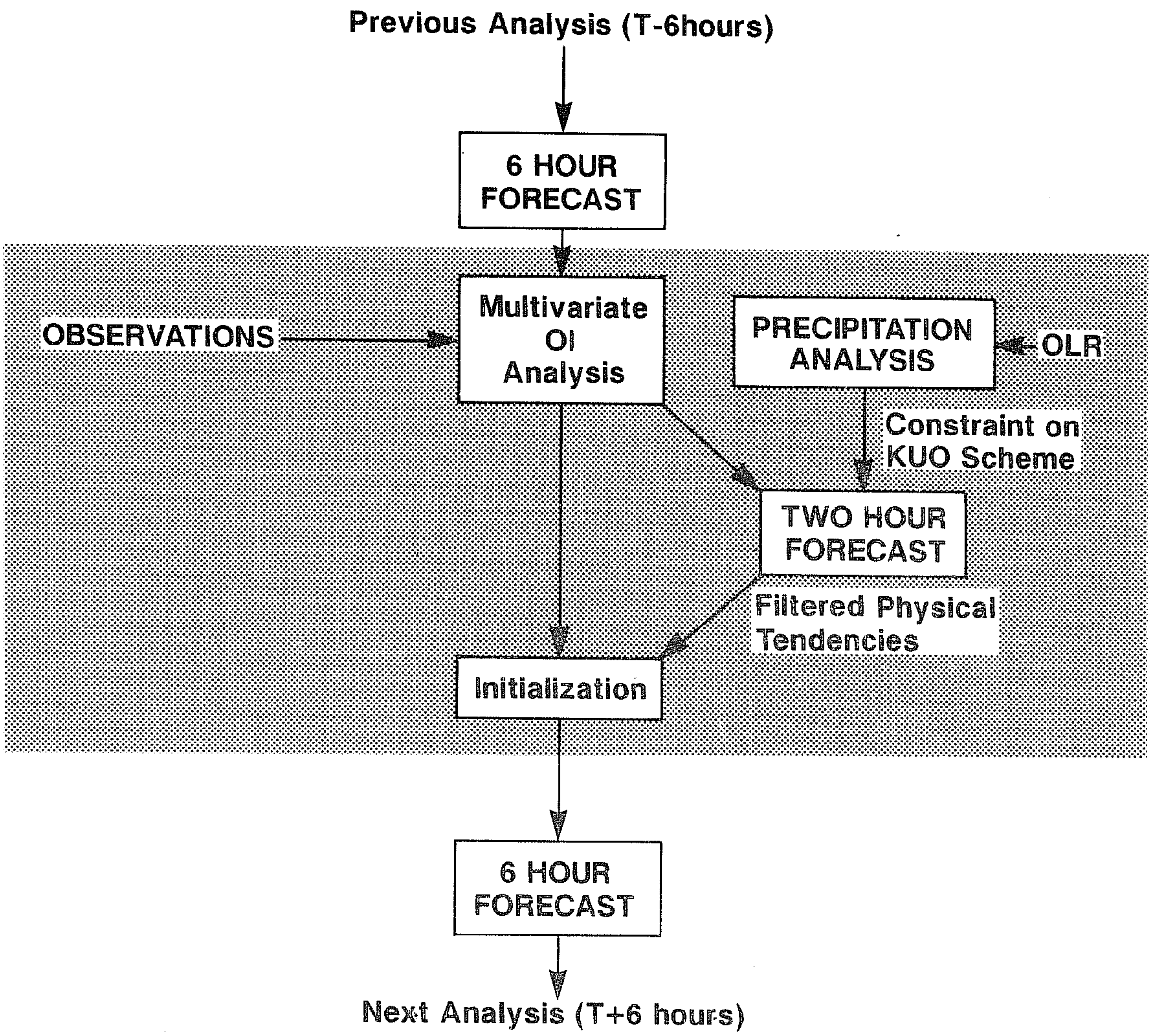


Fig.19 Schematic illustration of the assimilation scheme. The shaded area indicates analysis/initialization at time T. This procedure is repeated at intervals of 6 hours, interspersed by a six hour forecast which is used to provide a first guess for the subsequent analysis.

approach).

The appropriate input for the Kuo scheme is a moisture convergence into the grid area to produce precipitation but no moistening (moistening parameter β=0). Diagnosed values (DIAG) of moisture convergence are merged with the model generated values (MOD) as

$$(1\text{-wts})*\text{MOD} + \text{wts}*\text{DIAG},$$

depending on confidence in the diagnosed precipitation. Similarly the moistening parameter ß is set to (1-wts)*MODß, where MODß is the model ß parameter. In regions of high confidence the weights (wts) are close to unity, the moisture convergence is defined by the diagnosed precipitation rate and the moistening parameter is set to zero so that all the moisture convergence falls out as rain. Where there are no data, or in the extratropics, or where the data are far from the synoptic time, the weights (wts) will be close to zero and the moisture convergence will be set to its forecast value (MOD), as will the moistening parameter (MODß). Thus the scheme moves smoothly between forecast and diagnosed convection according to the specified field of confidence weights.

Figure 20 shows heating rates at model level 7 (about 200hPa), averaged over two hours of model integration from 12 UTC 2/2/87, for the Australasian region a) prior to filtering and (b) after filtering. No diagnosed heating has been used in this case. Note the 'spotty' nature of the model heating, and the intense local values. This is a region with little spin-up. Contrast the heating obtained when diagnosed heating is used in the model, Fig.20c shows the heating prior to filtering and Fig.20d after filtering. Diagnosed heating rates display a well organised and plausible structure. The effect of the filtering is to retain only the large scale information, reducing the local intensities considerably. However, even with the filtered fields (Figs.20b,d) there are substantial differences in the heating as diagnosed and generated by the model.

4.2.3 Analysis Impact

A data assimilation experiment has been carried out starting from 00 UTC on 1 February 1987 through to 12 UTC on 2 February 1987, using the version of the ECMWF forecast model and analysis system current in December 1988. As numerous changes have taken place since February 1987, a control assimilation (without the use of the diagnosed heating) has been carried out for the period 00 UTC on 1 February 1987 through to 12 UTC on 2 February 1987. This allows evaluation of the assimilation impact over seven analyses, six hours apart. This is the period of the AMEX experiment (Holland et al., 1986; Heckley and Puri, 1988), much additional conventional data were available during this time over the Australasian region.

It is interesting to look at the differences between the diagnosed heating and control assimilations cycle by cycle (smoothed using a spectral filter). Fig.21 shows the 200hPa streamline differences (contours are isotacs). Differences initially occur in the central and western Pacific, and to a lesser extent, over Indonesia. These differences grow in the areas of convective heating and extend both westwards and eastwards through advection, becoming largest in the eastern Pacific. By 12 UTC on 2 Feb 1987 differences of over $5ms^{-1}$ are not uncommon. Impact of the change is gradual, and increases with the length of the assimilation. Differences remain largely confined to relatively data sparse areas, as would be expected.

4.2.4 Discussion

Dominance of diabatic processes within the tropics suggests the utilisation of more direct measures of diabatic forcing within the assimilation. Use of satellite data to specify heating rates allows their use within the diabatic non-linear normal-mode initialisation. This not only introduces an additional source of data into the assimilation, it reduces the dependence of the analyses on the assimilating model's parametrization scheme (which is often in any case, suffering from spin-up problems). Use of diagnosed heating in the assimilation has a large impact, which suggests that the use of this additional data source is important.

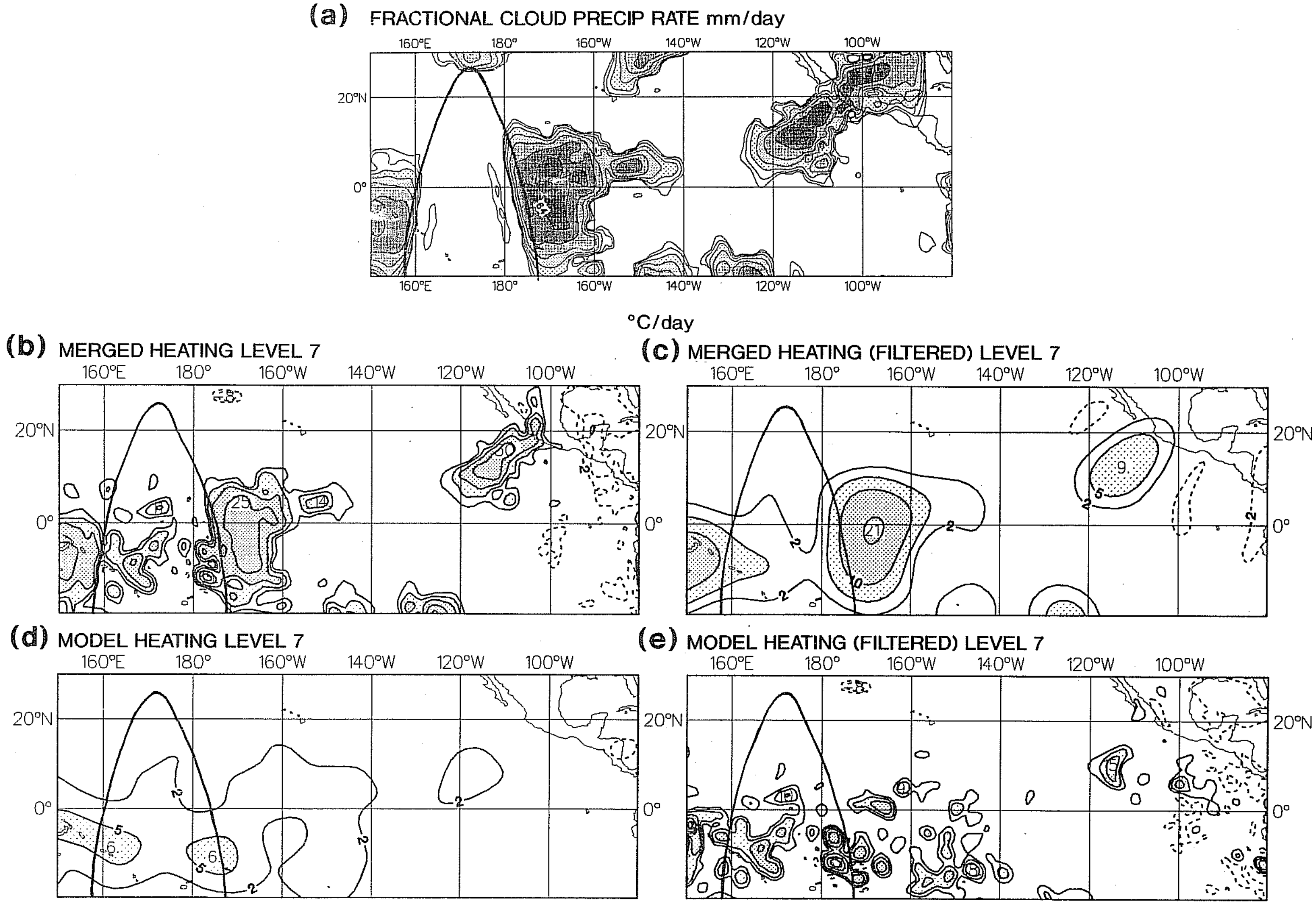


Fig.20 a) preciptation analysis 00 UTC 2/2/87, units: mm/day.
a)-d) Mean heating rates at model level 7 (about 200hPa) as averaged over two hours of integration from uninitialized analyses of 00 UTC 2/2/87.
a) prior to filtering, diagnosed heating
b) after filtering, diagnosed heating
c) prior to filtering, model only
d) after filtering, model only
Contours are at values of +/- 2, 5, 10, 20, 30, 40 °C per day. Negative values are dashed.

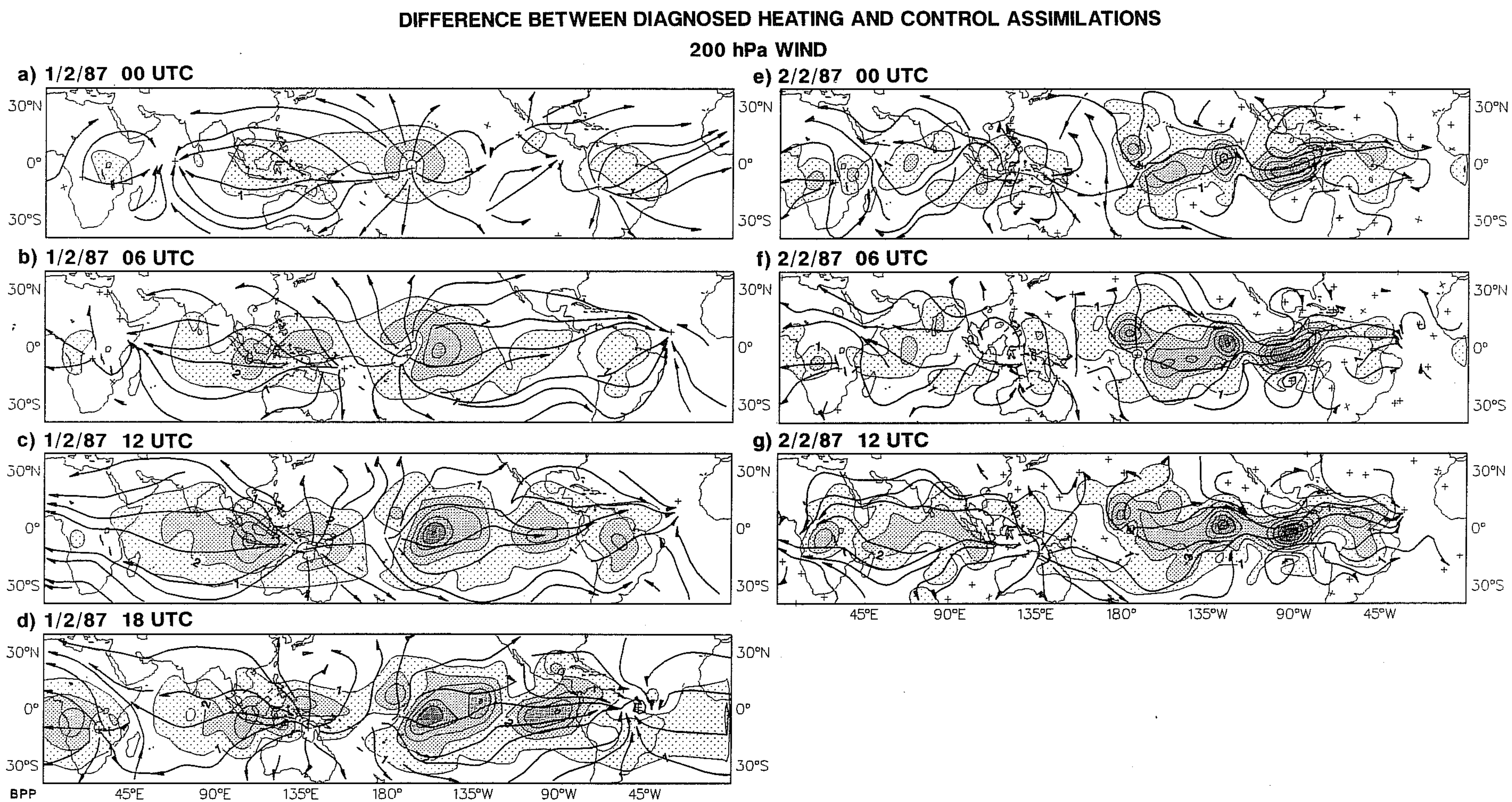


Fig.21 Difference between diagnosed heating and control assimilations cycle by cycle. 200hPa streamlines, contours are isotachs at interval 1 ms^{-1}. Fields are smoothed using a spectral filter.

The Heckley et al. (1990) study uses only very crude estimates of the precipitation and this information was heavily filtered through the way was used in the initialization. It should be possible to obtain better estimates of the precipitation, perhaps by taking into account other satellite information (humidity etc). Other instruments, such as SSM/I, will yield more direct and reliable estimates. Use of more reliable data such as this should also allow less filtering to be used in the initialization.

In view of the above reservations the work of Heckley et al. (1990) should perhaps be considered more as a sensitivity study rather than a definitive guide to initializing tropical forecasts. In this spirit it is clear that the use of diagnosed heating rates in the initialization has a fairly large impact in the analyses in data sparse regions. Evaluation of the quality of the analyses is not easy, but fit to observations suggest that both analysis quality and first guess quality were improved, at least below 150hPa. Some problems seemed to occur above 100hPa. Forecast impact was also significant and here too a modest improvement was seen.

5. PHYSICAL INITIALIZATION

The rotational part of the wind, temperature, and surface pressure can be represented, to some extent, by the current 4-D assimilation techniques using the currently available observations. The analysis of the humidity variable and divergent wind are in much worse shape.

Various efforts and proposals have been made to infer the divergence field, either based purely on the assumption that the analysis of the rotational wind is correct (Sardeshmukh, pers comm); or using additional data such as outgoing longwave radiation - using the strong relation that exists in the deep tropics between cold clouds and deep convection as discussed in section 4.

For the humidity field it has been suggested that it should be modified in regions of subsidence (within the tropics) such that a form of advective radiative balance exists, taking account of the observed OLR (Krishnamurti et al., 1984).

For regions of rain an adjustment of the initial humidity field, or humidity and temperature, through a reverse cumulus parametrization algorithm so that model precipitation rates agree with 'observed' rates (usually derived from OLR, or a combination of the latter with rain-gauge). Such techniques have been put forward by Krishnamurti et al. (1984), Donner (1988), Donner and Rasch (1989), amongst others.

More recently Krishnamurti et al. (1991) have advocated a more comprehensive 'initialization' which makes use of detailed data sets of outgoing longwave radiation. That along with the currently available tropical raingauge data sets are used to define the precipitation field. The initial surface fluxes (of water vapour and sensible heat) are determined from vertical integrals of the apparent moisture sink and the apparent heat source distributions (Yanai et al, 1973). These make use of the precipitation fields, which are further used to determine a consistent vertical structure of the humidity field from the use of reverse similarity and a reverse parametrization algorithm. The humidity analysis in the upper troposphere is structured to the satellite based outgoing longwave radiation. A Newtonian relaxation procedure (Krishnamurti et al. 1988) initializes the surface fluxes, rainfall rates and the outgoing longwave radiation. This procedure is shown to initialize the fluxes to within $1Wm^2$ (i.e. the reverse similarity theory and the Yanai fluxes converge to this degree of accuracy).

Krishnamurti et al. (1991) note that the accuracy of rainfall estimates is crucial for the determination of the surface moisture fluxes in the rain areas. Forecast experiments carried out by them showed improvement up to at least day five through the use of this procedure.

It is not clear, however, how applicable such techniques are in an operational context.

Constraints on spin-up and other initialization issues may be more easily addressed through the variational approach discussed below.

6. VARIATIONAL ASSIMILATION

All assimilation methods which are used (or have been used) in operational numerical weather prediction are sequential, in that they consist of a sequence of corrections performed on one temporal integration of the assimilation model over the assimilation period. The method of correction often performed is OI, which has been discussed earlier. In a sequential assimilation process each observation is used once without reference to anterior times. A comparison of the OI and variational approaches may be found in Pailleux (1990).

An alternative approach is a method in which one model solution is globally adjusted to the available observations, with propagation of the information contained in the observations both forward and backward in time. This is the basic idea which underlies variational assimilation. With present models and computers the only practical way to implement variational assimilation is through use of the so called adjoint of the assimilating model.

In general, the techniques require integration of the model equations forward, calculation of a 'cost function' based on the distance between the model trajectory and observations, and possibly other constraints. The gradient of this cost function has then to be calculated with respect to changes in the model state variables at the initial time, this involves using the adjoint of the model to carry the gradient information back to the initial state. The variational techniques adjust the initial state model variables such that the distance between the ensuing model trajectory and the observations is minimized. This minimization may require several iterations to converge, each of which requires an integration of the forecast model over the assimilation period. It follows that such techniques are likely to be horrendously expensive using current computer technology.

Variational assimilation of meteorological observations, with explicit use of the adjoint of the assimilating model, has been studied and used in the last few years by a number of authors (see, e.g., Lewis and Derber, 1985, Derber, 1987, Talagrand and Courtier, 1987, Courtier and Talagrand, 1987, Lorenc, 1988). Up till now most work has been done with barotropic or shallow water equation models.

At ECMWF a variational analysis scheme is being developed, based on the use of adjoint operators, which could use more data than the current OI technique, including data which are linked to the forecast model variables by non-linear operators (e.g. radiances linked to the temperatures by the radiative transfer equation). In this approach the satellite retrieval scheme and analysis scheme can be combined in one big minimization problem in the space of the model variables.

It is envisaged that the variational approach will yield benefits even in a 3-D mode, i.e. without the temporal component. In this case one still has the advantage of being able to relate model variables (e.g. temperature) to observed variables (e.g. radiances) in a consistent fashion. It should also be relatively straightforward to use other more exotic 'observations' such as precipitation, surface fluxes etc. Thereby introducing a more natural form of physical initialization.

7. NEW DATA SOURCES

New types of satellite data will be coming available in the first half of the coming decade:-

surface wind speed, integrated water vapour or integrated liquid water content, rain rate, from microwave imagers, such as SSM/I;

surface wind data from ERS-1 (1991) and from NASA scatterometer NSCAT on the Japanese satellite ADEOS (1994/5);

improved microwave temperature and humidity sounding data from AMSU (1994).

All the above instruments exist or are being built, and the time-scale for their launch is known to within a year or two. In the second half of the decade other, new, instruments are envisaged:-

higher resolution temperature and humidity sounding data from spectrometer instruments such as AIRS (100-4000 channels), or from interferometers, such as HIS or ISAS, each with up to 4000 channels;

global wind coverage with 1-2km resolution in the vertical from LAWS/ALADIN/BEST.

In this future time scale many of these projects are uncertain in terms of funding, time scales for deployment etc.

These new instruments offer a revolution in the amount and quality of data available to tropical analysis, but a great deal of research will have to go into the effective use of such data. A more comprehensive review of these new instruments and the data they offer may be found in the paper by Hollingsworth in these proceedings.

8. DISCUSSION

The aim of this work has been to convey the special nature of the problem of data assimilation within the tropics. In order to do this a selection of problems have been briefly described, which in the view of the author, are felt to be important. This is not a comprehensive list and reflects the interests and biases of the author. Whilst many of these problems are general in nature, examples have been largely drawn from ECMWF as a) these were readily available, and b) this is the system of which the author is most familiar.

Whilst many specific difficulties have been discussed, these all, by and large, arise from three basic problems.

1. Relative poor performance of the forecast model within the tropics - hence a relatively poor first guess.

2. A general lack of high quality data on the three-dimensional structure of the tropical flow.

3. Lack of any simple diagnostic relation between mass and wind in the deep tropics - difficulties with multivariate analysis.

One should not be too pessemistic. The quality of the forecast models is improving by leaps and bounds all the time. New data, in large quantities, will be available in the coming decades, from sophisticated instruments on space platforms.

Acknowledgements.
I should like to thank Dr Klaus Arpe for providing Fig.15. Anke Eriksson, Dr Martin Miller, and Robert Mureau provided invaluable comments on an early draft.

References

Anderson, A., A. Hollingsworth, S. Uppala and P. Woiceshyn, 1987: A study of the feasibility of using sea and wind information from the ERS-1 satellite. Part 1: wind scatterometer data. European Space Agency Contract Report. 121 pp. Available from ECMWF.

Andersson, E. and A. Hollingsworth, 1988: Typhoon bogus observations in the ECMWF data assimilation system. ECMWF RD Tech.Memo. No. 148. 25pp. Available from ECMWF.

Arkin, P.A., 1979: The relationship between fractional coverage of high cloud and rainfall accumulations during GATE over the B-scale array. *Mon.Wea.Rev.*, **107**, pp 1382-1387.

Arkin, P.A., and B.N. Meisner 1987: The relationship between large-scale convective rainfall and cold cloud over the Western hemisphere during 1982-84. *Mon.Wea.Rev.*, **115**, pp 51-74.

Arpe, K., 1989: Changes in the ECMWF analysis-forecasting scheme and the systematic error of the model. ECMWF RD Tech.Memo. No.161, 41pp.

Baer, F., 1977: Adjustment of initial conditions required to suppress gravity oscillations in nonlinear flows. *Beitr.Phys.Atmos.*,**50**, pp 350-366.

Barrett, E.C., and D.W Martin 1981: The use of satellite data in rainfall monitoring. Academic Press, 340 pp.

Bengtsson, L., M. Kanamitsu, P. Källberg and S. Uppala, 1982: FGGE 4-Dimensional Data Assimilation at ECMWF. *Bull.A.Met.Soc.*, **63**, pp 29-43.

Burt, S.D. and D.A. Mansfield, 1988: The great storm of 15-16 October 1987. *Weather*, **43**(3), pp 90-109.

Cats, G.J., and W. Wergen 1983: Analysis of large scale normal modes by the ECMWF analysis scheme. ECMWF workshop on current problems in data assimilation. pp 343-372. Available from ECMWF, Shinfield Park, Reading, England, RG2 9AX.

Chapman, S. and R.S. Lindzen, 1970: Atmospheric Tides, thermal and gravitational. D. Reidel Publishing Company, Dordrecht, Holland, 209pp.

Courtier, P. and O. Talagrand, 1987: Variational assimilation of meteorological observations with the adjoint vorticity equation. II: Numerical results.*Q.J.Meteorol.Soc.*,**113**,pp1329-1347.

Daley, R., 1983: Spectral characteristics of the ECMWF objective analysis system. ECMWF Tech. Rep. No. 40, 117 pp. Available from ECMWF.

Derber, J.C., 1987: Variational four-dimensional analysis using quasi-geostrophic constraints. *Mon.Wea.Rev.*,**115**,pp998-1008.

Donner, L.J., 1988: An initialization scheme for cumulus convection in Numerical Weather Prediction Models. *Mon.Wea.Rev.*, **116**, pp 377-385.

Donner, L.J. and P.J. Rasch 1989: Cumulus initialization in a global model for numerical weather prediction. *Mon. Wea. Rev.*,**117**,pp 2654-2671.

Dvorak, V.F., 1984: Tropical cyclone intensity analysis using satellite data. NOAA Technical Report NESDIS 11, US Dept of Commerce, NOAA/NESDIS, Washington DC, 43pp.

Eriksson, A. 1991: Use of cloud-motion winds at ECMWF. Proceedings of the 8th Meteosat user's meeting. August 1990, Norrköping, Sweden. Publ. by EUMETSAT.

Errico, R.M., 1989: The degree of Machenhauer balance in a climate model. *Mon.Wea.Rev.*, pp 2723-2733.

Errico, R.M, 1984: Dynamical balance produced by a general circulation model. *Mon.Wea.Rev.*, **112**, pp 2439-2454.

Errico, R.M. and D.L. Williamson, 1988: The behaviour of gravitational modes in numerical forecasts. *Mon.Wea.Rev.*, **116**, pp 1737-1756.

European Space Agency, 1985: Meteosat Image Bulletin. August 1985 and September 1985. Issued by MEP/Data Service European Space Operation Centre. Robert-Bosch-Str. 5 6100 Darmstadt, Germany.

Fiorino, M. and R.L. Elsberry, 1986: The role of vortex structure in tropical cyclone motion. Proceedings of the AMS 17th Conference on Hurricanes and Tropical Meteorology, Miami, 7-10 April, 1987. pp 55-59.

Girard, C. and M. Jarraud 1982: Short and medium range forecast differences between a spectral and grid point model: An extensive quasi-operational comparison. Tech. Rep. No. **32**, ECMWF, 176 pp. Available from ECMWF, Shinfield Park, Reading, England, RG2 9AX.

Hall, C.D., 1987: Verification of global model forecasts of tropical cyclones. *Meteor.Mag.*, **116**, pp 216-219.

Heckley, W.A., 1989: Data assimilation and anomalous lows. ECMWF RD Tech.Memo. No. 149. 16pp

Heckley, W.A., 1985: Systematic errors of the ECMWF operational forecasting model in tropical regions. *Quart.J.R.Met. Soc.*,**111**, pp. 709-738.

Heckley, W.A., 1983: A comparison of NMC and CAC sea-surface temperature analyses. ECMWF RD Tech.Memo. No. 83. 25pp.

Heckley, W.A., 1982: Adjustment in numerical weather prediction models in the tropics. ECMWF workshop on Current Problems in Data Assimilation. ECMWF 8-10 November, 1982. pp299-342.

Heckley, W.A., J.-J. Morcrette and E. Klinker, 1991: Applications of satellite radiance data in the medium range forecasting environment at ECMWF. *Advances in Space Research.* To appear.

Heckley, W.A., G. Kelly and M. Tiedtke, 1990: On the use of satellite derived heating rates for data assimilation within the tropics. *Mon.Wea.Rev.*,**118**, to appear.

Heckley, W.A. and K. Puri, 1988: The winter monsoon during AMEX. A quick look atlas. Available from ECMWF. 173pp. Available from ECMWF, Shinfield Park, Reading, England, RG2 9AX.

Heckley, W.A., M.J. Miller and A.K. Betts, 1987: An example of hurricane tracking and forecasting with a global analysis-forecast system. *Bull.Amer.Meteor.Soc.*,**68**, pp226-229.

Holland, G.J., 1983: Tropical cyclone motion: environmental interaction plus a beta effect.

J.Atmos.Sci., **40**, pp 328-342.

Holland, G.J., J.L. McBride, R.K. Smith, D. Jasper and T.D. Keenan, 1986: The BMRC Australian monsoon experiment: AMEX. *Bull.Amer.Met.Soc.*, **67**, pp 1466-1472.

Hollingsworth, A., 1986: Objective Analysis for Numerical Weather Prediction. Proceedings of the WMO symposium on Numerical Weather Prediction, Tokyo, August 1986.

Hollingsworth, A. and P. Lönnberg, 1986: The statistical structure of short-range forecast errors as determined from radiosonde data. Part 1: The wind field. *Tellus*, **38A**, pp 111-136.

Hollingsworth, A., D.B. Shaw, P. Lönnberg, L. Illari and A.J. Simmons, 1986: Monitoring of observation and analysis quality by a data assimilation system. *Mon.Wea.Rev.*, **114**, pp 861-879.

Illari, L., 1990: Spin-up in evaporation and use of satellite data. ECMWF RD Memo. R2295/2624. 26pp.

Illari, L., 1987: The "spin-up" problem. ECMWF RD Tech. Memo. No. 137, 31pp. Available from ECMWF.

Jones, D.E., 1976: The United Kingdom Meteorological Office objective analysis scheme for GATE. *Met.Magazine.* Vol **105**, No. 1250. pp 249-260.

Julian, P.R., 1984: Objective analysis in the tropics: A proposed scheme. *Mon.Wea.Rev.*, **112**, 1752-1767.

Kållberg, P., 1985: On the use of cloud track wind data from FGGE in the upper troposphere. ECMWF Research Department Tech. Memo. NO. **111**, 39 pp. Available from ECMWF, Shinfield Park, Reading, England, RG2 9AX.

Kanamitsu, M., 1985: Data processing, selection and quality control as used at JMA. Proceedings of the ECMWF workshop on the use and quality control of metorological observations, Reading 6-9 November 1984, pp 369-396.

Kasahara, A., R.C Balgovind and B.B. Katz, 1988: Use of Satellite Radiometric Imagery Data for Improvement in the Analysis of Divergent Wind in the Tropics. *Mon.Wea.Rev.*,**116**,866-883.

Klinker, E., 1990: Momentum budgets. ECMWF RD Tech.Memo. No.173, 26pp.

Krishnamurti, T.N., J. Xue, H.S. Bedi, K. Ingles and D. Oosterhof, 1991: Physical initialization for numerical weather prediction over the tropics. *Tellus*. To appear.

Krishnamurti, T.N., H.S. Bedi, W.A. Heckley and K. Ingles, 1988: Reduction of Spin-up Time for Evaporation and Precipitation in a Spectral Model. *Mon.Wea.Rev.* **116**, pp. 907-920.

Krishnamurti, T.N. and S. Low-Nam, 1986: On the Relationship between the Outgoing Long Wave Radiation and the Divergent Circulation.*J.Meteor.Soc.Japan*,**64**,709-719.

Krishnamurti, T.N., K. Ingles, S. Cocke, T. Kitade and R. Pasch, 1984: Details of low latitude medium range numerical weather prediction using a global spectral model II. Effects of orography and physical initialization. *J.Meteor.Soc.Japan*,**62**,613-649.

Krishnamurti, T.N., P. Ardauny, Y. Ramanathan and R. Pasch, 1981: On the onset vortex of the summer monsoon. *Mon.Wea.Rev.*, pp 344-363.

Leith, C.E., 1980: Nonlinear normal mode initialization and quasi-geostrophic theory. *J.Atmos.Sci.*, **37**, pp 958-968.

Lewis, J.M. and J.C. Derber, 1985: The use of adjoint equations to solve a variational adjustment problem with advective constraints. *Tellus*,**37A**,pp309-322.

Lönnberg, P., 1989: Developments in the ECMWF analysis system. Proceedings of the 1988 ECMWF Seminar on Data assimilation and the use of satellite data. Vol.I. pp75-119.

Lönnberg, P. and A. Hollingsworth, 1986: The statistical structure of short-range forecast errors as determined from radiosonde data. Part II: The covariance of height and wind errors. *Tellus*, **38A**, pp 137-161.

Lorenc, A.C., 1988: Optimal nonlinear objective analysis. *Q.J.R.Meteorol.Soc.*, **114**,pp205-240.

Lorenc, A.C., 1981: A global three-dimensional multivariate statistical interpolation scheme. *Mon.Wea.Rev.*, **109**, pp 701-721.

Lorenc, A.C., R.S. Bell and B. Macpherson, 1991: The Meteorological Office 'Analysis Correction ' data assimilation scheme. *Quart.J.R.Met.Soc.*, **117**. To appear.

Lyne, W.H., R. Swinbank and N.T. Birch, 1982: A data assimilation experiment and the global circulation during the FGGE special observing periods. *Quart.J.R.Met. Soc.*, **457**, pp 575 - 594.

Machenhauer, B., 1977: On the dynamics of gravity oscillations in a shallow water model, with applications to normal mode initialization. *Beitr.Phys.Atmos.*, **50**, pp 253-271.

deMaria, M., 1986: The effect of vortex structure on barotropic hurricane track forecasts. Proceedings of the AMS 17th Conference on Hurricanes and Tropical Meteorology, Miami, 7-10 April, 1987. pp 52-54.

Marks, D.G., 1986: A technique to improve the MFM 12-hour forecast by modifying the spin-up. Proceedings of the AMS 17th Conference on Hurricanes and Tropical Meteorology, Miami, 7-10 April, 1987. pp 64-66.

Mather, M.B., 1986: NMC's high resolution hurricane model. Proceedings of the AMS 17th Conference on Hurricanes and Tropical Meteorology, Miami, 7-10 April, 1987. pp 60-63.

Milne-Thomson, L.M., 1968: Theoretical Hydrodynamics. MacMillan Press, London, 743pp.

Neumann, C.J., 1988: The National Hurricane Center NHC83 model. NOAA Technical Memorandum, NWS NHC 41, 44pp. Available from National Hurricane Center, Coral Gables, Florida, USA.

Pailleux, J., 1990: Data assimilation: optimum interpolation approach/variational approach. Proceedings of the ECMWF Seminar, 4-8 September, 1989. Volume I. pp 55-67. Available from ECMWF.

Phillips, N.A., 1963: Geostrophic motion. *Rev.Geophys.*,**1**, 123-176.

Puri, K., 1987: Specification of diabatic heating in the initialization of numerical weather prediction models. ECMWF workshop on Diabatic Forcing pp 151-176. Available from ECMWF, Shinfield Park, Reading, England, RG2 9AX.

Reed, R.J., A. Hollingsworth, W.A. Heckley and F. Delsol, 1988: An evaluation of the performance of the ECMWF operational system in analyzing and forecasting easterly wave disturbances over Africa and the tropical Atlantic. *Mon.Wea.Rev.*, **116**, pp 824-865.

Rowntree, P.R., M.F. Wilson and A.B. Sangster, 1985: Impact of land surface variations on African rainfall in general circulation models. UK Meteorological Office Dynamical Climatology Technical Note No. 30. 52pp. Available from UKMO, Bracknell, UK.

Ruml, M., A.J. Simmons and D. Vasiljevic, 1986: Comparison of Tropical analyses produced by ECMWF and the UK Meteorological Office. CAS/JSC/WGNE-II. Minutes.
Shaw, D.B., P. Lönnberg, A. Hollingsworth and P. Undén, 1987:Data assimilation: The 1984/85 revisions of the ECMWF mass and wind analysis. *Q.J.R.Meteorol.Soc.*, **113**, pp 533-566.

Shaw, D.B., P. Lönnberg, A. Hollingsworth and P. Undén, 1987: Data assimilation: The 1984/85 revisions of the ECMWF mass and wind analysis. *Q.J.R.Meteorol.Soc.*,**113**, 361-378.

Shukla, J. and Y. Mintz, 1982: The influence of land-surface evapotranspiration on the Earth's climate. *Science*, **215**, 19th March 1982, pp 1498-1501.

Talagrand, O. and P. Courtier, 1987: Variational assimilation of meteorological observations with the adjoint vorticity equation. I: Theory. *Q.J.Meteorol.Soc.*,**113**,pp1311-1328.

Tiedtke, M., W.A. Heckley and J. Slingo, 1988: Tropical forecasting at ECMWF: the influence of physical parametrization on the structure of forecasts and analyses. *Q.J.R.Meteorol.Soc.*, **114**, pp 639-664.

Undén, P., 1989: Tropical Data Assimilation and Analysis of Divergence. *Mon.Wea.Rev.*, **117**, pp 2495-2517.

Vasiljevic, D., 1989: Soil wetness analysis. ECMWF RD Memo R2327/2538. 25pp.

Wergen, W., 1990: The effect of model errors on four-dimensional variational assimilation. Proceedings of the International Symposium on Assimilation of Observations in Meteorology and Oceanography. Clermont-Ferrand, France, 9-13 July, 1990. Published by WMO, pp 45-46.

Wergen, W., 1989: Normal mode initialization and atmospheric tides. *Q.J.R.Meteorol.Soc.*, **115**, pp 535-545.

Wergen, W., 1988: The Diabatic ECMWF Normal Mode Initialization Scheme. *Beitr.Phys.Atmosph.*, **61**, pp 274-302.

WMO, 1986: Workshop on comparison of simulations by numerical models of the sensitivity of the atmospheric circulation to sea surface temperature anomalies. WMO/TD-No. 138., WCP-121, World Meteorological Organization, Geneva, 188pp.

Yanai, M., S. Esbensen and J.H. Chu, 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budgets. *J.Atmos.Sci.*,**30**,pp611-627.