1. INTRODUCTION

It is well known that the evaporation from tropical oceans, particularly in the warm pool areas, plays an important role in the tropical circulation. This has also become clear from recent experiments with the ECMWF model. An artificial increase by 1 K of the sea surface temperature in the Indonesia area and the Western Pacific (between 10°S and 20°N and between 95°E and 140°E) produces a much more realistic climatological rainfall pattern than the current operational model. Ninety day averages of precipitation are shown in Fig. 1 for T42 NH summer simulations with and without the 1 K SST increase. One of the typical deficiencies of the current ECMWF model "climate" is the tendency to produce two rain bands over the Indian ocean and Indonesia. The 1 K anomaly experiment shows much less of this tendency.

Because warm pool regions are areas with relatively low wind speed, it is expected that the model climatology will also be sensitive to the evaporation parametrization at low wind speeds. In this report the current parametrization is briefly described and improvements are proposed. The model response will be illustrated with a preliminary implementation of the improved scheme.

Evaporation from the sea is determined by the wind speed, by the moisture and temperature difference between the lowest model level and the sea surface and by the transfer coefficients. The neutral part of the transfer coefficients is related to the sea surface roughness lengths and is specified on the basis of empirical data (e.g. the Charnock relation). The stability corrections are based on Monin Obukhov theory.

It will be shown that for low wind speeds the ECMWF scheme (model cycle 34) has three deficiencies: (i) the stability functions are in error by 50% over very smooth surfaces, (ii) the sea surface roughness lengths do not have the correct smooth surface limits for low wind speeds and (iii) the model does not account for the near-surface wind induced by the convective motion.

The purpose of this report is to illustrate the importance of the air-sea flux parametrization at low wind speeds over sea. This is done by showing the model sensitivity to the details of the parametrization. It will also become clear that virtually no ocean data exists to guide us with regard to the three problems mentioned above. It is hoped that the model sensitivity demonstrated here will provide additional stimulus to experimental work on the transfer coefficients at low wind speeds such as during the TOGA/COARE program.

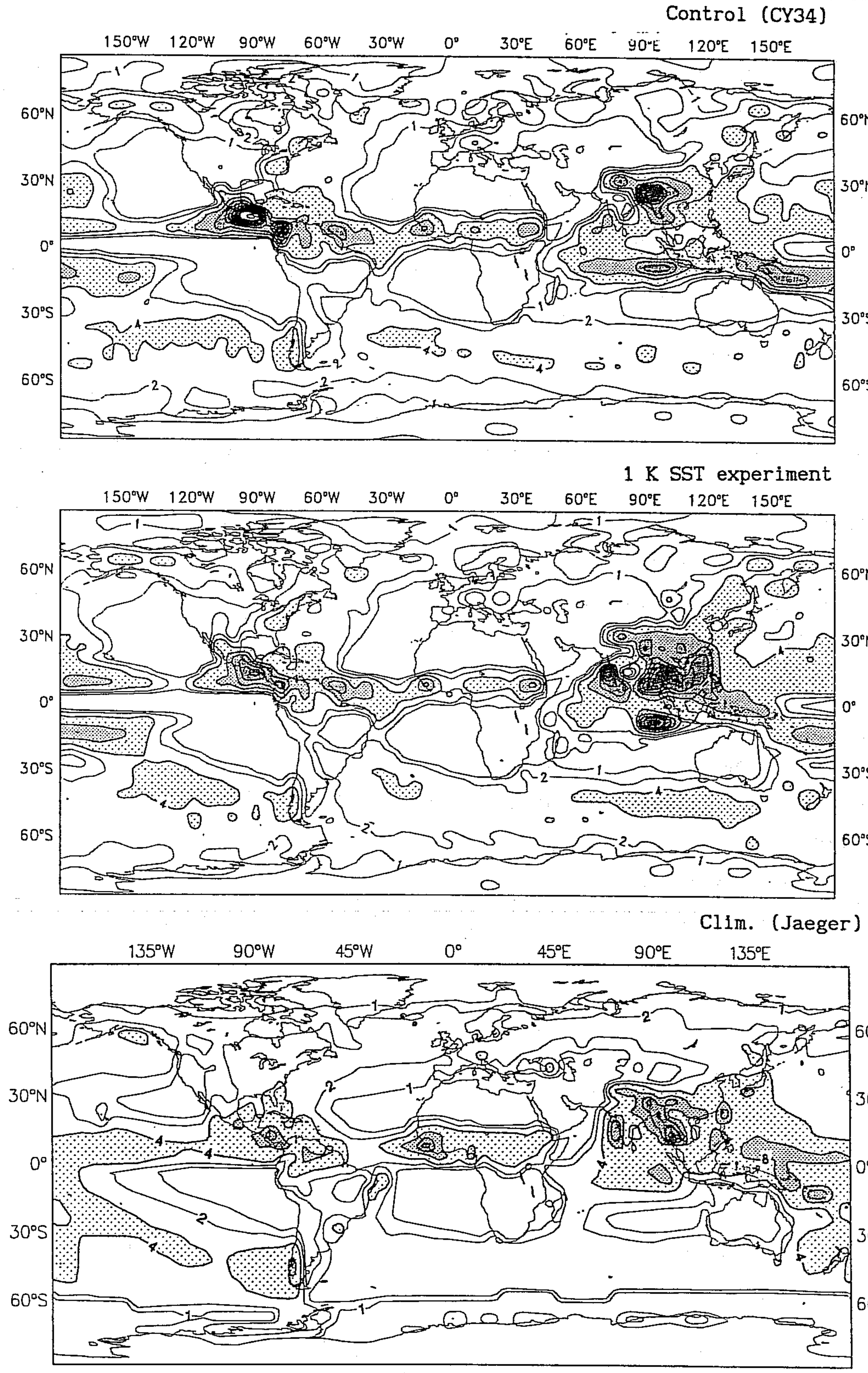

Fig. 1 Total rainfall averaged over 90 days from a T42 90-day experiment with observed SST (June, July and August; upper panel). The middle panel shows the rainfall for a similar run with the SST artificially increased by 1 K in the area between 10°S and 20°N and between 95°E and 140°E. The lower panel shows the climatology by Jaeger for June, July and August.

2. PARAMETRIZATION OF THE TRANSFER COEFFICIENT FOR MOISTURE OVER SEA

In the ECMWF model the evaporation is parametrized by means of

$$-\frac{E}{\rho} = C_Q \, | \underline{U}_1 | \, (q_1 - q_s) , \tag{1}$$

where E is the evaporation, ρ the air density, C_Q the transfer coefficient for moisture, $\underline{U}_1$ the horizontal velocity vector at the lowest model level, q_1 the specific humidity at the lowest model level and q_s the specific humidity at the surface (saturation value at SST). According to Monin Obukhov similarity, C_Q can be expressed in a neutral part C_{QN} and a stability correction F_Q (e.g. Louis, 1979; Holtslag and Beljaars, 1988)

$$C_Q = C_{QN} \, F_Q(Ri_o, z_1/z_{oM}, z_1/z_{oH}, z_1/z_{oQ}) , \tag{2}$$

$$C_{QN} = \frac{k^2}{\ln\{z_1/z_{oQ}\} \, \ln\{z_1/z_{oM}\}} , \tag{3}$$

where Ri_o is the bulk Richardson number between the surface and the first model layer based on the virtual potential temperature difference, k is the VonKarman constant (0.4), z_1 is the height of the first model level above the surface and z_{oM}, z_{oH}, z_{oQ} are the sea surface roughness lengths for momentum, heat and moisture respectively.

In the current ECMWF model the roughness lengths for heat and moisture are taken equal to the one for momentum, with z_{oM} according to the Charnock relation

$$z_{oM} = 0.018 \, \frac{u_*^2}{g} \; , \; z_{oM} > 1.5 \; 10^{-5} ,$$

$$z_{oH} = z_{oQ} = z_{oM} \; . \tag{4}$$

In these equations u_* represents the friction velocity and g the gravitational constant. The stability function F_Q, that is presently in use in the ECMWF model is depicted in Fig. 2 for a ratio $z_1/z_{oM} = 10^6$, which is typical for a weak wind situation over sea. When this function is recomputed from the Monin Obukhov profile functions as proposed by Dyer and Hicks (1970) it becomes clear that the current functions underestimate the stability correction (see Fig. 2 for a comparison).

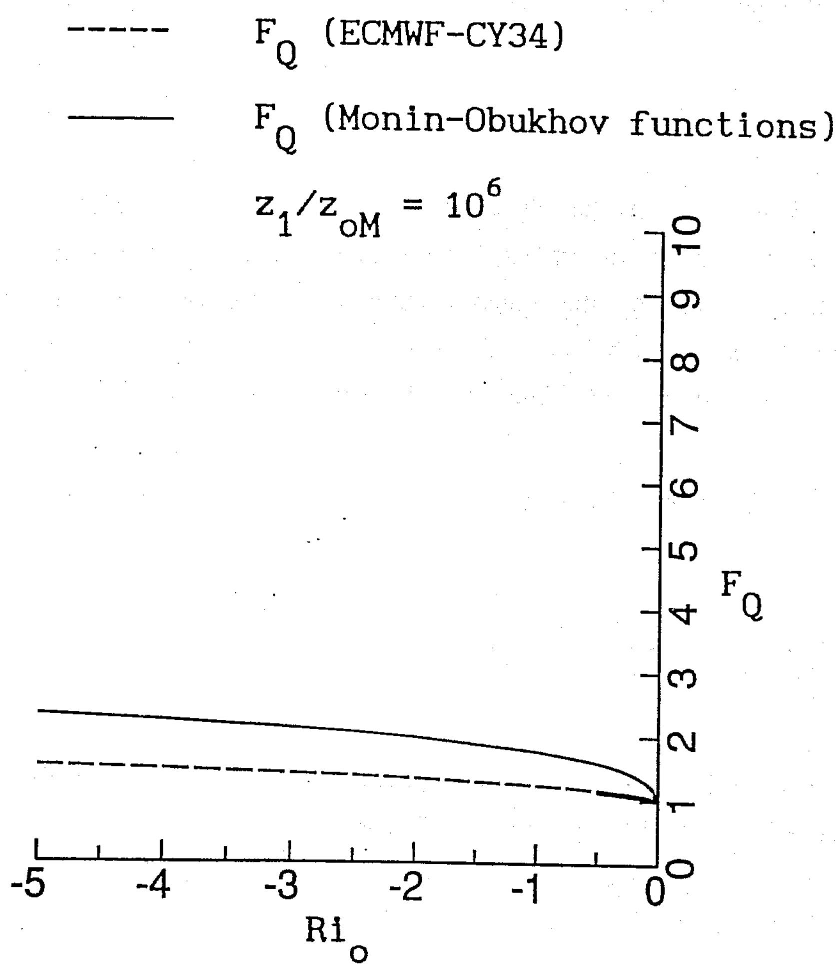

Fig. 2 The stability function of the bulk Richardson number as used in the ECMWF model (CY-34) and as computed from the Dyer and Hicks similarity functions (these are functions of height divided by the Obukhov length).

A second shortcoming of the parametrization in the ECMWF model is that equation (4) does not account for smooth surface scaling at low wind speeds. In weak wind conditions the sea surface becomes aerodynamically smooth and the roughness lengths (or rather integration constants in the logarithmic profiles) become proportional to ν/u_*, where ν is the kinematic viscosity of air ($\approx$ 1.5 10^{-5} m^2/s). The constants in the three roughness lengths are different and depend on the Prandtl and Schmidt number (e.g. Brutsaert, 1982).

In the rough regime the wind speed dependence of C_{MN} is well described by the Charnock relation (Smith, 1988), although some dependence of the wave spectrum is to be expected. For C_{HN} and C_{QN} the situation is quite different. In a recent review Smith (1988, 1989) suggests that the wind speed dependence in C_{HN} and C_{QN} is much weaker than in C_{MN}, if present at all. He suggests constant values for z_{oH} and z_{oQ}. By combining the smooth surface values (e.g. according to Brutsaert, 1982) with the Charnock relation for the aerodynamic roughness length and the constant values for heat and moisture in accordance with Smith's suggestions, we find

$$z_{oM} = 0.11\, \nu/u_* + 0.018\, u_*^2/g \ ,$$

$$z_{oH} = 0.40\, \nu/u_* + 1.4\ 10^{-5} \ , \qquad (5)$$

$$z_{oQ} = 0.62\, \nu/u_* + 1.3\ 10^{-4} \ .$$

The neutral transfer coefficient for moisture that results from equations (5) is compared with data by Large and Pond (1982) in Fig. 3. Also the current ECMWF parametrization is shown. For wind speeds below 8 m/s, the new parametrization gives larger transfer coefficients; for strong winds the evaporation is reduced. For wind speeds below 5 m/s, hardly any data exists, but it is felt that the evaporation from the ocean can not be smaller than the evaporation from an aerodynamically smooth surface for which the transfer properties are well established from laboratory experiments (Brutsaert, 1982).

Equation (1) indicates that the absolute magnitude of the near-surface wind is very important for the evaporation. In the case of heating from the surface the horizontal wind near the surface never drops to zero even if the horizontally averaged wind vector (as resolved by the model) is negligible. The convective motion driven by the surface buoyancy flux, results in a finite wind near the surface (e.g. Schumann, 1988; Liu et al.,

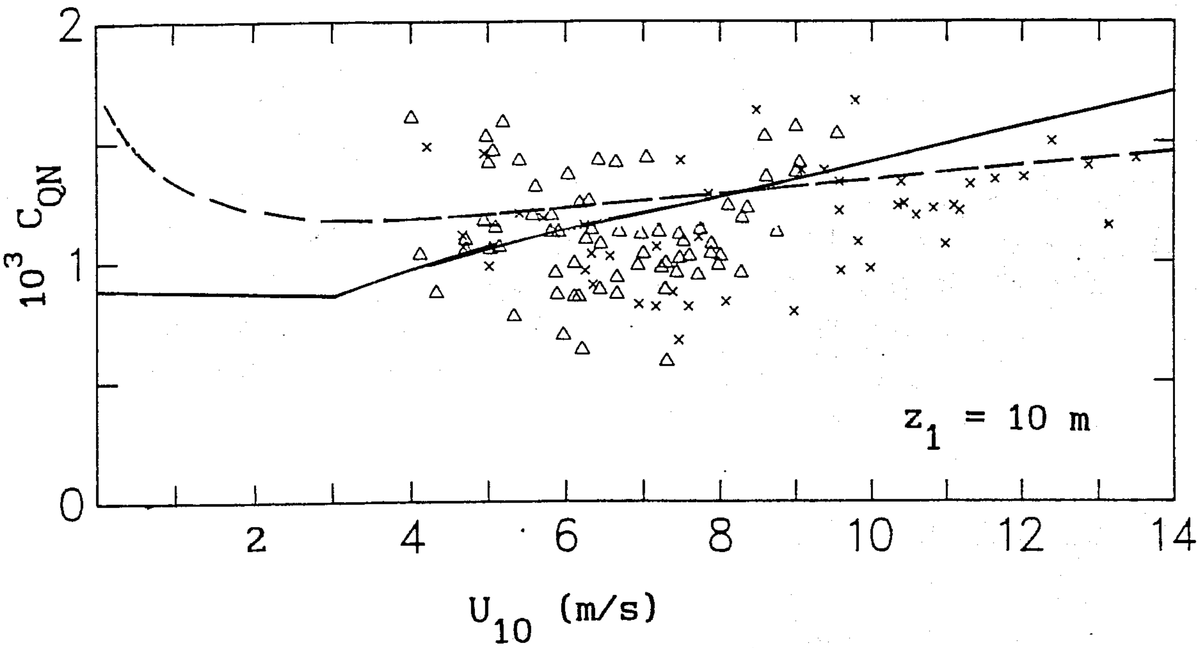

Fig. 3 Neutral transfer coefficients for moisture computed from the ECMWF scheme (CY34) for a reference height of 10 m (thick solid line). The dashed line has been derived from the improved scheme with equations (3) and (5). The triangles and crosses represent data by Large and Pond (1982).

1979), which also drives the evaporation. Deep convection will induce much larger surface winds and bring dryer air into contact with the surface. Parametrization of these effects is currently in progress. Here we therefore consider only the dry convective motions for which the scaling behavior is well established (e.g. Panofsky et al., 1977). Inclusion of the so-called mixed-layer scaling in the parametrization results in a modification of equation (1):

$$-\frac{E}{\rho} = C_Q \{ |\underline{U}_1|^2 + w_*^2 \}^{1/2} (q_1 - q_s) , \tag{6}$$

$$w_* = \{ z_i \, g \frac{\overline{(\rho' w')}}{\rho} \}^{1/3} , \tag{7}$$

where z_i is the boundary layer height and $g\overline{(\rho' w')}/\rho$ is the buoyancy flux at the surface (depends on heat and moisture flux). Since the dependence on z_i is weak, it is sufficient to use an order of magnitude estimation of z_i. We will use $z_i = 1000$ m.

It can be shown that the three ingredients of the new parametrization, (i) improved stability functions, (ii) appropriate smooth surface limits for the roughness lengths and (iii) convection induced surface wind, result approximately in the free convection limit found in tank experiments (Liu et al., 1979; Deardorff, 1972; Townsend, 1964)

$$\frac{E}{\rho} = 0.17 \{ \frac{g}{\theta_v} \frac{\kappa^2}{v} \}^{1/3} \{ \theta_{vs} - \theta_{v1} \}^{1/3} \{ q_s - q_1 \} , \tag{8}$$

where $\theta_{vs} - \theta_{v1}$ represents the virtual potential temperature difference between the surface and the first model layer, θ_v the absolute virtual potential temperature and κ the thermal diffusivity of air.

A comparison of the new parametrization with the current operational scheme is shown in Fig. 4 for a typical situation over a tropical ocean with a virtual potential temperature difference between the surface and the first model level of 1.5 K and a specific humidity difference of 7 g/kg. For low wind speeds the new scheme enhances the evaporation; for strong winds the moisture flux is reduced.

Unfortunately, the current model framework does not allow the distinction between different roughness lengths for momentum, heat and moisture. Since we want to concentrate on the effects of the parametrization at low wind speeds, a preliminary

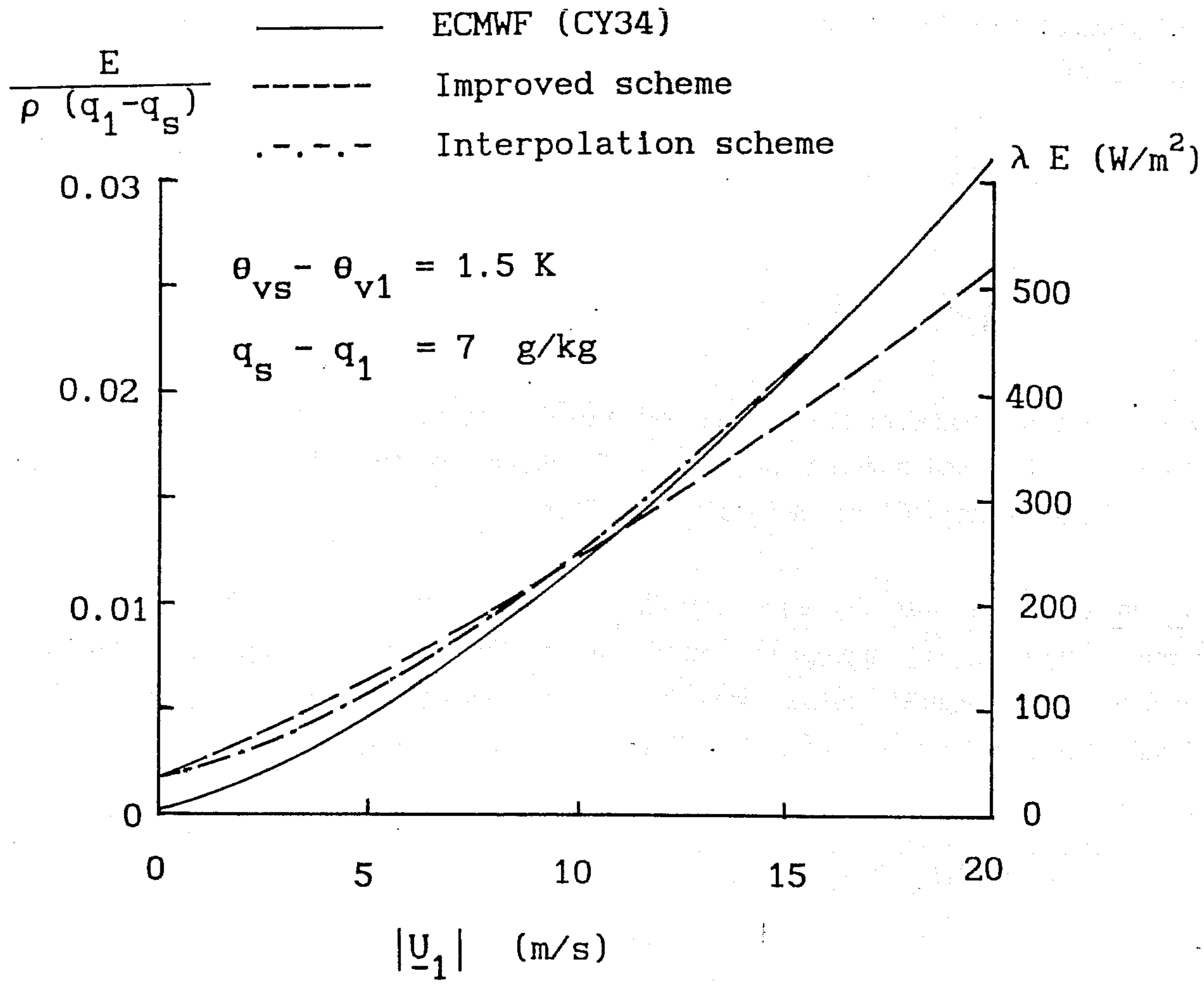

Fig. 4 The "conductivity" of the lowest model layer for moisture transfer as a function of wind speed. A typical tropical situation has been chosen with an SST of 30 C and a virtual potential temperature difference over the lowest model layer of 1.5 K. The right hand scale indicates the latent heat flux for a specific humidity difference of 7 g/kg.

implementation is sought. An empirical interpolation is used between the free convection limit of equation (8) and the neutral approximation of the current formulation (this is very similar to a proposal by Sill and Asce, 1983 as reviewed by Gordon and Hunt, 1989)

$$C_Q = C_{QN} (1 + C_R^{\gamma})^{1/\gamma} ,$$

$$C_Q = C_H , \gamma = 1.25 ,$$

$$C_{QN} = \left(\frac{k}{\ln(z_1 / z_{oM})} \right)^2 ,$$

$$z_{oM} = \text{according to the current formulation (eq. 4)} ,$$

$$C_R = \frac{0.0016}{C_{QN} \mid \underline{U}_1 \mid} (\theta_{vs} - \theta_{v1})^{1/3} . \qquad (9)$$

The preliminary formulation (indicated by "interpolation scheme") is also shown in Fig.4 in comparison with the current parametrization and the one that will ultimately will be used after the model framework has been adapted. The interpolation scheme has the correct free convection limit for negligible wind speed and provides a smooth transition towards the current formulation at higher wind speeds.

3. MODEL SENSITIVITY

To investigate the sensitivity of the model to the parametrization of transfer coefficients between the sea surface and the lowest model level, a winter and a summer 90 day T42 experiment have been carried out with and without the improved scheme according to equations (9). This scheme is in fact an interpolation between the free convection limit for smooth surfaces and the neutral transfer coefficients as currently in use in the ECMWF model. We will refer to the new parametrization as the "interpolation scheme".

The 90 day averages of the rainfall (see Fig. 5 and 6 for global maps of summer and winter experiments and Fig. 7 for the zonal average) show major changes. The differences are most distinct over the Indian ocean and the Western Pacific where the current model (CY34 for the control experiment) tends to develop a North-South division in the ITCZ. Although such split ITCZ's are observed locally, it is not an observed climatological feature but a known, documented model systematic error (Mohanty, personal communication). The

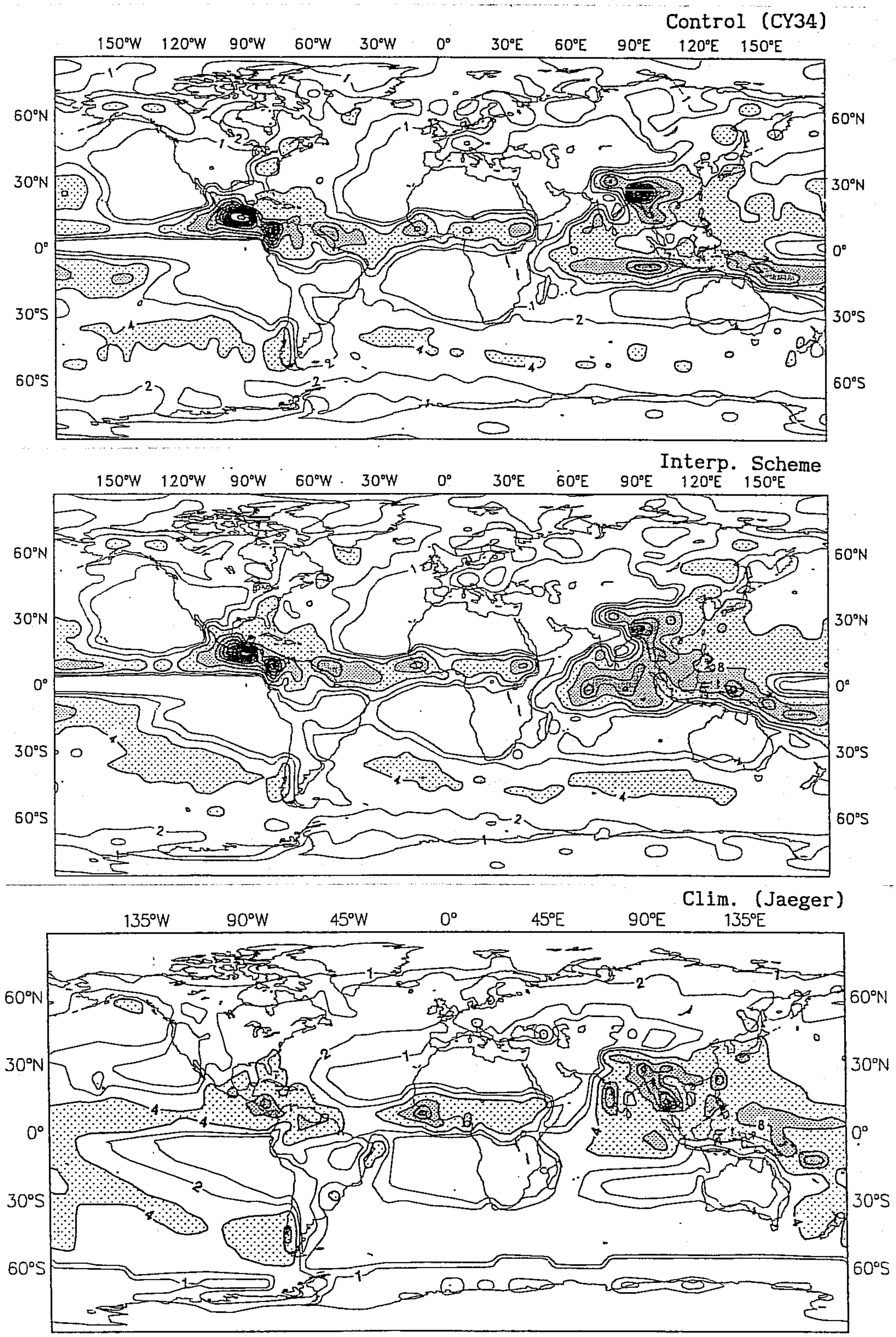

Fig. 5 Rainfall averaged over 90 days of a T42 NH summer experiment. The upper panel refers to the current operational model (CY34); the middle panel shows the experiment with the interpolation scheme and the lower panel shows the climatology by Jaeger for June, July and August.

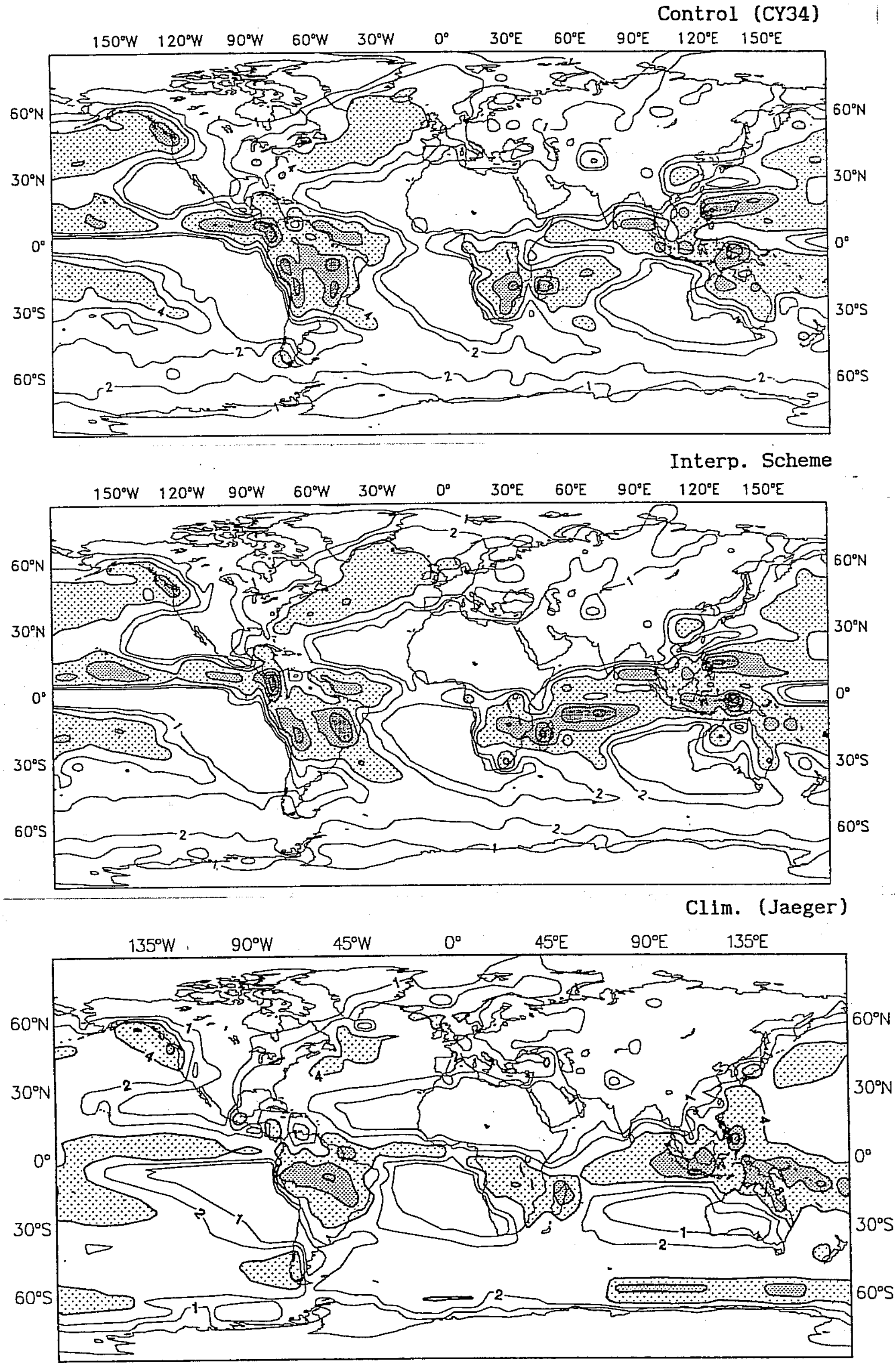

Fig. 6 As Fig. 5 but for December, January and February.

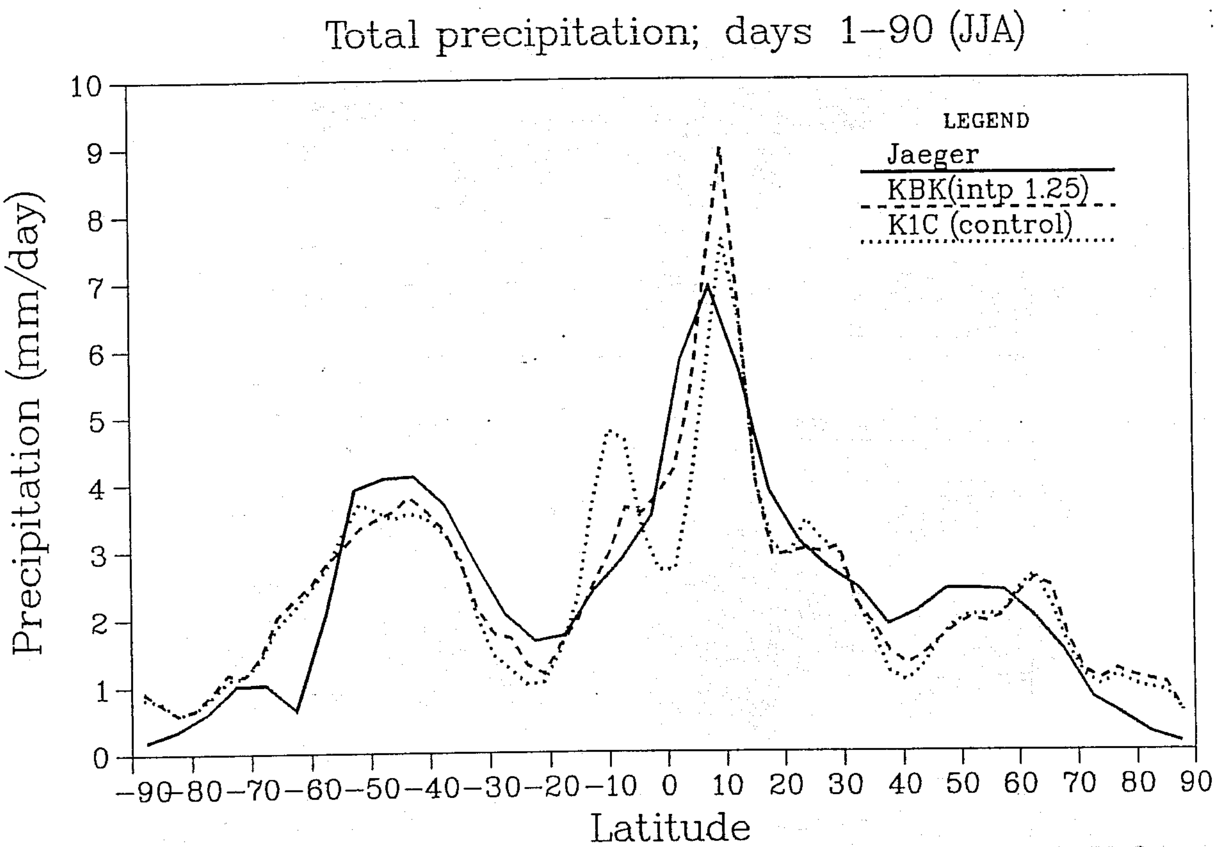

Fig. 7 Zonally averaged precipitation for the revised transfer coefficients (interpolation scheme, dashed line) in comparison with the results of the control experiment (dotted line) and the climatology by Jaeger (solid line). These are 90 day averages for the June-July-August experiments.

revised parametrization virtually eliminates the split. This is very clear for the summer case where the Southern branch of the ITCZ has been moved North. Consistent with this improved forecasted rainfall distribution for the summer is a greatly improved Indian Monsoon circulation (Fig. 8).

Tropical wind errors are much reduced in winter, but more mixed in summer. This is clear from the zonally averaged U-errors as shown in Fig. 9. The omnipresent upper tropospheric easterly error of about 8 m/s at 200 hPa in the tropics has been reduced to about 4 m/s. The main contribution to this reduction comes from the Eastern Pacific (see Fig. 10). However, the changes in the parametrization influence the entire tropical circulation, particularly through the Walker circulation.

4. CONCLUSIONS

Apart from the 90 day "climate" type results described above we have carried out extensive high resolution (T106) experimentation (including data assimilation) in the context of medium-range weather prediction. This confirms our main conclusion that the tropical circulation is extremely sensitive to apparently minor changes in the way the evaporation from tropical oceans is parametrized. By increasing the latent heat flux at low wind speeds (< 5 m/s) by no more than 25 W/m^2, we find major impact on the systematic wind errors and on the rainfall patterns.

Although the proposed parametrization is more realistic than the one currently operational, there is still very little oceanographic data with which to verify. The rationale for the new parametrization is that it seems impossible that the tropical oceans would evaporate less than an aerodynamically smooth water surface. (So the "free convection limit" can be seen as a safe lower bound.)

Much work remains to be done however. To provide guidelines for parametrization, oceanographic experiments would be needed that concentrate on the following aspects:

- Measurements of the transfer coefficients for moisture are needed at low wind speeds (below 5 m/s). Virtually no data exists at this moment (cf. Fig. 3).

- In areas with deep convection, the convection generates a wind near the surface that is completely sub-grid to the model and therefore not resolved by the model. This surface wind is relevant to the parametrization of evaporation as it enhances the moisture flux. The magnitude of the convection-induced surface wind and its impact on the evaporation is not well know and should be investigated experimentally.

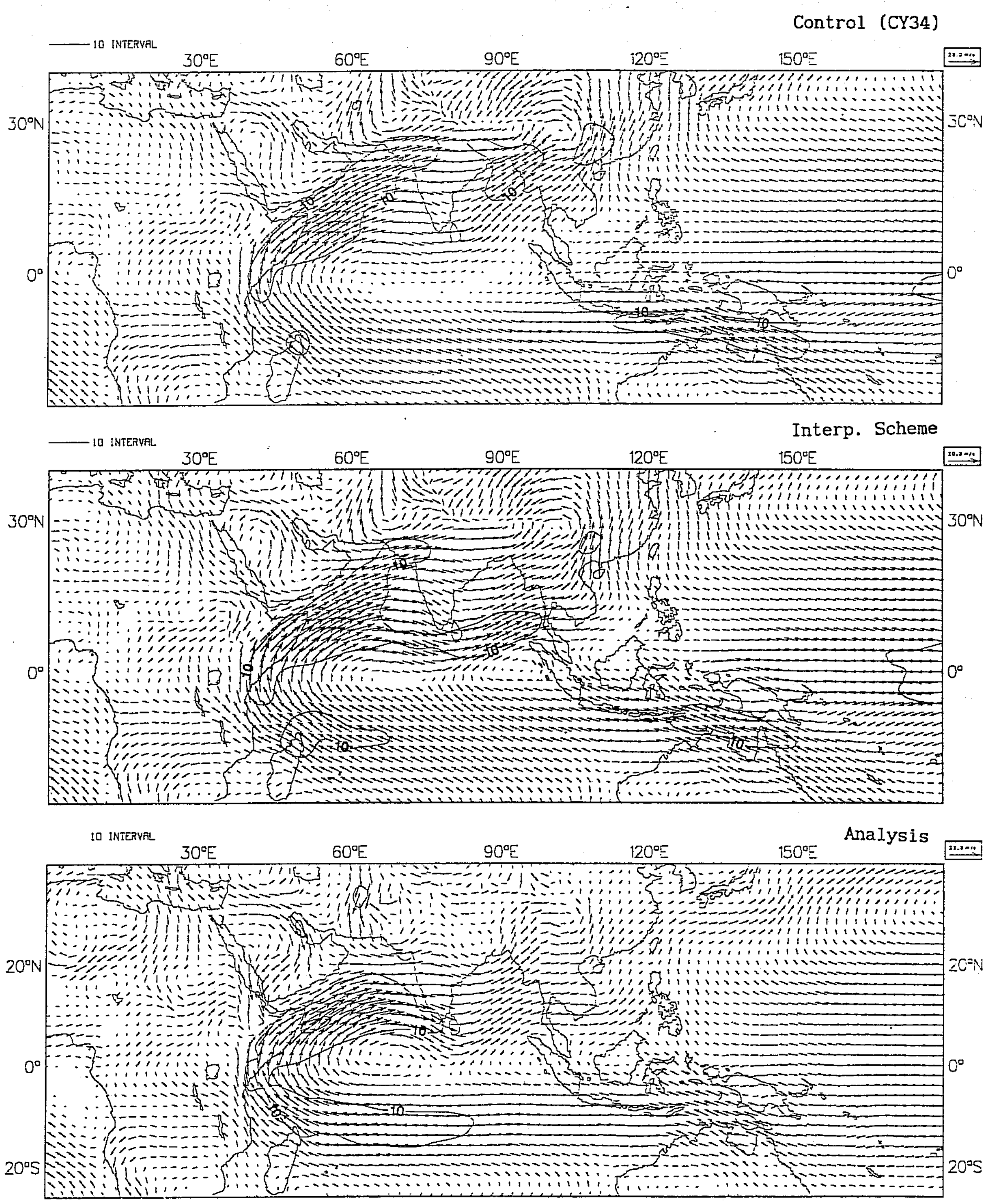

Fig. 8 90 day average of the level 15 (≈ 850 hPa) wind field for the 90 day T42 NH summer run. The upper panel shows the control experiment (CY34); the middle panel refers to the experiment with the interpolation scheme and the lower panel shows the operationally analyzed wind field.

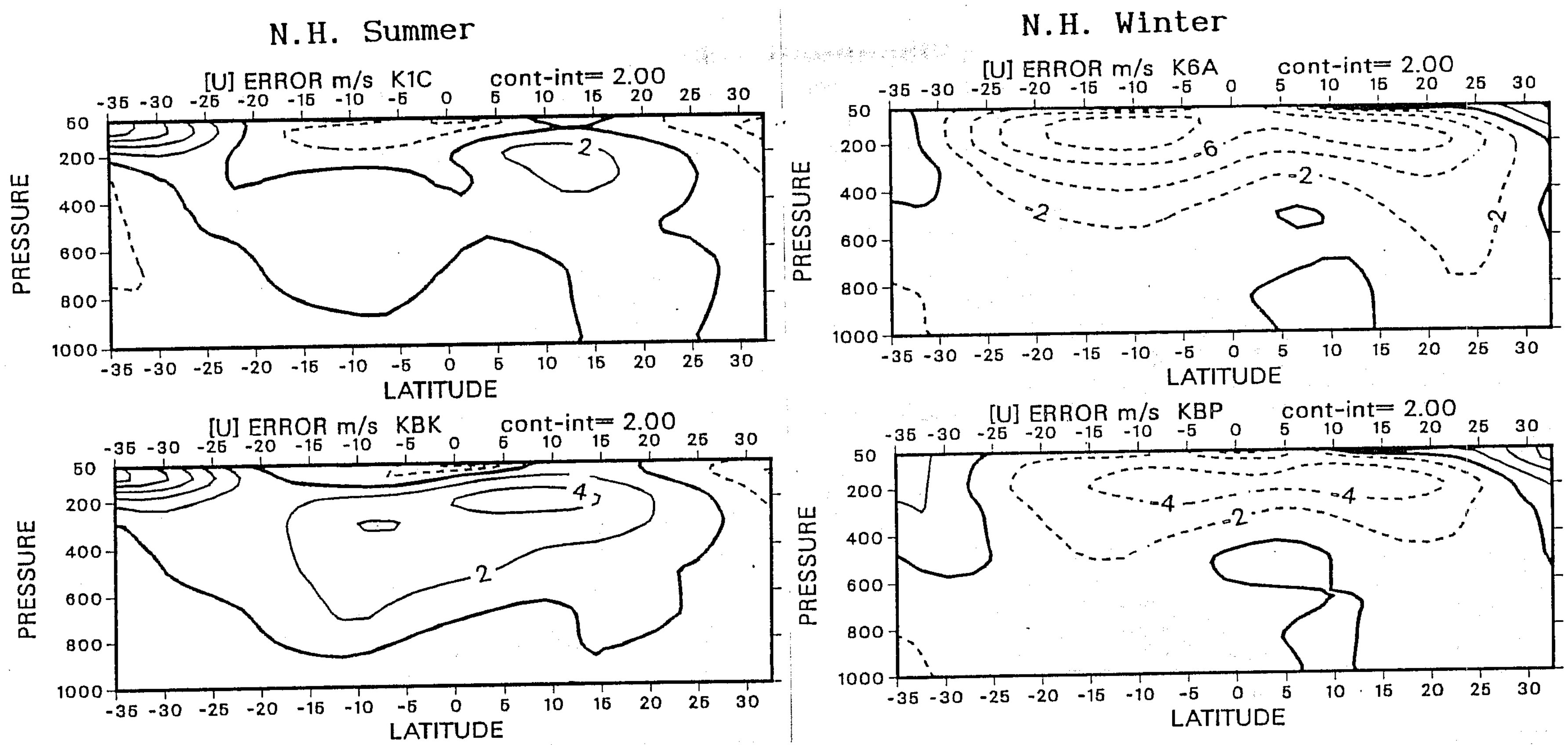

Fig. 9 Zonally averaged U-errors for the 90-day control experiments (CY34; upper panels) and the errors with the interpolation scheme (lower panels).

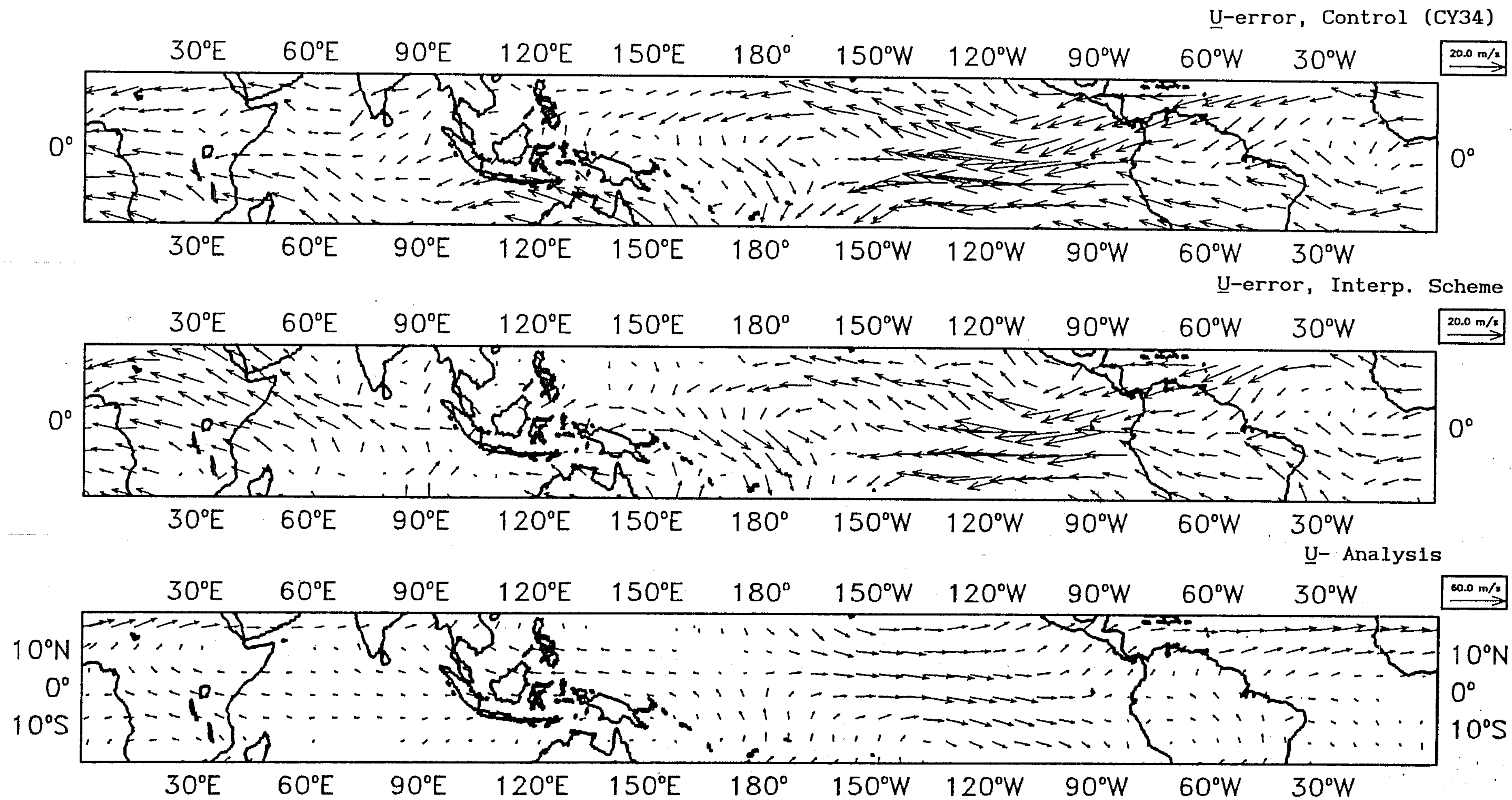

Fig. 10 Mean wind errors (difference between forecast and analysis, averaged over 90 days) at the 200 mbar level for the T42 90-day NH winter experiment. Results for the control experiment (CY34) are shown in the upper panel; results with the interpolation scheme in the middle panel. The lower panel shows the wind pattern as analyzed by the ECMWF operational model.

- The stability effects on the transfer coefficients are usually based on Monin-Obukhov functional relations as established over land. Although there is little reason to believe that the stability effects are different over sea, Large and Pond (1982) find large differences between stable and unstable data points after stability correction. Because the Large and Pond data set is one of the very few this needs further investigation.

- With model sensitivity as demonstrated above, it is probably necessary to distinguish between the skin temperature of the ocean and the bulk temperature. This means that the skin temperature needs to be parametrized in terms of the energy budget of the sea surface (Robinson et al., 1984). Support from oceanographic experiments is essential.

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