INTRODUCTION TO THE PARAMETRIZATION OF PHYSICAL PROCESSES AND PRESENTATION OF BOUNDARY LAYER ASPECTS

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This contribution is designed both as an introduction to the following papers on various aspects of physical parametrizations (Tiedtke, Miller) and as a specific presentation of developments in boundary layer aspects.

Part I Introduction to developments in physical parametrization

A numerical weather prediction model contains a number of approximations used for the representation of the complex ensemble of processes taking place in the atmospheric system. It is convenient to distinguish between the approximations built in the basic set of equations, the so-called primitive equations adopted in most large-scale models, and the parametrization techniques. The expression "parametrization" generally refers to the ensemble of methods used to express the effect on the model variables of physical processes not resolved (or not entirely resolved) by the model in function of parameters resolved and predicted by the model. The main goal is then to provide a correct feedback from subgrid scales to resolved scales.

In a spectral model, a natural scale boundary between processes explicitly computed by the model and those which have to be parametrized is half the smallest computed wavelength, that is approximately 200 km near the equator with a T106 resolution. Similarly, in the present configuration of the ECMWF model, one is unable to deal with events taking place at time scales smaller than twice the timestep of the dynamics (that is 30 minutes at T106 resolution). Parametrization schemes should then typically deal with processes which occur at space and time scales appreciably smaller than the above limits. This includes turbulent processes which take place at the lower atmospheric boundary, most turbulent and convective processes taking place within the atmosphere, and a large fraction of internal waves from orographic or convective origin. The parametrization of meso-scale processes which occur at scales smaller or of the same order as the model gridscale is certainly also of importance but the problem has not received sufficient attention up to now. In addition to the parametrization of thermodynamic and dynamic exchanges, the formulation of radiative fluxes has to take into account both grid-scale and subgrid scale aspects.

The examination of a schematic diagram of the atmospheric system (Fig. 1) enables to identify the main processes an atmospheric model should represent or parametrize. The units used here are in percent of the incoming flux of solar radiation at the top of the atmosphere, they should be multiplied by 3.4 to convert in W/m^2 (assuming a solar constant of 1360 W/m^2). The main energy reservoir in the atmosphere corresponds to the internal and potential energy fed partly by radiative input (solar and infrared radiation absorbed by the atmosphere) and partly by direct heat input. This heat input comes in major part from the latent heat release (24% or 81.6 W/m^2) and to a lesser extent from the direct input of sensible heat at the lower boundary (6% or 20.4 W/m^2). The evaluation of those terms is, in the model, strongly determined by physical parametrization schemes. The surface sensible heat flux has to be entirely parametrized as well as its redistribution within the atmosphere. The latent heat release occurs partly at the resolved scale (gridscale condensation) but approximately three quarters of it in our present model take place at subgridscale (as part of the convection scheme). As ultimately the hydrological cycle of the atmosphere should be in balance in a sufficiently long simulation, the total latent heat release is determined by the evaluation of the surface latent heat flux, which is also entirely parametrized.

Another branch of the global energy cycle, which deals with terms of lesser magnitude than the previous ones (about 1% of the total solar input), is the kinetic energy cycle. Its accurate representation is of course one of the main requirements of an atmospheric model, and deficiencies, for example in the level of global kinetic energy, are an important subject of preoccupation for modelers. The terms of the kinetic energy budget are, on one hand the generation of kinetic energy which occurs mostly at computed scales but is also dependent on parametrization aspects (for example spatial distribution of convective heat sources), on the other hand the dissipation by turbulence at the boundary and within the atmosphere. This second part has to be entirely parametrized, mostly by the vertical diffusion scheme, but contributions are also made from the gravity wave drag and the horizontal diffusion schemes (cf M. Miller, these proceedings).

The above components illustrate the major role of parametrization schemes for an adequate representation of the atmospheric evolution as soon as one deals with time scales larger than 1-2 days. For example the recycling time scale of kinetic energy in the atmosphere (ratio of global kinetic energy to its generation rate) is of the order of 5-6 days, and the recycling time scale of water vapour of the order of 10 days. Global budgets of the various parts of the atmospheric energy cycle are good indicators of the performance of the model physics. They should be judged by the equilibrium level of the various terms and also by the way initial imbalances evolve within the first few days of a forecast. This is the so-called spin-up problem discussed, for example, in the paper by M. Tiedtke (these proceedings).

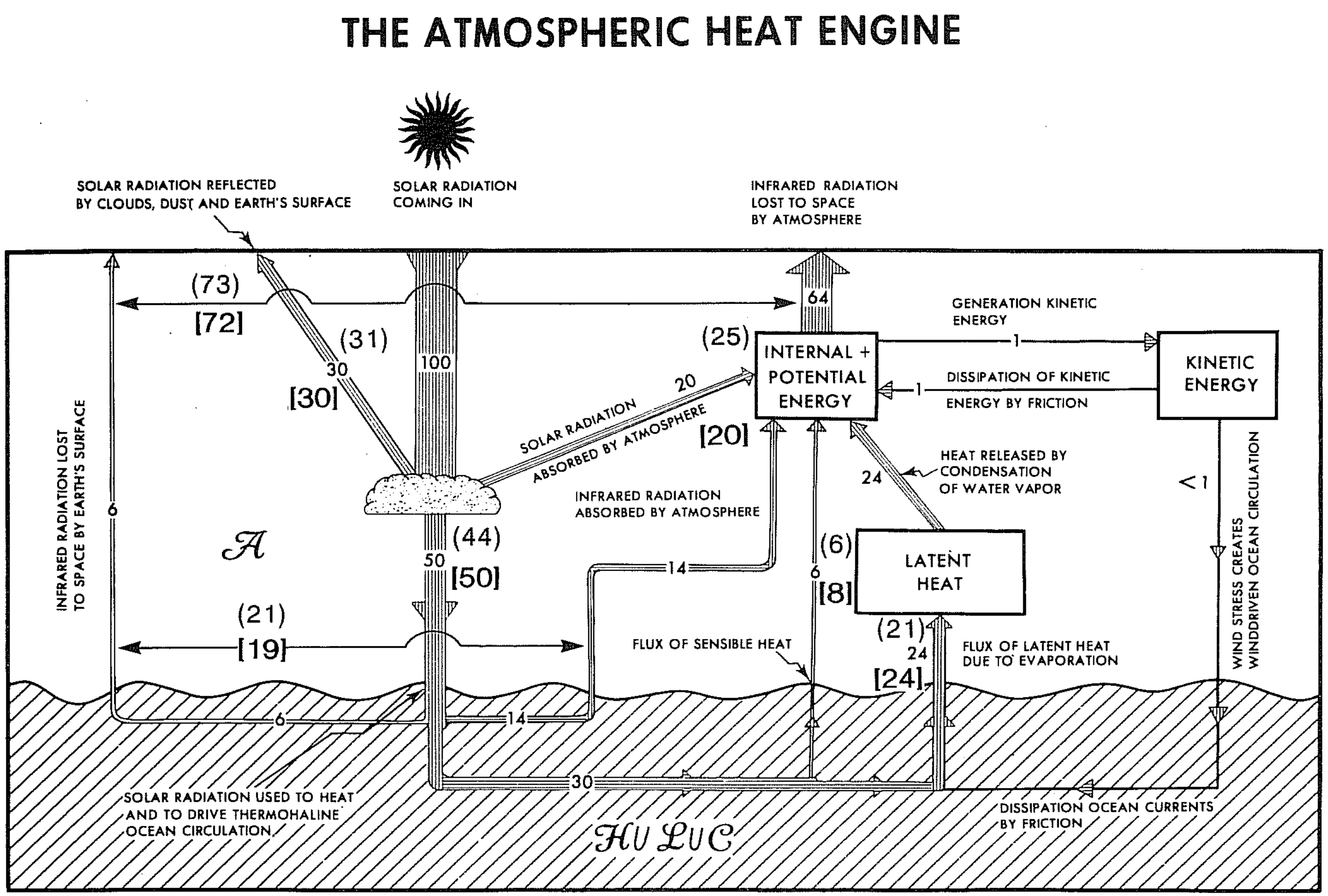

Fig. 1 Schematic diagram of the flow of energy in the climatic system. A value of 100 units is assigned to the incoming flux of solar energy. Figures have to be multiplied by 3.4 to obtain units in W/m², assuming a solar constant of 1360 W/m². All values represent annual averages for the entire atmosphere (from Peixoto and Oort, 1984). Figures in parenthesis are model values for Northern Hemisphere summer 1988 and in brackets for summer 1989 with the new physics.

The importance of the quality of parametrization schemes for medium-range weather forecast had been realized early in the development of the ECMWF model. We will recall here the main evolution of the physics schemes in the recent years, since the introduction in operation of the first spectral model in 1983 (Table 1). At that time the physics package had already benefited from a number of important developments compared to other models and the "ECMWF physics" was fairly advanced. Most of its features were already implemented in the first operational model in 1979. Among others, it contained a stability dependent vertical diffusion formulation including a Monin-Obukhov surface layer scheme following J.F. Louis (1979), a convection scheme based on the Kuo (1974) formulation, and a radiative scheme taking into account clouds computed diagnostically by the model.

The main stages of development are mentioned in Table 1. The introduction of the diurnal cycle was associated with a three layer representation of land surfaces (as described below) and marked a step towards a more realistic physics. The impact tested at the time was mostly seen on local weather elements, but more recent experiments seem to indicate that the impact on the general circulation with the more recent physics is not negligible.

A revised infra-red radiation scheme based on the representation of transmission functions by sums of exponentials (Fast Exponential Sums Filtering technique) was implemented in December 1984 to correct biases observed in the stratospheric temperature field. In May 1985, simultaneously with the implementation of the higher resolution model (T106), three major physics changes took place, a revision of the deep convection scheme, the implementation of a shallow convection scheme and a modification of the cloud formulation. This had a large impact on the tropical circulation, improving the boundary layer thermal structure, increasing the hydrological cycle and reinforcing the Hadley circulation and subtropical highs. The impact on tropical and temperate Northern Hemisphere scores was probably the largest seen in the ECMWF forecast history. The implementation of a parametrization for gravity wave drag (July 1986) following the increase of stratospheric resolution led to a significant improvement of the model climate. It was complemented in January 1988 by a removal of most of the vertical diffusion above the boundary layer (M. Miller, these proceedings). The implementation of a fractional vegetation scheme (April 1987) is discussed below.

The latest major physics change (May 1989) is discussed in detail in the papers by Tiedtke, Miller and Morcrette (all in these proceedings): it involves the introduction of a new convective scheme based on the mass flux approach, of a new radiation scheme similar to the one developed at the University of Lille, and a revision of the gravity wave drag scheme. The effect of those changes, studied during a series of 13 parallel analyses and forecasts in April 1989, is a further improvement of the temperature and moisture structure

MAIN PHYSICS CHANGES 1983-1989

SPECTRAL MODEL (T63) ENVELOPE + 'ECMWF' PHYSICS –Stability dependent vertical diffusion –'Kuo' convection –Interactive clouds	April 1983
DIURNAL CYCLE	May 1984
REVISED RADIATION (FESFT)	December 1984
REVISED PHYSICS (T106) –Revised 'Kuo' –Shallow convection –New cloud form	May 1985
(19 levels)	May 1986
GRAVITY WAVE DRAG	July 1986
FRACTIONAL VEGETATION	April 1987
REVISED VERTICAL DIFFUSION	January 1988
REVISED PHYSICS –Mass flux –'Lille' radiation –Revised gravity wave drag	May 1989

Table 1

in the tropics, leading to an increase of the thermal energy cycle (around 20%) and of the hydrological cycle (around 20%) and a more realistic and permanent Hadley circulation. Tropical winds at low level are improved, although errors in the upper level winter remain high. The terms of the atmospheric dynamical energy cycle are also improved to a higher and more realistic level. Another important impact is the decrease of the spin-up of the hydrological cycle corresponding to a decrease of the precipitation excess during the first 2-3 days of the forecast.

The developments of the physics schemes were guided all along the past ten years and are still guided by three major preoccupations: to improve the representation of physical processes and validate results on selected case studies, to improve the forecast's systematic errors by going somehow backwards to understand which formulation aspects led to specific deficiencies, and finally to improve the forecast of weather elements. Those aspects are discussed in Part II of this paper and in the following papers, but two examples have been chosen here as a matter of illustration. The first one is the evolution of the systematic error of the temperature field at day 5 for the various winters 83/84 to 86/87 (Fig. 2) and for the recent parallel forecasts (Fig. 3).

The upper tropospheric and stratospheric error has been first decreased from winter 83/84 to winter 84/85 due to the change of radiation scheme which improved the representation of upper cloud effects. The large negative systematic error seen in most of the tropical troposphere during winter 84/85 and previous ones is largely decreased in and after winter 85/86, due to the 1985 physics change. One remains until 1988 with a slight negative temperature bias in the lower troposphere and a positive bias of the order of 1-15 deg/day in the tropical upper troposphere and stratosphere. A large part of this positive bias (which resulted in a too stable atmosphere with convective activity decreasing with time) is removed by the 1985 physics (Fig. 3) due to the combined effect of the new radiation and convection scheme.

Another example deals with the forecast of 2 meter temperature, T2m (Fig. 4). The diagram compares the forecasted T2m with the observed one, taking all 48 hour forecasts from the 13 parallel runs mentioned above and using all available European stations. The distribution of T2m forecasts is clearly translated towards warmer temperatures but is also better centered compared to observations. Throughout the model history the computation of T2m, as an example of weather element, has been affected by several physics changes, mainly the diurnal cycle, the vegetation parametrization, the last radiation change and, to a lesser extent, modifications of the convection schemes. Its computation is now based on sounder physical formulations which however may still lead to significant discrepancies in cases where at least one of the processes involved is not correctly taken into account. It should be noted that, after operational implementation of the 1989 physics, T2m diurnal maxima

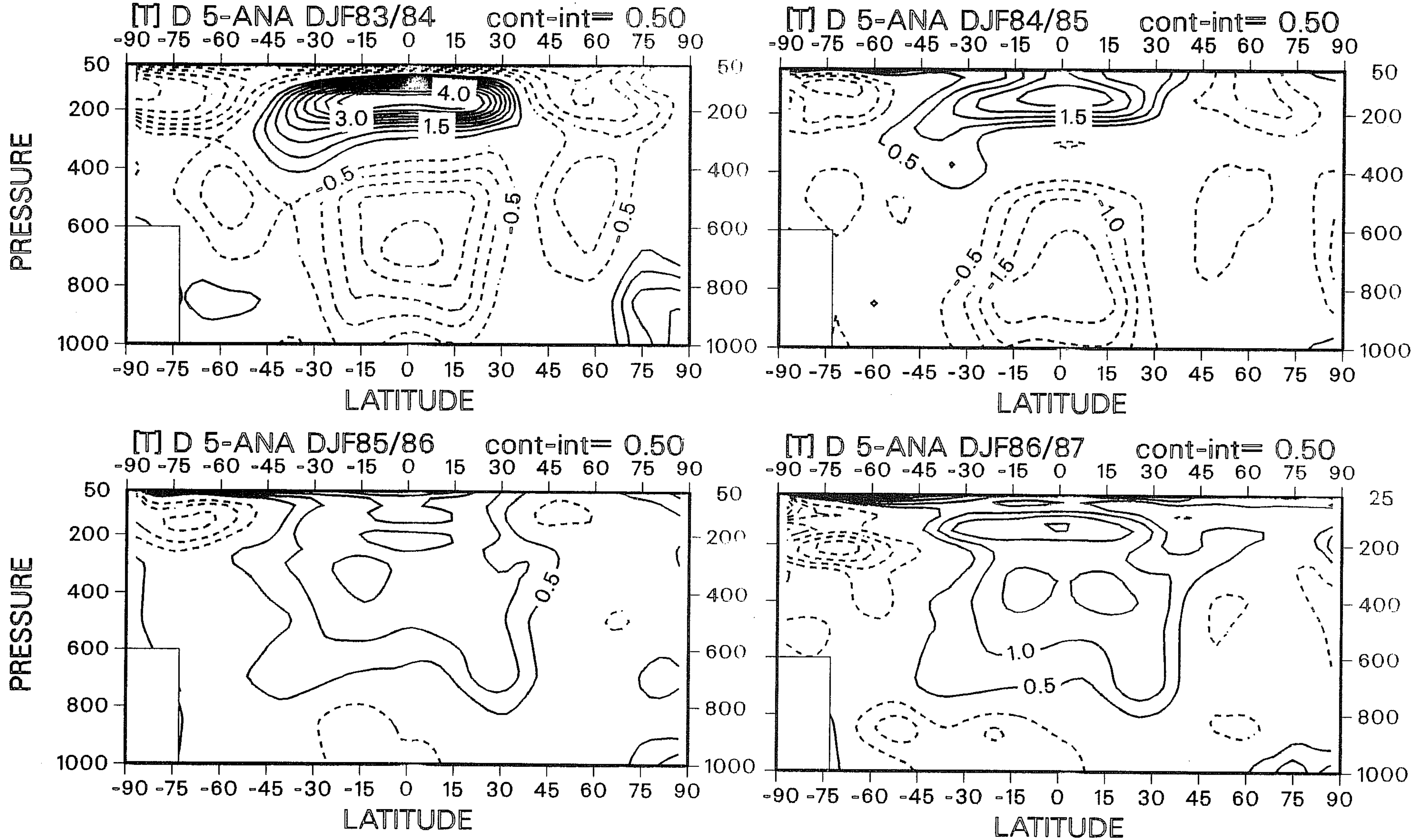

Fig. 2 Systematic error of the temperature field at day 5 averaged for all operational forecasts in December-January-February for winter 83/84, 84/85, 85/86 and 86/87 (from C. Brankovic and K. Arpe).

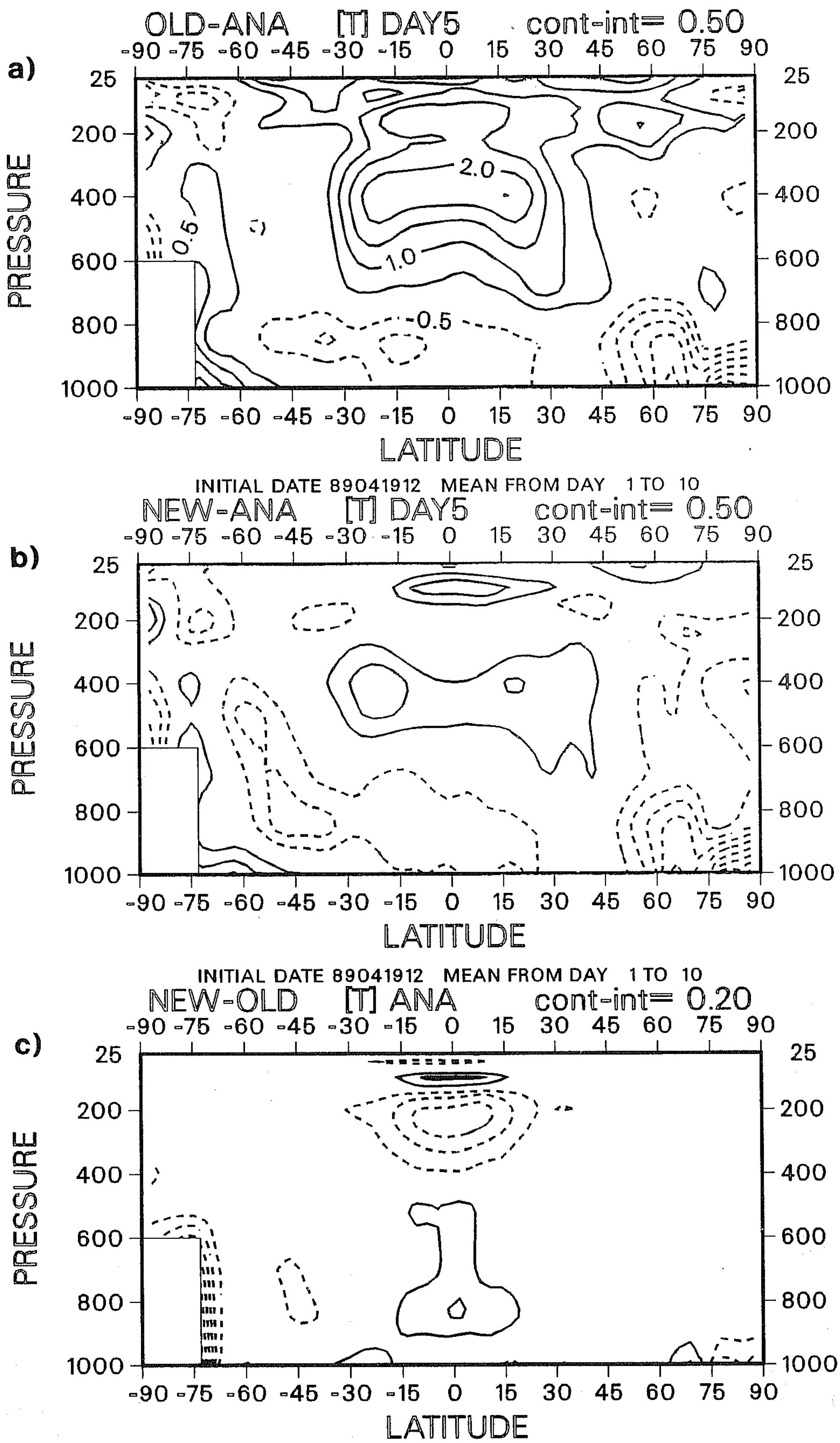

Fig. 3 Systematic error at day 5 averaged over 13 parallel forecasts prior to the implementation of the 1989 physics (19 April - 1st May 1989). a) with the control physics, b) with the new physics, c) systematic difference between the new and the old analysis (from K. Arpe).

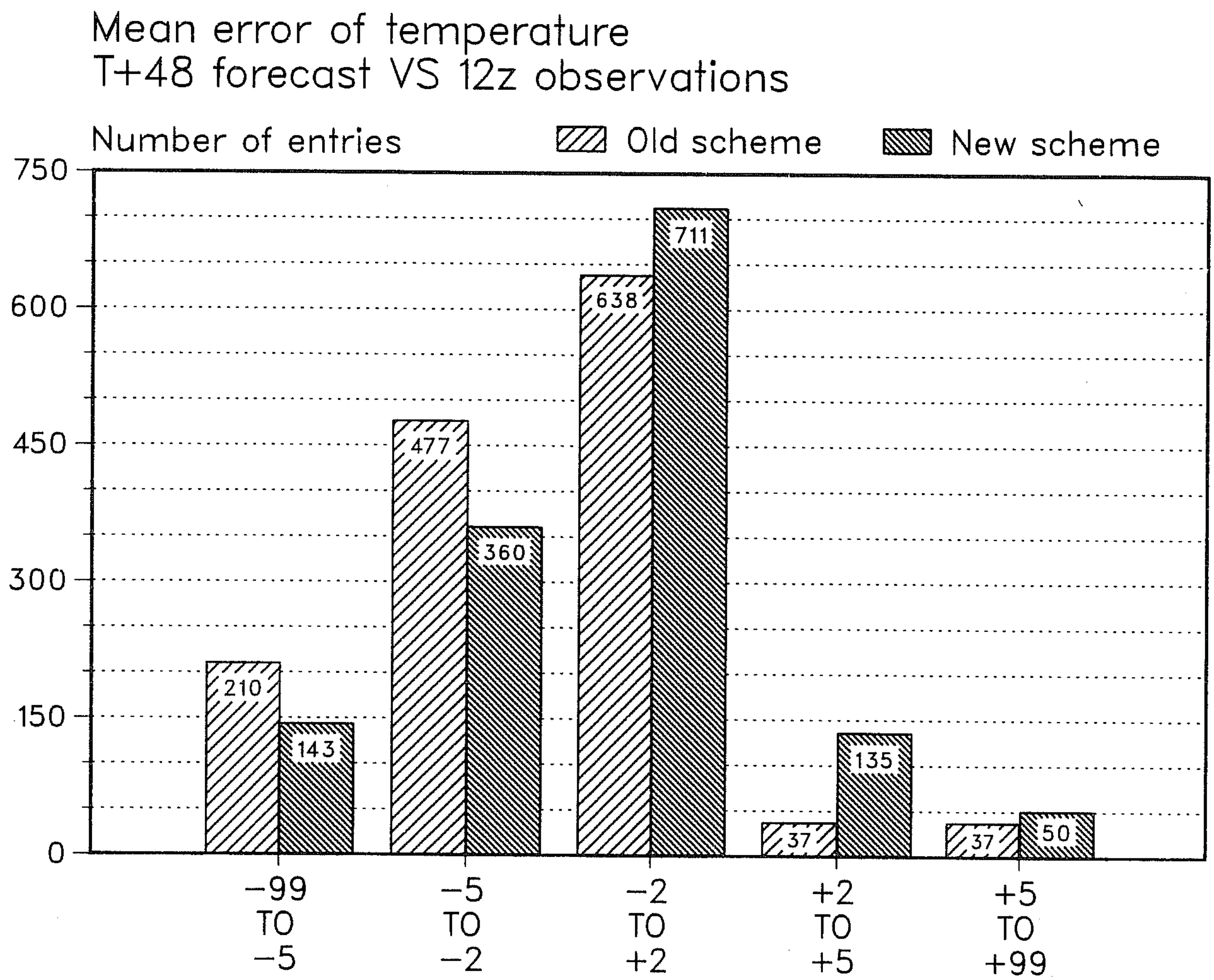

Fig. 4 Distribution of mean T2m temperature errors for all European stations (average for the 13 parallel forecasts mentioned above).

have displayed a systematic positive bias over temperate continents, which is under investigation at the time of writing. Possible contributions for this positive bias are an underestimate of low level clouds, of aerosol effects and of surface albedo.

The summary given in the above pages has hopefully shown that progress has been made in the last ten years in the formulation of physical processes. The most striking advance has been towards the decrease or removal of basic systematic errors in the forecast at various ranges, although important errors remain regionally and zonally and are the object of special efforts for on-going and future research as is discussed in the following papers. Another guideline for future developments is the improvement of forecast products which were, up to now, considered as experimental, but which become more realistic as a by-product of improvements in physical parametrization. This is particularly the case for precipitation, cloud and near surface temperature forecast.

In addition, a greater variety of non-conventional data, often derived from satellite observations, will be used as initialization and validation data, requiring reliable formulations of the relevant physical processes.

Part II Parametrization of surface fluxes and boundary layer aspects

1. INTRODUCTION

This part reviews the state of the surface and boundary layer (PBL) parametrization scheme in the ECMWF model, describes recent developments and provides indications on future trends. Recent developments are mainly related to the introduction of the properties of a vegetation cover in the land surface scheme and on-going research focuses on the parametrization of boundary layer clouds.

If one goes back to the general motivation for a PBL parametrization scheme it can be presented as threefold:

(1) To provide estimates of vertical fluxes of momentum, heat and water vapour in function of the grid-scale field of variables computed by the basic model equations. This includes surface fluxes and their distribution with height, generally computed at levels intermediate between main model levels.

(2) To provide an evaluation of the effect of subgrid processes on the computed field of variables. This is indeed the primary goal of a parametrization scheme. The source terms for the model variables are simply expressed as minus the vertical derivatives of the parametrized fluxes. However the interaction of the PBL scheme with other parametrized processes, such as convection, condensation, radiative effects may have to be taken into account.

(3) To provide forecasts of weather elements which are closely linked to boundary layer properties, namely 2 metre temperature and dew points, 10 m wind, low level cloudiness or other parameters which can be used for model validation, such as surface temperature and fluxes. This aspect, which is not a primary goal of parametrization schemes, becomes increasingly important as the forecast quality improves and weather elements become useful forecast products.

The present paper includes first a general description of current PBL parametrization schemes based on experience at ECMWF, followed by a more detailed account of two important aspects, the representation of surface fluxes over vegetation and of turbulent fluxes in cloudy boundary layers.

2. CURRENT BOUNDARY LAYER SCHEMES

If one excludes simple approaches based on drag coefficients, a PBL parametrization scheme contains two main components: the computation of surface fluxes and of the way those fluxes are distributed with height.

a) Surface fluxes

A prerequisite for the computation of surface fluxes is a specification of surface properties, as is done over the ocean, or their forecast, as is done over land. Over the ocean, temperature is currently specified as its analysed value throughout the forecast, and the surface specific humidity is taken as the corresponding saturation value. The roughness length is assumed to be proportional to the mean surface stress following Charnock's formulation, with a minimum value for weak winds. Tests are currently being made at ECMWF to evaluate the impact of a more advanced stress formulation which would take into account the effect of the wave field, as forecasted by a global wave model.

Over land, surface variables are usually computed through budget equations involving the exchanges of thermal energy and water between a thin surface soil layer and the atmosphere, with some account of exchanges with underlying soil layers. In the ECMWF model, a three-layer land-surface configuration has been implemented in December 1982 as a prerequisite for a realistic description of the diurnal cycle. The thermal energy budget for the first soil layer (Fig. 5) involves the net input of solar energy $(1-\alpha_g)\ S\downarrow$, the net infra-red radiative flux (the incoming flux from the atmosphere $\varepsilon_g\ R_L\downarrow$ minus the surface thermal emission $\varepsilon_g \sigma T_s^4$) the turbulent exchanges of sensible and latent heat (F_H and LF_W) and the diffusion of heat into the soil (F_G). In those expressions α_g is the surface albedo, ε_g the surface emissivity, σ the Stefan-Boltzmann constant, L the specific latent heat for evaporation, T_s the surface temperature. The above budget is applied to a soil layer of density ρ_G, specific heat capacity C_g and depth Δz_1 (equal to 7 cm).

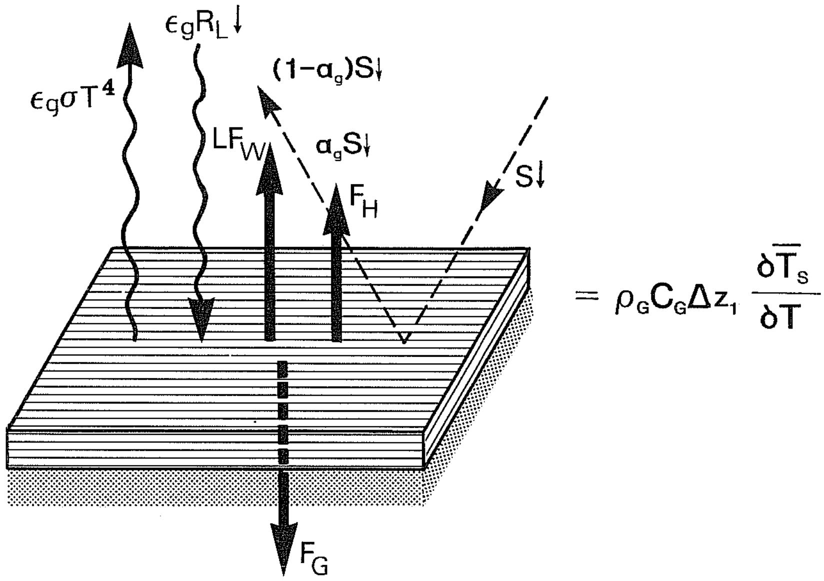

Fig. 5 Schematic representation of the land-surface energy budget.

As for the surface water budget, it is convenient to consider separately the budget for liquid water and snow. The liquid water input is by liquid precipitation and snow melt, and this is compensated by percolation into the ground, run-off, and by the turbulent surface moisture flux over snow-free areas. The snow mass per unit surface S_n evolves through the effects of solid precipitation, snow melt and the fraction of surface moisture flux which corresponds to sublimation over snow covered areas. The result of the land surface prametrization is the forecast of the three variables T_s, S_n and surface soil water content W_s which can be expressed as a water depth within the upper soil layer. The computation is dependent on a number of parameters which have to be defined at each model grid-point, namely α_g, ε_g, ρ_g, C_g defined earlier, as well as the diffusivity properties of the soil for heat and water and the soil maximum water capacity. For lack of adequate data, the only parameter which is geographically dependent in snow-free conditions is the albedo, although a seasonal and soil moisture dependency are also considered for future model versions. The treatment of snow-covered land is the object of a special computation which is not detailed here. Whereas an adequate land-surface parametrization is obviously crucial for the computation of near surface weather elements, it is also important for the evaluation of surface turbulent fluxes. One of the key factors there is the ratio between latent and sensible heat, determined mainly by the evapo-transpiration formulation and the way vegetation effects are taken into account.

Once the surface variables are computed, surface fluxes can be expressed as a function of those and of the model variables at the first level (z_1). A general hypothesis is that the first model level (approximately 30 metres for the ECMWF model) lies within the surface layer or "constant flux layer" of the atmosphere for which Monin-Obukhov similarity laws apply. Typical expressions for surface fluxes are then:

$$-\overline{(u'w')}_o = a^2 \, |U_1|^2 F_m \left(\frac{z_1}{z_o}, R_{iB}\right) \qquad (1)$$

$$\frac{F_H}{\rho \; C_p} = \frac{a^2}{R} \, |U_1| \; \Delta\theta \, F_h \left(\frac{z_1}{z_o}, R_{iB}\right) \qquad (2)$$

$$\frac{F_W}{\rho} = \frac{a^2}{R} \, |U_1| \; \Delta q \, F_h \left(\frac{z_1}{z_o}, R_{iB}\right) \qquad (3)$$

where $\overline{(u'w')}_o$ is the surface momentum flux, ρ the density and C_p the specific heat capacity of the near surface air, a^2 and $\frac{a^2}{R}$ constants corresponding to drag coefficients in neutral conditions, $|U_1|$ the velocity at first model level, F_m and F_h "universal" functions of two dimensionless parameters which characterize the first model level depth (z_1 divided by a roughness length z_0) and its thermal stability (bulk Richardson number R_{iB} for the first model layer). $\Delta\theta$ is the difference between the potential temperature at the surface and at first model level. An upper limit for Δq is the difference between the saturating specific humidity at the surface and the specific humidity at the first model level, providing a value for the so-called "potential" evapo-transpiration. The formulation of Δq will be discussed below with respect to the evaluation of vegetation effects.

The functions F_h and F_q are derived from local field experiments and the expressions derived by J.F. Louis (1979) from the Kansas experiment data and are used in several models. Revised expressions derived from experimental data obtained at Cabauw (The Netherlands) are being presently tested at ECMWF in order to improve the formulation, leading mainly to smaller fluxes in cases of stable stratification. A particular problem is the specification of z_0 at the grid scale, discussed for example in André and Blondin (1986). In the present ECMWF model, z_0 includes the contribution from the vegetation and is greatly enhanced over mountainous areas. A new formulation is being tested where the values of z_0 for heat and moisture will be different from the z_0 used for momentum, with smaller values for the vegetation contribution and no orography contribution.

b) Fluxes within the planetary boundary layer

It is convenient to distinguish between multi-level schemes where fluxes are represented by a simple flux-gradient relationship (eddy-coefficient type), multi-level schemes where more or less advanced turbulence closure methods are applied to relate fluxes to grid-averaged variables (generally based on a turbulent kinetic energy formulation) and bulk schemes where a schematic flux distribution is assumed a priori and the key problem is the representation of the interface between the PBL and the free atmosphere. Another method where fluxes are defined not only from nearby model layers but also include some contribution from layers further apart (so-called transilient method) has been applied to PBL modelling (Stull and Driedonks, 1987) but not, to my knowledge, to forecast models. Whereas most methods of parametrization have been extensively tested for clear boundary layers, (especially the classical case of mixed layer under an inversion), formulations for cloud-topped boundary layers are still at a research stage, as will be discussed further in Section 4.

The eddy diffusivity formulation, presently the most commonly used in forecast models, states that the vertical sub-grid turbulent fluxes are proportional and of opposite sign to the vertical derivatives of the corresponding mean quantities. The eddy diffusion coefficients usually depend on height, shear and thermal stability. As an example, the formulation used in the ECMWF model is similar to the one described in Louis (1979):

$$\overline{w'x'} = -K_x \frac{\partial \overline{X}}{\partial x} \qquad (4)$$

where X is any of the following variables, horizontal components of velocity, dry static energy $s = C_p T_v + gz$, or specific humidity q and where K_x are corresponding eddy coefficients. The overbar denotes an averaging over the model grid. The form chosen for K_x is derived from dimensional considerations and a schematic interpretation of the Turbulent Kinetic Energy (T.K.E.) budget:

$$K_x = c_x \ell^2 \left|\frac{\Delta v}{\Delta z}\right| F_x(R_i) \qquad (5)$$

where c_x is a constant, $|\frac{\Delta v}{\Delta z}|$ the absolute magnitude of mean shear between two model levels, F_x a function defined from the similarity theory and R_i the mean Richardson number between the two model levels. One of the key problems is the estimate of the characteristic length scale, ℓ, used in this type of formulation. The present choice at ECMWF is the so-called "Blackadar" formulation:

$$\ell = \frac{\kappa z}{1 + \kappa z/\lambda} \qquad (6)$$

where λ is a parameter which may be different for momentum and heat or moisture, and which corresponds to an asymptotic value of ℓ, and κ the von Karman constant.

In forecast models the PBL parametrization is often used throughout the atmosphere as a vertical diffusion scheme which represents clear air turbulence or small scale vertical transports associated with clouds and orography. This may require a different formulation for ℓ near the top and above the boundary layer. As discussed in the paper by Miller, the vertical diffusion above the boundary layer has been reconsidered after the implementation of the gravity wave drag in the ECMWF operational model. Since January 1988 there is no diffusion above a diagnosed boundary layer height, except to remove dry instability.

The second type of multi-level parametrization attempts to better represent some of the physics involved in the turbulent exchanges. Whereas turbulence closures up to the third order (Andre et al., 1978), have been used for detailed PBL modelling, parametrization schemes for forecast models usually limit themselves to the use of a prognostic equation for turbulent kinetic energy. The formulation first tested in the GFDL model by Miyakoda and Sirutis (1977), based on Mellor-Yamada (1974), is the following:

$$\frac{D\bar{e}}{Dt} = \frac{\partial}{\partial z}\left(K_e \frac{\partial \bar{e}}{\partial z}\right) - \overline{u'w'}\frac{\partial \bar{u}}{\partial z} - \overline{v'w'}\frac{\partial \bar{v}}{\partial z} + \beta g\, \overline{w'\theta'_v} - c_\varepsilon \frac{\bar{e}^{\,3/2}}{\ell_\varepsilon} \tag{7}$$

where $\bar{e}$ is the grid-averaged T.K.E., K_e an eddy-coefficient chosen to represent the turbulent diffusion of T.K.E. and the pressure correlation term, c_ε a constant used for the expression of molecular T.K.E. dissipation, ℓ_ε a characteristic length scale prescribed from empirical considerations. The resulting expressions for turbulent fluxes, such as $\overline{u'w'}$, $\overline{v'w'}$, $\overline{w'\theta'_v}$, computed with the help of their corresponding diagnostic equations, are complex functions of $\frac{\partial \bar{u}}{\partial z}$, $\frac{\partial \bar{v}}{\partial z}$, $\frac{\partial \bar{\theta}}{\partial z}$, ℓ_ε and $\bar{e}$.

Modified formulations have since been proposed by various authors, such as Therry and Lacarrère (1983) Mailhot and Benoit (1982), with slightly different sets of approximations, and used for example in the French Peridot model (Bougeault and Lacarrère, 1989) and the Canadian regional model (Staniforth and Mailhot, 1988). The turbulent kinetic theory is then mainly used to scale the eddy coefficients with expressions such as:

$$K_x = c_x \ell_x \bar{e}^{\,1/2} \tag{8}$$

and fluxes are derived from expressions similar to (1), but including the possibility of counter-gradient values, especially for heat.

The third approach, so-called bulk parametrization, does not attempt to resolve the PBL level by level but, assuming a typical PBL structure, focuses on the computation of fluxes at the surface and at the PBL top. It usually relies upon two main features, a prognostic equation for the PBL height and evolution equations for the mean PBL properties. In the UCLA model where this method was first introduced following the work of Deardorff (1972), the first model level is chosen to be at the top of the PBL, and is time-dependent (Suarez et al., 1983). This overcomes the problem of having to forecast variables within the PBL and which would, at the same time, have to be solutions of the model equations and to verify the bulk PBL assumptions. Surface fluxes are computed from a "bulk" similarity formulation, using bulk variables instead of surface layer variables. A delicate problem is the computation of entrainment, made from a series of hypotheses related to the energy budget of the PBL.

Both multi-level schemes, eddy diffusivity and T.K.E. formulations, attempt to resolve the PBL vertical structure with the vertical resolution of the model, that is in practice with a few model levels within the first 2000 m, which is the typical height of the PBL. This inevitably crude resolution tends to annihilate the advantages of a sophisticated method of parametrization. Tests made at ECMWF by Manton (1983) have shown that the vertical truncation error on the T.K.E. field in a 15-level G.C.M. leads to spurious vertical diffusion, which destroys the precision expected from the method. A vertical resolution of approximately 100 metres is required to obtain satisfactory precision of the computation. Cruder methods like the eddy diffusivity approach are physically less satisfactory but can be adequately tuned within a given model. They may in fact include in one formulation several physical processes such as clear air turbulence and the breaking of internal gravity waves above the PBL. Their validity is, however, questionable in some circumstances which should allow for counter-gradient fluxes and they tend not to preserve sharp temperature structures, such as inversion layers. Another problem noted with local eddy-diffusion schemes is their inability to represent entrainment at the interface between turbulent and stable layers. As shown in experiments done by L. Dümenil (1987) with the ECMWF grid point model, this is overcome in a T.K.E. scheme by the vertical diffusion of kinetic energy. This may indeed be one of the main advantages of T.K.E. formulations compared to first order ones.

A bulk scheme is inexpensive with respect to computer resources and is designed in principle to provide the basic information, surface fluxes and fluxes at the upper interface. As it is based on an ideal PBL model, it is well adapted for cases which do not depart too much from it, typically a mixed layer under a temperature inversion, frequently observed over the oceans and in the day-time over land. It is in general adapted to undisturbed situations, in the absence of large temporal or spatial variations. It however requires specific assumptions, for example at the day-night transition, and has to be replaced by a simple diagnostic scheme in the case of stable boundary layers. Another advantage is that the PBL equations appear as a budget computation which does not suffer from computational truncation errors unavoidable from a diffusion scheme. The kinetic energy budget can provide a way to compute entrainment at the top of a turbulent layer. Some suggestions have been made to use this type of entrainment formulation as a complement to local diffusion scheme for example in the framework of an eddy-coefficient formulation.

3. RESEARCH ON THE REPRESENTATION OF SURFACE FLUXES OVER VEGETATION

The sensitivity of numerically simulated climates to the formulation of evaporation over land has been a subject of growing interest since the work of Mintz (1984). This lead to the development of advanced (or at least complex) schemes to represent the properties of the vegetation cover in GCMs, such as BATS (Biosphere Atmosphere Transfer Scheme, Dickinson et al., 1986) and SIB (Simple Biosphere Model, Sellers et al., 1986). Less work has been devoted to the sensitivity of global atmospheric simulations at short or medium-range to land-surface conditions. A first study on this subject has been done by Rowntree and Bolton (1983), demonstrating the impact of persistent soil wetness anomalies on atmospheric moisture and surface precipitation at a 2-3 day time scale.

Problems were noted in the ECMWF precipitation forecast at the time of implementation of the 1985 physics and were partly related to an excess of evaporation over land. A series of studies were then undertaken to develop a vegetation scheme for the ECMWF model. Although it does not contain all the features of the advanced schemes mentioned above, it takes into account the basic properties of a vegetation cover, namely the regulation of evapo-transpiration in function of meteorological and soil conditions, and the ability of the root system to pump from underground layers the moisture available for evaporation. It can also be upgraded at a later stage to make use of global vegetation data sets.

The main features of the vegetation scheme implemented operationally in April 1987 are described in Blondin (1988) and summarized in Figure 6. Presently only two types of land-surface are considered, namely bare land and a standard canopy-type vegetation, the fractional vegetation cover being defined at each grid point from the Wilson and Hendersson-Sellers (1985) dataset.

Over the bare land fraction of the grid and in the absence of snow, the evaporation flux is obtained from equation (3) with:

$$\Delta_q = q_l - h\, q_{sat}(T_s, p_s) \tag{9}$$

where p_s is the surface pressure, q_{sat} the saturation specific humidity and h a relative humidity of the surface based on the value of the upper soil layer water content W_s. The main difference between this formulation and the one used at all land points in the previous version of the model is that the factor h is directly applied to $q_{sat}(T_s, p_s)$ instead of to the difference $q_l - q_{sat}(T_s, p_s)$. As a result the surface moisture flux is more sensitive to the surface soil moisture and gets to zero as soon as the relative soil water content reaches a value of the same order as the relative humidity in the atmospheric layer above.

Over the vegetated grid fraction, Δ_q is replaced in equation (3) by an expression of the type:

$$\Delta_q = \frac{R_a}{R_a + R_c}\,(q_l - q_{sat}(T_s, q_s)) \tag{10}$$

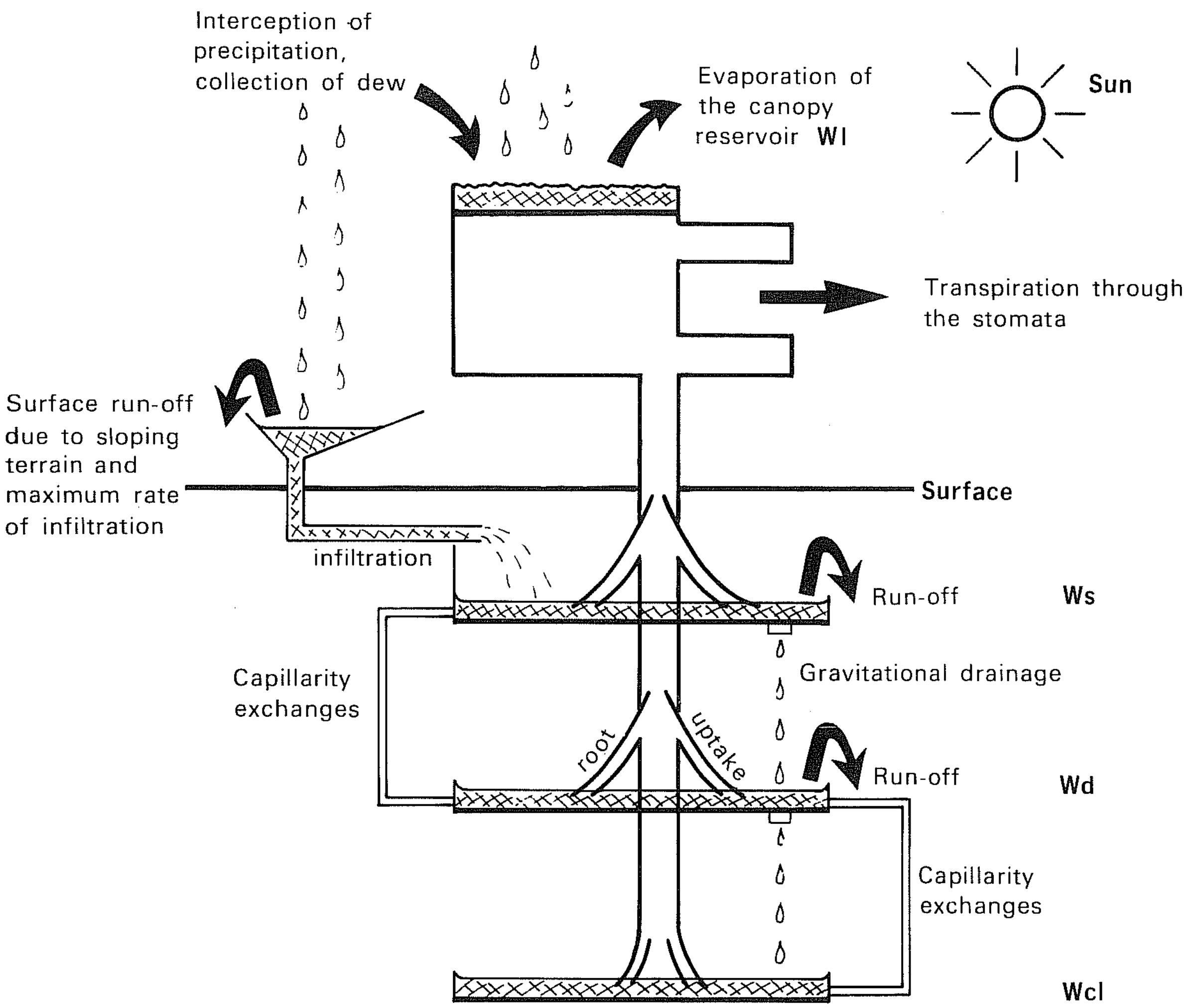

Fig. 6 Schematic representation of the vegetation scheme (from C. Blondin).

where the factor $\frac{R_a}{R_a + R_c}$ represents the regulation effect of the vegetation cover. R_a is the aerodynamic resistance of the first model layer (the inverse of the usual drag coefficient) and R_c the so-called canopy resistance, which characterizes the physiological control of the water loss by the vegetation. R_c should in principle contain the effect of vegetation properties and the way it responds to meteorological factors. The formulation chosen in the 1987 ECMWF scheme derived from Sellers (1985) is the following:

$$R_c = R_{co}\,(PAR)/F(W) \qquad (11)$$

where R_{co} characterizes the canopy response to the Photosynthesis Active Radiation (PAR) and F(W) characterizes the availability of moisture. In a more realistic vegetation scheme the function R_{co} should be a characteristic of the plant type, the leaf area index and meteorological conditions to which the plant may react, all factors which are not taken into account here. PAR is taken, also for simplicity, as 55% of the short wave net radiation flux. F(W) is 1 when there is enough water available in the soil ($W>W_{cr}$) and 0 if there is not enough water available for the plants to transpire ($W<W_{pwp}$). F(W) varies linearly between those two thresholds. W_{pwp} is the so-called permanent wilting point and W a weighted average of the water content of the three soil layers. The weighting factors R_s, R_d and R_{cl} ($R_s + R_d + R_{cl} = 1$) represent the distribution of the root densities within the soil layers. All the above constants are assumed to be the same on all vegetated areas.

The above relations describe the transpiration of the canopy. In addition the scheme allows for direct evaporation of liquid water deposited on the leaves, resulting from precipitation interception and dew collection. This is done by a "canopy" reservoir which intercepts a fraction (presently equal to .25) of the precipitations and evaporates at the potential rate over a fraction of the vegetation. Presently the upper limit for this reservoir is .8 mm of water depth.

The soil hydrology has also been modified to take into account both the diffusion of water by capillarity and the gravitational drainage. The computation of infiltration and run-off is also modified to take into account the increased run-off over subgrid orography and a reasonable maximum infiltration rate. As for the vegetation characteristics, no geographical dependence of the soil properties is considered.

A series of sensitivity tests showing the impact of this new surface scheme on local meteorological conditions and on the model climate has been performed; part of them presented in Blondin (1988). The local impact of the vegetation cover can be summarized as follows. In the absence of precipitation and with identical relatively high water content of the surface reservoir, the latent heat flux is decreased by the vegetation resistance leading to a larger diurnal cycle of surface temperature, especially a higher temperature maximum. On the contrary if the surface water reservoir is low, the vegetation parametrization pumps water from deeper layers leading to a higher latent heat flux and a smaller diurnal cycle than over bare soil. Overall the vegetation cover has a regulating effect on the diurnal cycle over periods of a few days to 1-2 weeks. In case of a rain event, the interception of rain by the canopy causes a rapid increase of the canopy evaporation, reducing the sensible heat flux and the surface temperature at the same time. The upper reservoir captures the infiltrated water but this water can be used efficiently to feed the transpiration only when it enters the intermediate reservoir. The increase of evaporation is thus spread over a few days, in addition to the short term outburst of evaporation from the interception layer.

Although the local impact of a land-surface scheme is relatively large, often leading to temperature differences of the order of 5 degrees in the first model layer, the effect on the atmospheric circulation in the medium-range is small. The main impact is observed on the various terms of the hydrological cycle and a lesser impact on the Hadley circulation itself. A clearer picture is given by longer range forecasts and a set of T42-90 day experiments have been carried out to assess the impact of the various features of the surface scheme. When the present scheme was implemented, it replaced a formulation where the moisture flux was proportional to the potential evapo-transpiration, with a proportionality coefficient function only of the water content of the superficial soil layer. This led to large moisture fluxes as long as the superficial layer was sufficiently moist, and to a fast local recycling of the precipitation water. The implementation of the scheme presented here resulted in a decrease of the latent heat flux over continents and consequently in a slight increase of temperature maxima. At the same time the occurrence and intensity of precipitation events were reduced, and the Hadley circulation slightly weakened.

If one now starts from the present scheme and studies the sensitivity of the model climate to some of the key factors, a relatively large response is obtained by changing the basic vegetation resistance and by removing the "interception reservoir". A higher vegetation resistance leads to a smaller evaporation, a higher sensible heat flux and higher temperature maxima. Suppressing the interception layer has a similar effect by suppressing the fast local recycling of water, and has been shown to be approximately equivalent to a multiplication by 10 of the stomatal resistance.

The example displayed here shows the sensitivity of the model hydrological cycle to the prescription of the vegetation cover. 30 day averages of the surface latent heat flux and of the total precipitation obtained during a T63 global forecast in northern hemisphere summer conditions (Figs. 7 and 9) are compared with the same fields in a simulation where the vegetation cover has been removed (Figs. 8 and 10). In this experiment the vegetated areas are simply replaced by bare soil with respect to the evaporation properties, no other change being made in the model in relation for example with the prescription of roughness lengths or albedo. The precipitation amount is reduced by 3.3% globally and by 16% over continental areas. The latent heat flux is reduced by 4.3% globally and 30% over continental areas. Main differences occur over the tropical belt, for example in Africa, where the extension of the moist convection region is greatly reduced.

The land-surface scheme presented here has been slightly modified after its operational implementation in April 1987 in response to deficiencies observed under specific climatic conditions. This led to a modification of the root distribution with a reduction of the input of evaporation water from the climate reservoir, thus a reduction of the climatic control of the surface moisture flux. At the same time the thermal diffusivity of the soil was reduced under a radiatively active vegetation cover, in order to simulate the shading effect of the canopy. This results in an increase of the diurnal temperature maxima, which previously remained often too low over temperate regions in summer.

Further revisions are presently under experimentation, including a complete cut of the evaporation contribution from the climate reservoir and a distinction of vegetation types in the resistance formulation. It will now include an atmospheric temperature and moisture dependency, characteristic of each vegetation type, enabling a better representation of the plant behaviour under various climatic conditions. At the same time a more appropriate specification of the surface albedo, including seasonal and possibly moisture dependency, and of the roughness parameters is required and presently under study. This has to be accompanied by a careful validation of the surface scheme and of the surface radiation budget, which is being undertaken with station data and satellite observations.

In parallel with the special effort dedicated to the improvement of land surface parametrization, the treatment of the snow cover has also been the object of specific attention during the last few years. Revisions have been made in 1987 to the thermal properties of snow-covered land with a reduction of thermal diffusivity function of the grid-averaged snow cover. The possibility of positive atmospheric temperatures over a partially snow covered ground is also allowed by a distinct heat budget for the snow-free and the snow-covered fraction of a grid-box. As it was already done before, the fractional snow cover is assumed to be proportional to the grid-averaged snow depth until it reaches a value of 1 for a grid-averaged snow depth equivalent to 1.5 cm of water. Further modifications are under experimentation, mainly a revision of snow albedo including a dependency with surface temperature with deposition of ice in the interception reservoir and with the height of the vegetation (as represented by its roughness length).

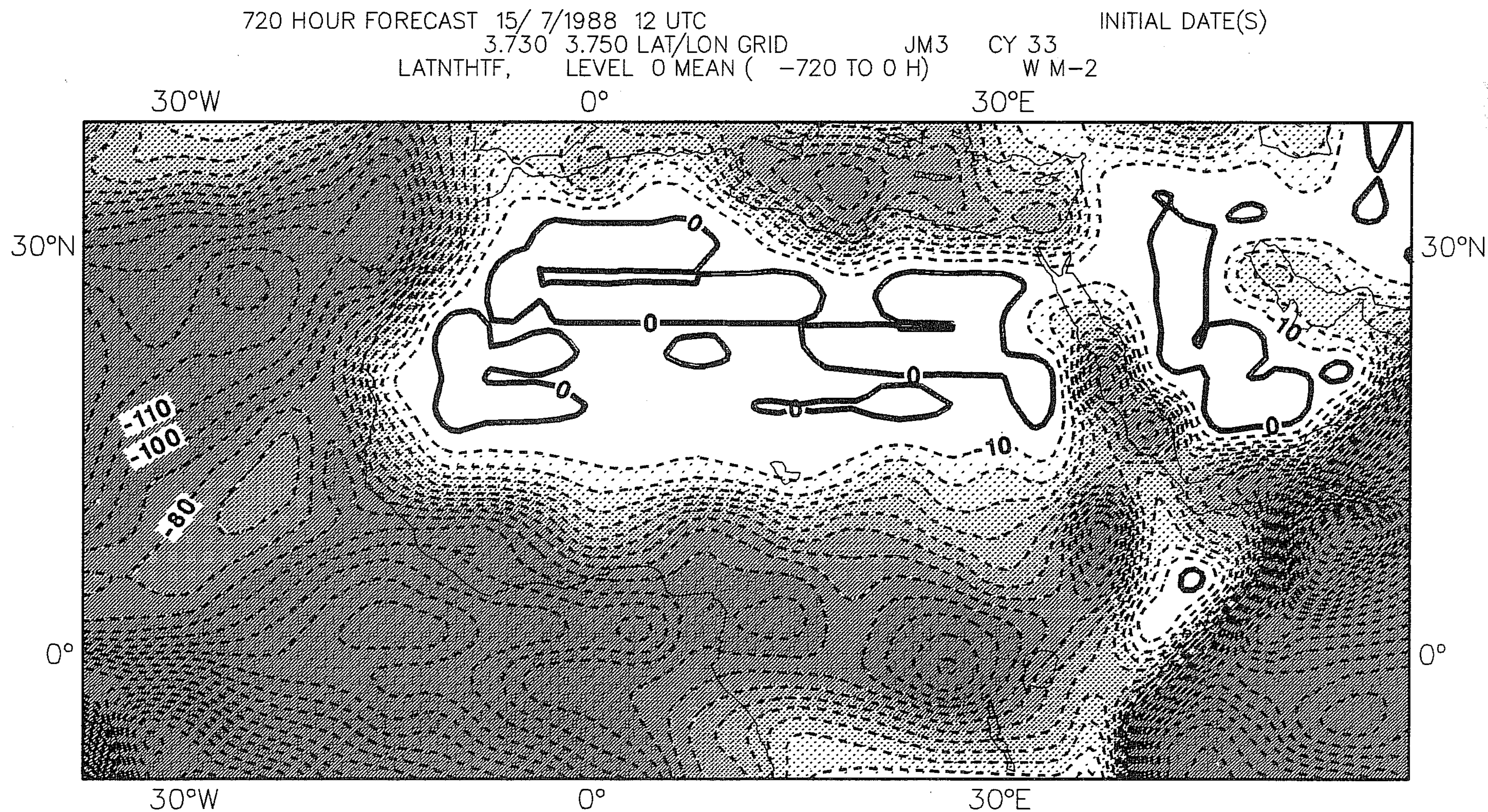

Fig. 7 Latent heat flux from a control T63 30 day experiment over an area centred on tropical Africa. (30 day average, forecast from 15 July 1988 12 00Z).

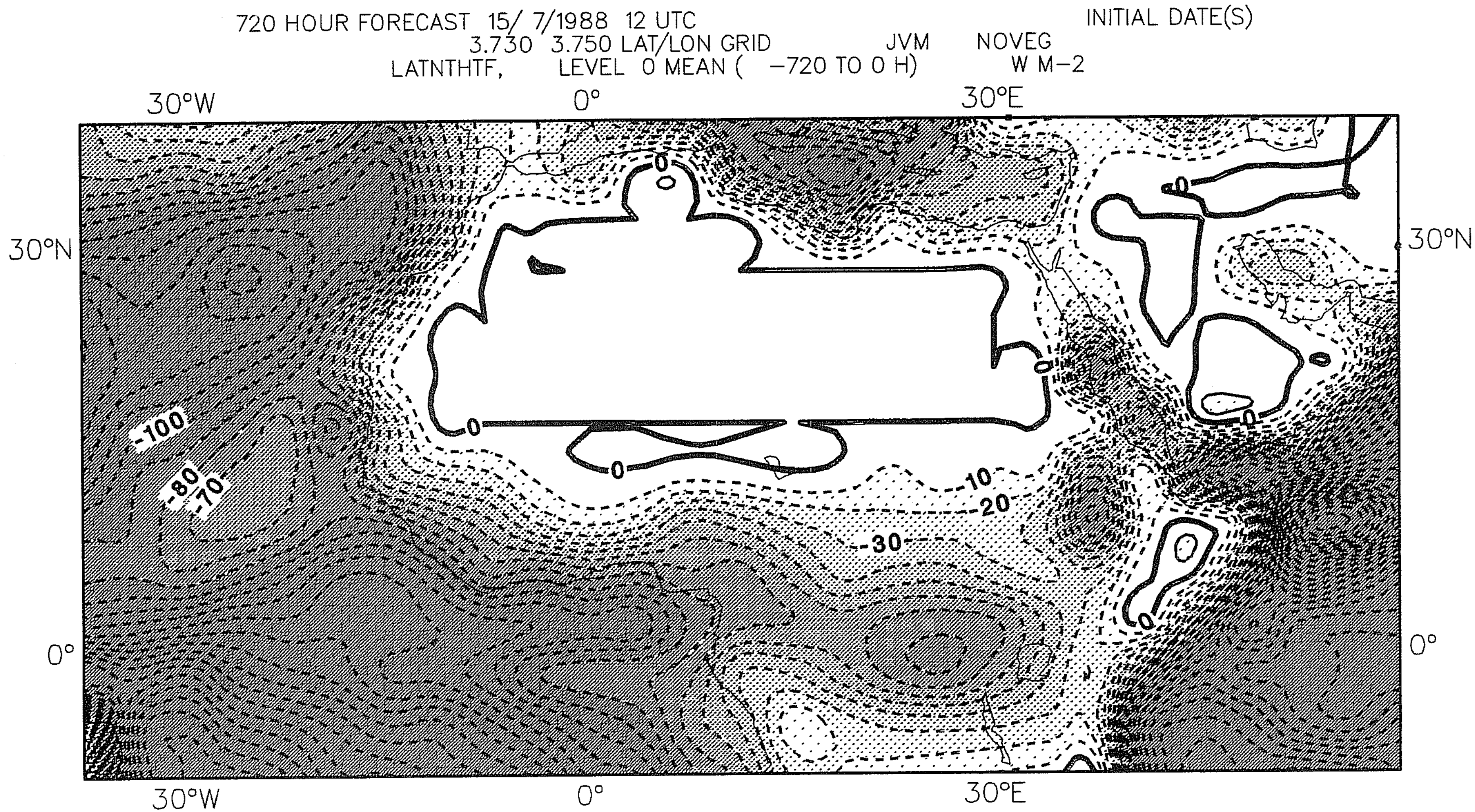

Fig. 8 Same as Fig. 7 for an experiment where vegetation cover has been set to zero.

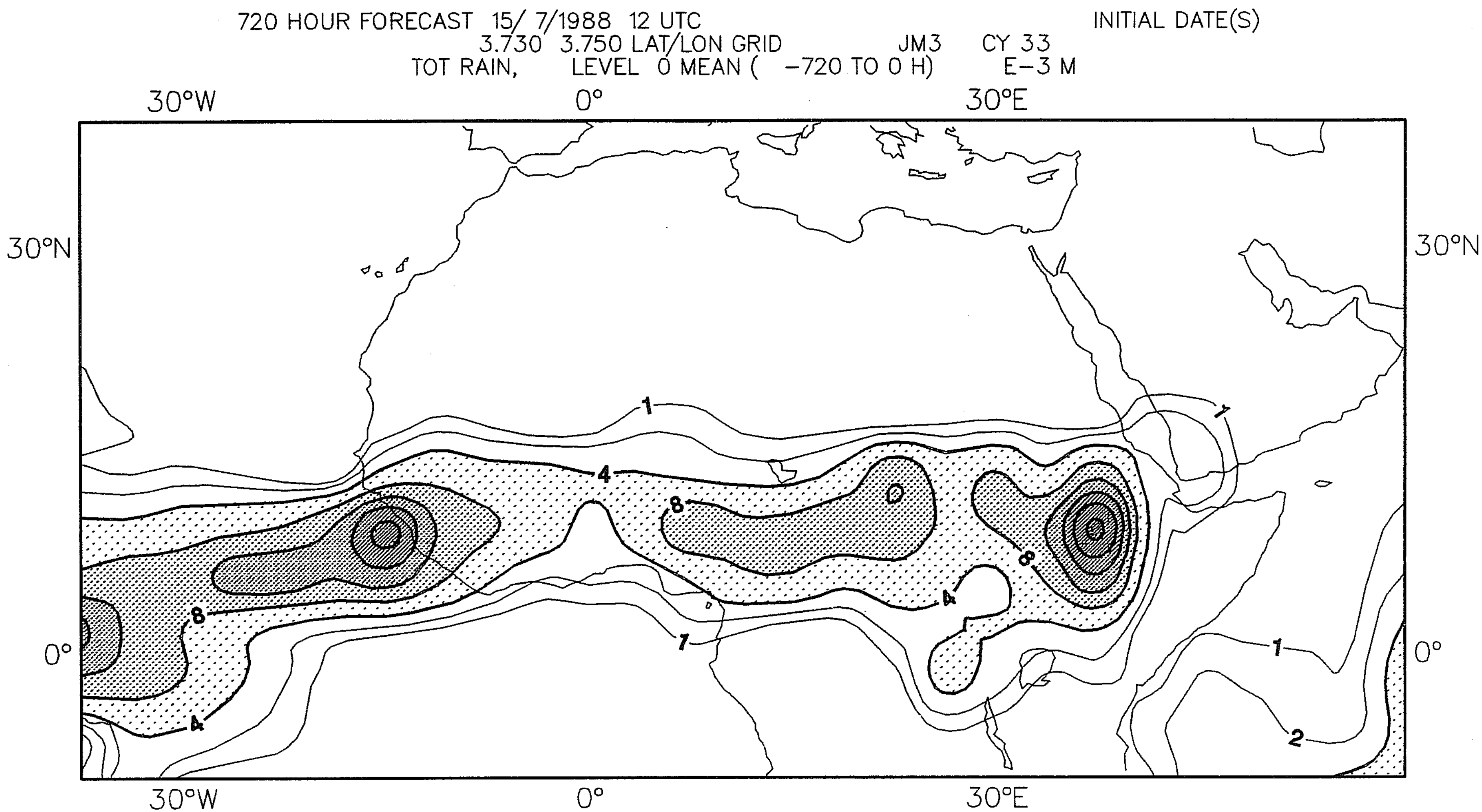

Fig. 9 Total precipitation forecast (large scale + convective) from the same experiment as Fig. 7 (30 day average).

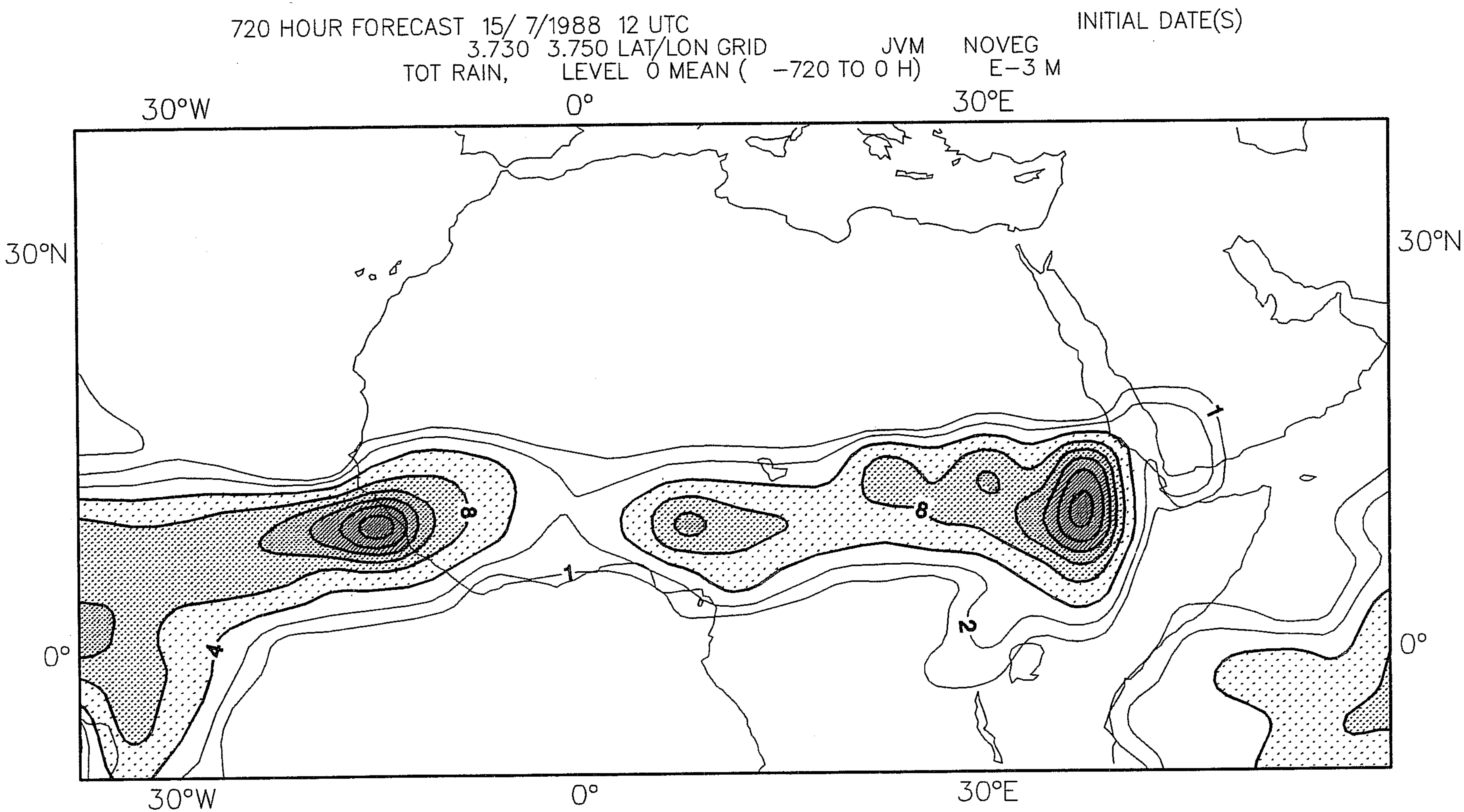

Fig. 10 Same as Fig. 9 for the experiment where vegetation cover has been set to zero.

4. RESEARCH ON THE PARAMETRIZATION OF BOUNDARY LAYER CLOUDS

Most PBL schemes have been developed for undisturbed cases in the absence of cloud cover. In disturbed conditions, the convection scheme would generally supersede the PBL scheme with respect to the vertical transports involved. However, it appears now clearly that specific schemes are required to represent cloud-topped PBL in the absence of deep convection. Planetary boundary layer cloud processes have basically two main effects of interest to the formulation of meteorological models. First, they tend to increase turbulent fluxes of heat, moisture and momentum, not only within the cloud layer, but throughout the PBL and at the Earth's surface. Second, they modify the radiative fluxes throughout the atmosphere and particularly at the surface.

Any PBL scheme can in principle be modified to include cloud processes but the difficulty comes in great part from the sparse information available in the GCM on PBL characteristics. In particular, the lack of vertical resolution associated with a coarse horizontal grid does not allow full use to be made of what is known from micrometeorological studies about PBL cloud processes.

At the gridscale of a GCM, it is usually assumed that clouds occur between a cloud base height above ground (h_B) and a cloud top height (h_T), and cover a fraction α of a horizontal grid. A comprehensive parametrization scheme should in principle provide a coherent formulation for radiative and turbulent processes associated with a given cloud cover. A first attempt is the scheme developed by Randall in the UCLA model, which is presently valid for stratocumulus decks (α=1) and for clear boundary layers [α=0) in the framework of a bulk PBL parametrization (Randall et al., 1985). The cloud layer is considered as part of the bulk PBL. However the extension to fractional cloud cover remains, to my knowledge, an open problem.

A natural framework to discuss the parametrization of shallow convection, defined as nonprecipitating moist convection, is to use thermodynamic variables which are conserved in the condensation process (formation of cloud droplets which remain small enough to be carried along with the air). This has the advantage of eliminating the need for a detailed formulation of the condensation-evaporation process in the turbulence equations. A set of quasi-conservative variables is

$$q_w = q + q_\ell \qquad (12)$$

$$\theta_\ell = \theta \frac{L\, q_\ell}{c_p T} \ , \qquad (13)$$

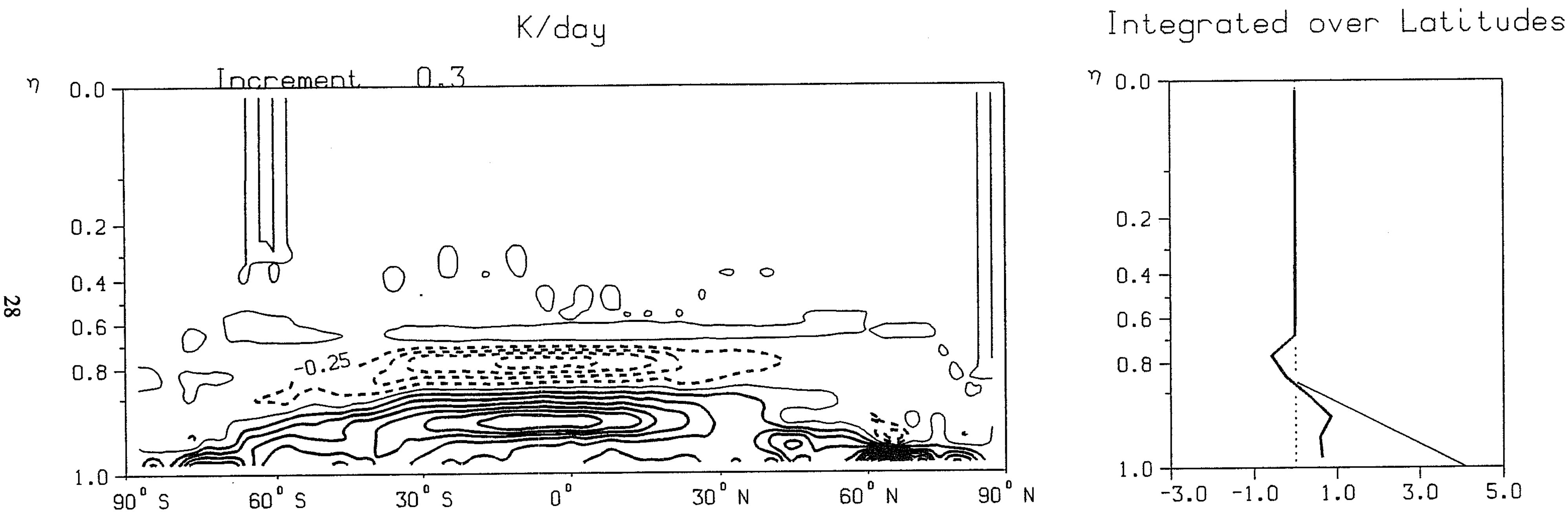

Fig. 11 Latitude-vertical cross-section of the temperature tendency for shallow convection in a T42 90 day control forecast from 1 December 1987 12 00Z (average from day 61 to 90). The global average is displayed on the right along with the total including shallow convection plus dry vertical diffusion.

where q_ℓ and q_w are the mixing ratios of cloud liquid water and total water, respectively, c_p is the specific heat of air at constant pressure, and θ_ℓ the liquid water potential temperature. The approximations entering the definition of θ_ℓ are discussed in Betts (1973), and the advantages of its use in a model are presented in Deardorff (1976).

However, these variables are not generally used in the main GCM equations because of the need to go back to more usual variables in the other computations: q and q_ℓ are required for any precipitation formulation or subgrid parametrization; T, q and q_ℓ are required for radiation computations; the virtual potential temperature $\theta_v \simeq \theta(1 + 0.61q - q_\ell)$ is required for buoyancy computations. A method to diagnose θ, q, q_ℓ and the buoyancy flux from θ_ℓ and q_w and some hypotheses on their statistical distribution is proposed in Sommeria and Deardorff (1977).

If one uses the q_w, θ_ℓ variables, the parametrization of moist PBL processes can be expressed in a way similar to the parametrization of dry processes with the variables q and θ. This can be done in the framework of second-order closure or with any simpler approach using eddy-coefficient relations. The main additional problem to the dry case is the occurrence of terms involving correlations with θ_v in the expressions for vertical fluxes and in the TKE equation. Even the simplest eddy-coefficient formulations make use of some form of the TKE equation and thus require an evaluation of the vertical buoyancy flux, $\overline{w'\theta_v'}$. In particular, this flux is implicitly used in the definition of the Richardson number entering most eddy-coefficient formulations.

As discussed in Deardorff (1976b) and Sommeria and Deardorff (1977), the expressions for $\overline{w'\theta_v'}$ in terms of the q_w and θ_ℓ fluxes in unsaturated and saturated situations are very different from each other. To be used over a grid volume which is partially covered with clouds, the unsaturated and saturated expressions have to be combined with the respective weights $1 - \alpha$ and α, assuming that the fractional cloud cover α is computed by other means. This approach also assumes, but this can be questioned as discussed by Randall (1987), that clear air and cloud processes can be separated in space within a grid volume, and that gridscale effects are simply a linear combination of both.

Various methods have been proposed for the computation of α from the knowledge of the grid-averaged variables as, for example, in Albrecht (1981) and Manton (1983). Manton defines α from the knowledge of the surface buoyancy flux, the height of the mixed layer and the condensation level for the near-surface air. This value is then used in the turbulent flux parametrization including a TKE equation. Tests carried out in the ECMWF model in a case of cold air outbreak over the China Sea were satisfactory, but problems remained in the use of the scheme in areas where results are sensitive to the fractional cloud cover computation. If the model uses a second-order closure or at least a TKE equation, hypotheses on the statistical distribution of θ_ℓ and q_w allow the derivation of expressions for α as a function of the departure of the mean grid properties from saturation and of the variances and covariances of the distribution of q_w and θ_ℓ. The same approach provides an estimate for $\overline{q}_\ell$, the grid-averaged $\overline{q}_\ell$, thus allowing to compute $(\overline{q}, \overline{\theta})$ from $(\overline{\theta}_w, \overline{q}_\ell)$ and vice versa.

A first application of a shallow convection scheme to a GCM was done in the ECMWF 1985 physics (Tiedtke et al., 1988). Vertical diffusion is enhanced by an empirical value throughout the depth of the assumed cloud layer. The effect of such a scheme can be appreciated on Fig. 11, which displays the zonally averaged temperature tendency due to the shallow convection scheme averaged between day 61 and 90 in a T42 N. Hemisphere winter integration. It produces mixing between the upper part of the boundary layer (800-900 mbars, warmer with respect to potential temperature, and generally thermally stable), and the lower part (often a well mixed layer). The net result is a cooling of the upper boundary layer (by .5 deg or more per day in the belt 30°N - 30°S) and a warming below (up to about 1 deg/day in the tropical belt). The warming comes in addition to the dry vertical diffusion effect (as indicated in Fig. 11) which is approximately the same with or without shallow convection.

The main consequences of having a shallow convection scheme are discussed in Tiedtke et al., (1988), but the effect is combined there with the effect of other model changes. In the example shown here, the shallow convection only is included in or removed from the 1988 model physics. Its main impact is, by allowing an additional mixing at the top of the boundary layer, to provide better conditions for the development of deep convection, resulting in a slightly more intense but mainly more organized tropical circulation. The global convection precipitations are increased in the present case by approximately 7% mostly through an intensification of the intertropical convergence zone over oceans. The latent heat release by large scale condensation is globally increased by about 10% and takes place, in the simulation with shallow convection, away from the tropics and above the boundary layer (Fig. 12). In the simulation without shallow convection (Fig. 13) large-scale condensation is less organised at the global scale with, in addition, spurious condensation occurring at the top of the boundary layer, specially over tropical oceans. The better organization of tropical precipitations, with well defined and more intense convergence going in the forecast with shallow convection, is illustrated on total precipitation maps (Fig. 14 and 15). Changes in the organization of temperate systems over the North Atlantic cyclone track are also noticeable.

The object of the above remarks is not to give a comprehensive account of the effects of a shallow convection scheme but to illustrate the importance of having a sound parametrization of this process. Following the implementation of a shallow convection scheme at ECMWF, another simple formulation has been proposed by J.F. Geleyn (1987) for the

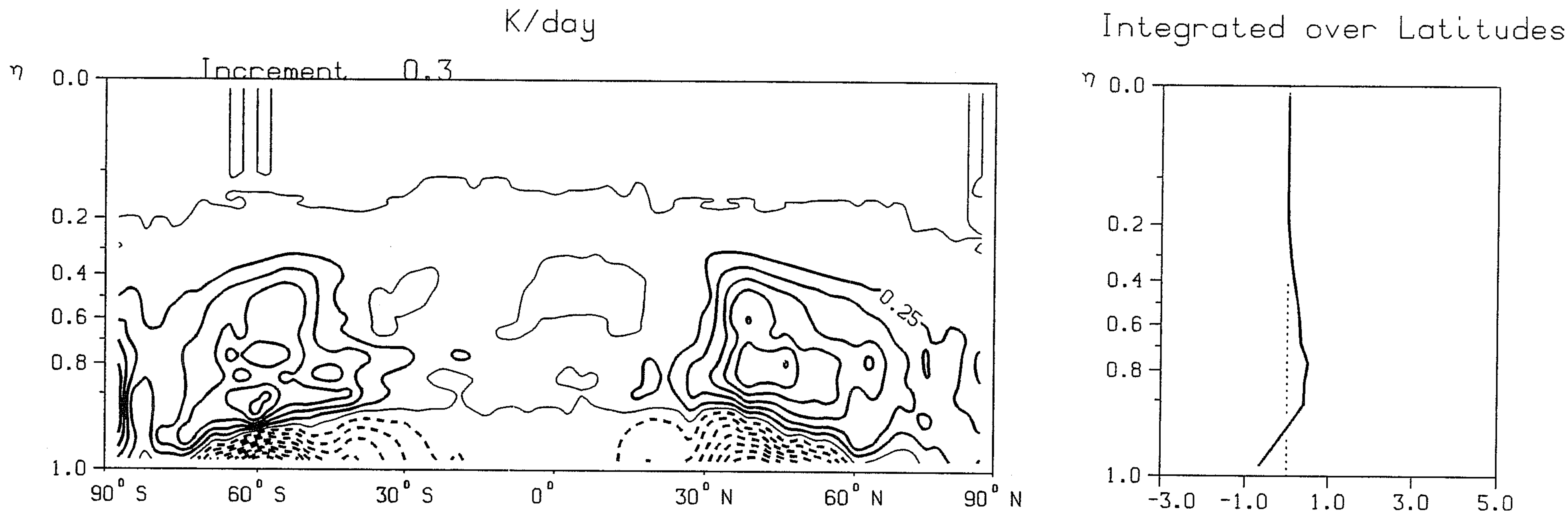

Fig. 12 Same as Fig. 11 for large-scale condensation.

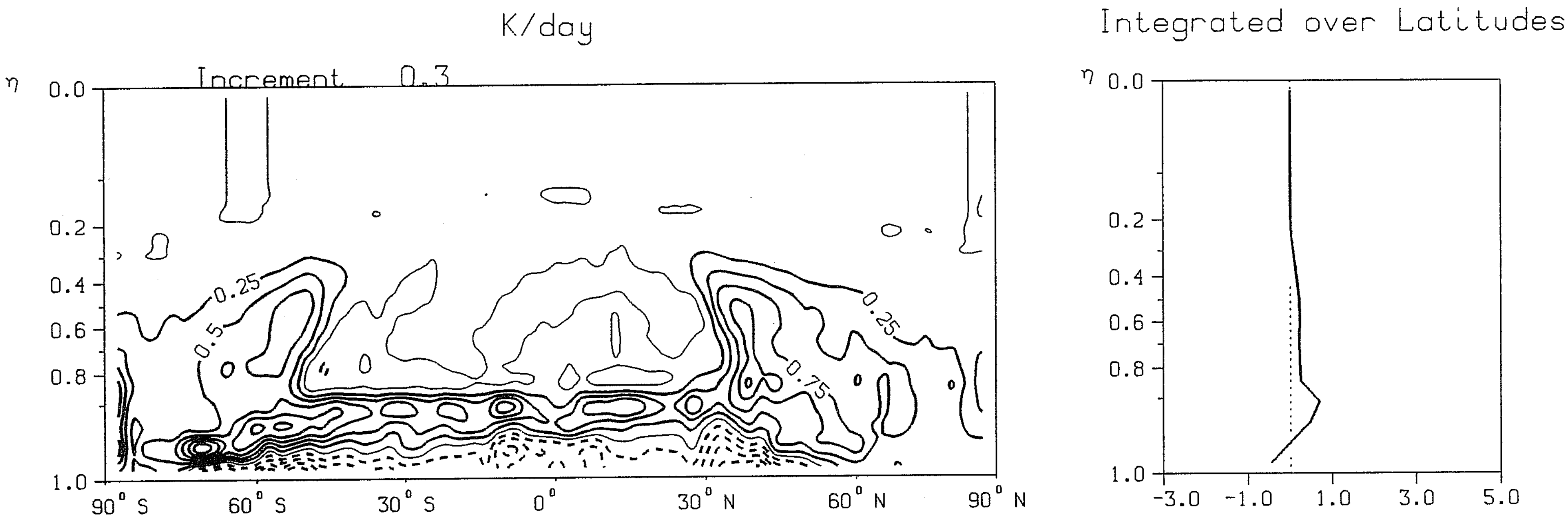

Fig. 13 Same as Fig. 12 for the same forecast but with a physics from which shallow convection has been removed.

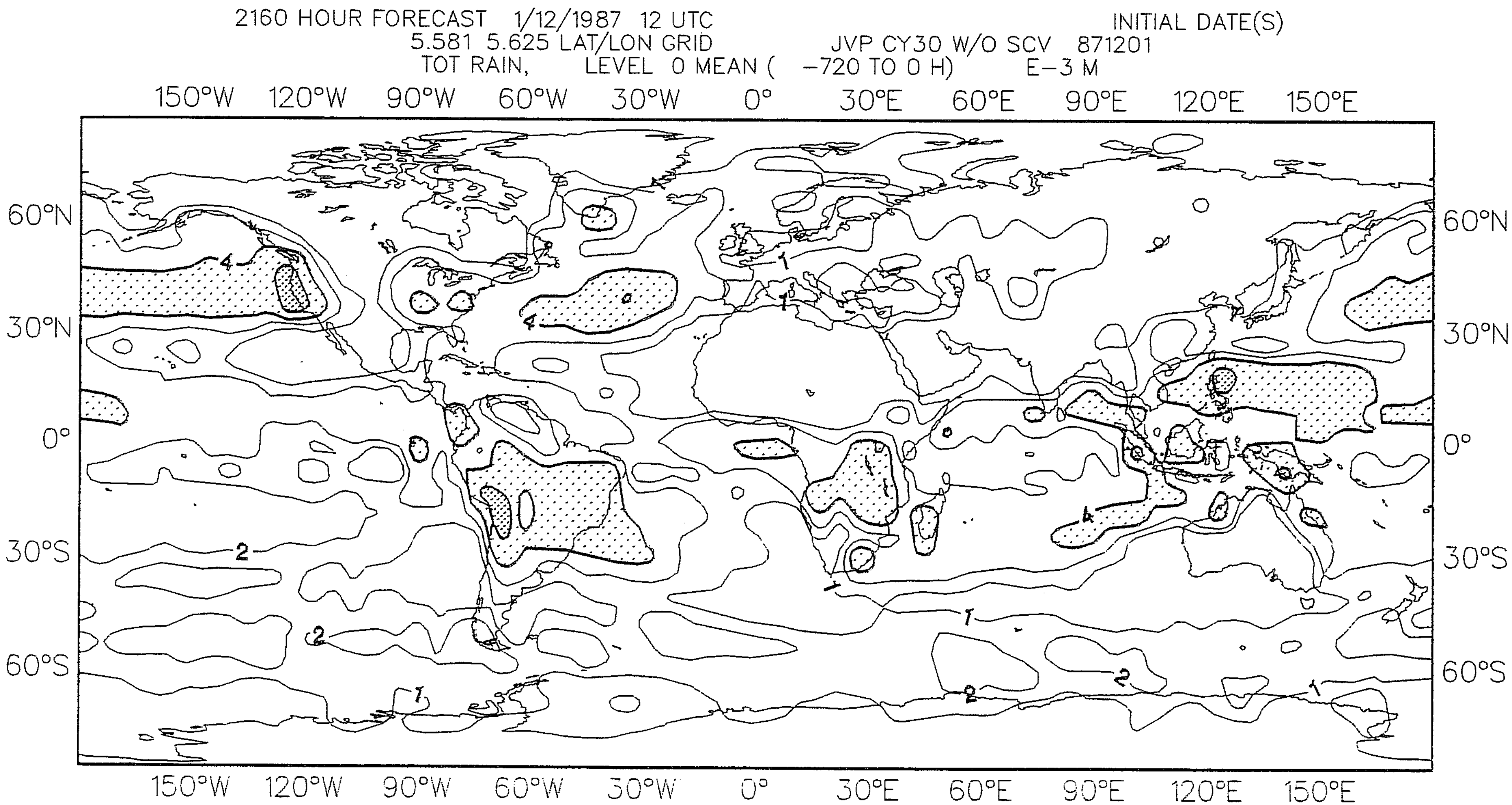

Fig. 14 Total precipitation map for the control forecast (averaged between day 61 and 90).

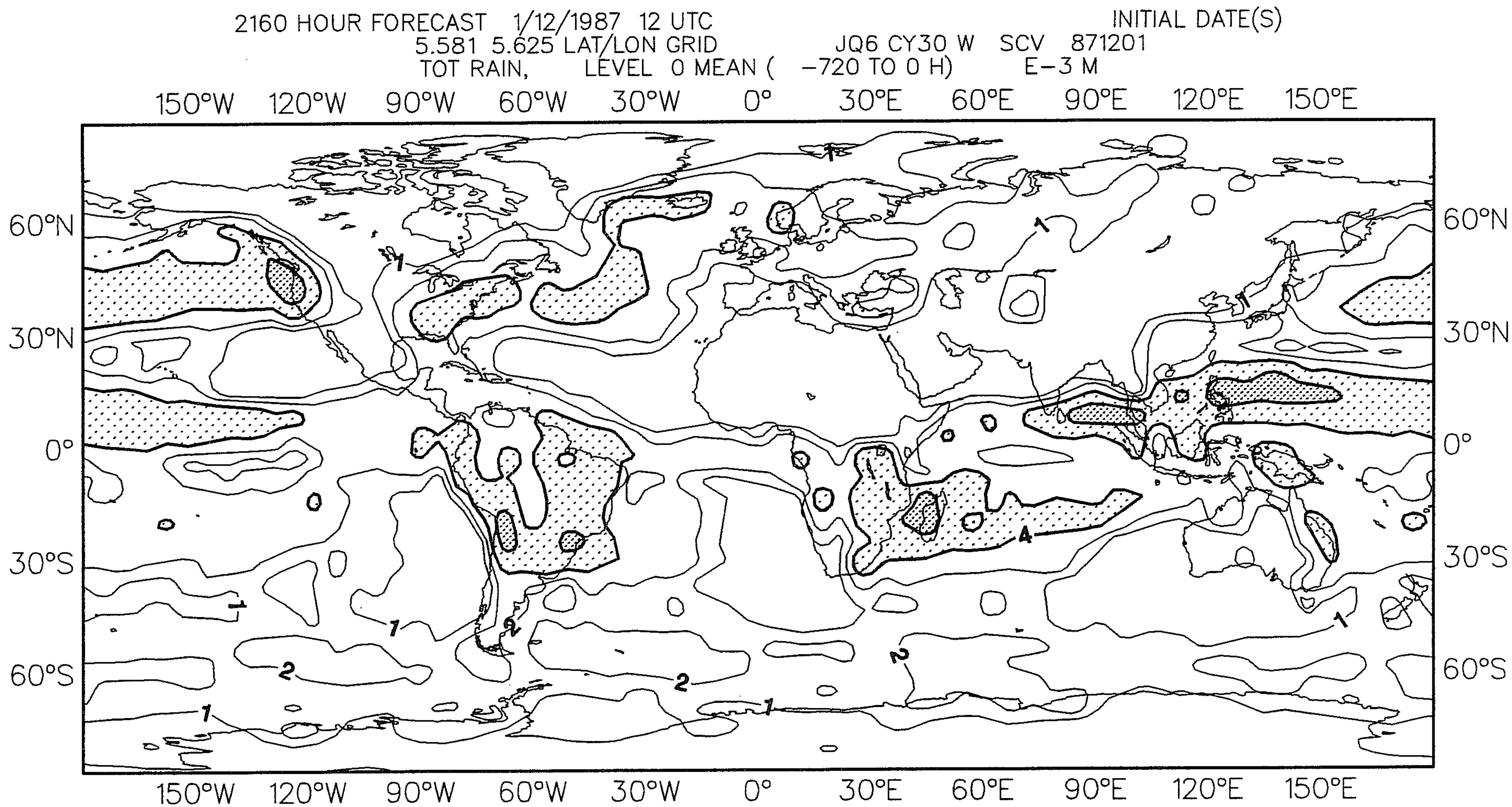

Fig. 15 Same as Fig. 14 but for the experiment without shallow convection.

French "Emeraude" model, where the buoyancy effect of shallow convection is formulated by a modification of the Richardson number in the vertical diffusion computation.

Smith (1988) has implemented in the UK Met. Office climate model a formulation similar to the one described above, with the non-gaseous cloud water, $q_{\ell} + q_f$, q_f is the solid cloud water, as an additional model variable and assumptions of triangular probability distributions for conservative turbulent variables. A similar type of scheme, but at a less advanced stage is being tested in the ECMWF model by H. Le Treut. It deals only with liquid water for the moment and assumes a top-hat-distribution for subgrid turbulent variables. Following R. Smith's initiative, a more complete scheme with prognostic turbulent kinetic energy has been implemented by Smith, Golding and Ballard in the UK meso-scale model. In addition to the features mentioned above, Burk (1985) has implemented in the U.S. Navy regional model a scheme including a prognostic equations for turbulent kinetic energy and the computation of fractional cloud cover. It is vertically nested in the main model, interacting with the large scale variables at a small number of levels but allowing at the same time a good vertical resolution of boundary layer processes. It is felt that this technique is probably the most appropriate to follow as well in a global forecast model.

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