ATMOSPHERIC PARAMETERIZATION SCHEMES IN METEO-FRANCE'S ARPEGE N.W.P. MODEL

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1. INTRODUCTION

ARPEGE is the name at Météo-France for the ARPEGE/IFS general NWP software jointly developed with ECMWF since 1988. For our purpose we shall here only concentrate on the forecasting and diagnostic features present in the operational forecasting version at Météo-France, especially concerning the atmospheric part of the physical parameterizations (i.e. no surface scheme's description, since the rules of development of that part are rather specific), and with less details about the radiation scheme (owing to the great complexity of the preparation of the data supposed to represent the optical properties of the various atmospheric constituants). This paper will be strictly limited to the presentation of the methods and algorithms, the mass of results that could be shown in support of the many choices done during more than ten years of development being left to other occasions.

The main characteristics of the Météo-France ARPEGE physical package are :

* full consistency between the different parameterization schemes, the set of universal constants and of basic thermodynamic functions and the full model equations ;

* no fictitious source/sink of energy and/or water ;

* no implicit value for any existing or even potential tuning parameter ; all such parameters expressed in SI units ;

* no a priori dependency of the physical calculations upon the number and/or the positioning of the model's vertical levels.

Priority over any other consideration was given to the absolute respect of these rules, in order to have the maximum flexibility and so-called "plug-compatibility" for our physical computation routines. Several technical arrangements, which will not be reported here, have been worked out to simplify the enforcement of the above-mentioned rules. In the following, only the "generality" aspect of this strategy will be noticeable through the explained choices of equations and/or algorithms, but the reader should keep in mind that this is only the "visible part of the iceberg" ! In the same Seminar Proceedings, one

paper by J.-F. Geleyn tries to go further in that direction, with some results mainly to qualify the importance of the thermodynamic choices outlined below.

The presentation of the exact algorithms will only be made when no literature is available on the particular subject. The state of the description is that of first of September 1994.

2. THERMODYNAMIC FRAMEWORK

We start by defining the following principles :

- The atmospheric system consists of a mixture of two perfect gases, dry air and water vapour ;
- The atmospheric system is in thermodynamic equilibrium ;
- The condensed phases of water which appear in the atmosphere are immediately eliminated from the system (in one time step) at least up to now ;
- The specific volume of the condensed phases of water is equal to zero ;
- The condensed phases of water which leave the atmospheric system remove mass, momentum and energy (kinetic, potential and internal) from the atmosphere ;
- During their fall they can exchange mass, momentum and energy in the underlying layers ;
- The exchange of momentum is obtained by assuming that the precipitations are in equilibrium of horizontal velocity in the layers they cross ;
- The temperature of the precipitations is assumed to be that of the air they cross (even if the wet-bulb temperature would be more logical) ;
- The potential energy of the precipitations is converted into kinetic energy which is dissipated and converted into internal energy.

Like in most current NWP models we consider the coupled variations of C_p and R with the specific humidity of water vapour q_v in case of moist air (and especially in the so-called conversion term of the thermodynamic prognostic equation $RT(dp/dt)/(C_p.p)$). Assuming, for the time being, that there are no prognostic variables in our system for the liquid and ice water contents of the atmosphere, this reads :

$$R = R_d+(R_v-R_d).q_v \tag{1}$$

$$C_p = C_{pd}+(C_{pv}-C_{pd}).q_v \tag{2}$$

But, unlike most "state of the art" NWP models, we also take into account the variation of the latent heats of vaporization and sublimation with temperature, e.g. :

$$L_v(T) = L_v(T*)+(C_{pv}-C_w).(T-T*) \tag{3a}$$

$$L_s(T) = L_s(T*)+(C_{pv}-C_i).(T-T*) \tag{3b}$$

$T*$ being the triple point temperature.

Given the fact that this item has little influence on the results of short/medium range NWP integrations (we verified it), this may appear as an unnecessary refinement. However, provided one accordingly rewrites the algorithms for condensation and moist adiabatic calculations (see below), the induced changes to the code are very limited and they provide the framework to have an exact description of all energy cycles in the atmosphere (including the Clausius-Clapeyron cycle), a fact that may become relevant when this type of code is used inside the data assimilation cycle of a NWP operational suit.

The inherent non-linearity of the saturation specific humidity functions $q_s(T,p)$ makes it necessary, even when neglecting the time variations of pressure, to use iterative Newton-type algorithms in order to find (T,q_v) solutions to the coupled set of equations expressing exact saturation of the final state as well as conservation of moist enthalpy, either at constant geopotential ("stratiform" condensation-evaporation process) or along the vertical (moist adiabatic ascent).

It is interesting to note here that taking into account both the variation of C_p with q_v and that of L with T in the course of such processes allows to construct simpler algorithms for the Newton-type loops than when only one of the two effects is considered.

Indeed, starting from initial values T_0,q_{v0},C_{p0} and L_0 one readily gets (for the "stratiform" case) :

$$C_p = C_{p0}+(C_{pv}-C_{w/i}).(q_v-q_{v0}) \tag{4}$$

and

$$L = L_0+(C_{pv}-C_{w/i}).(T-T_0) \tag{5}$$

which coupled with

$$C_p.dT+L.dq_v = 0 \tag{6}$$

leads to

$$L.C_p = L_0.C_{p0} \tag{7}$$

Linearizing the saturation function for each iteration and eliminating the terms of the form $(T_{i+1}-T_i)^2$ leads then to the following "symmetric" algorithm :

$$q_{vi+1}-q_{vi} = (q_s(T_i,p)-q_{vi})/(1+(L_i/C_{pi}).dq_s(T_i,p)/dT_i) \quad (8a)$$
$$C_{pi+1}-C_{pi} = (C_{pv}-C_{w/i}).(q_{vi+1}-q_{vi}) \quad (8b)$$
$$T_{i+1}-T_i = -(L_i/C_{pi+1}).(q_{vi+1}-q_{vi}) \quad (8c)$$
$$L_{i+1}-L_i = (C_{pv}-C_{w/i}).(T_{i+1}-T_i) \quad (8d)$$

For the moist adiabatic case the situation is a little bit more complicated since one has to consider both a transport from the base of an atmospheric model slab (subscript "b") to its top (subscript "t") and the associated change of geopotential ϕ (at fixed pressures). Nevertheless, choosing to do the whole Newton-type loop at "t", and considering the property

$$L.C_p+(C_{pv}-C_{w/i}).(\phi_t-\phi_b) = L_b.C_{pb} \quad (9)$$

one can obtain a very similar algorithm e.g.

$$q_{vi+1}-q_{vi} = (q_s(T_i,p_t)-q_{vi})/(1+(\underline{L}_i/\underline{C}_{pi}).dq_s(T_i,p_t)/dT_i) \quad (10a)$$
$$\underline{C}_{pi+1}-\underline{C}_{pi} = (C_{pv}-C_{w/i}+\underline{R}_v^t).(q_{vi+1}-q_{vi}) \quad (10b)$$
$$T_{i+1}-T_i = -(\underline{L}_i/\underline{C}_{pi+1}).(q_{vi+1}-q_{vi}) \quad (10c)$$
$$\underline{L}_{i+1}-\underline{L}_i = (C_{pv}-C_{w/i}+\underline{R}_v^t).(T_{i+1}-T_i) \quad (10d)$$

with

$$\underline{C}_p = C_p+\underline{R}_b^t+\underline{R}_v^t.(q_v-q_{vb}) \quad (11a)$$
$$\underline{L} = L+\underline{R}_v^t.T \quad (11b)$$

where the change of geopotential $(\phi_t-\phi_b)$ depends on T_b,q_{vb},T_t and q_{vt} only through three iteration-independent parameters of the type R.dln(p), ($\underline{R}_b^b$,$\underline{R}_b^t$ and $\underline{R}_v^t$), chosen from the model vertical coordinate discretization so that:

$$\phi_t-\phi_b = \underline{R}_b^b.T_b+(\underline{R}_b^t+\underline{R}_v^t.(q_{vt}-q_{vb})).T_t \quad (12)$$

Finally, using the $L_{v/s}$ dependency upon temperature, the Clausius-Clapeyron relationship (exactly integrated for vanishing volumes of condensed water forms) readily leads to :

$$ln(e_s(T)) = \alpha-\beta/T-\gamma.ln(T) \quad (13)$$

with

$$L(T) = R_V.(\beta-\gamma.T) \text{ (and } \alpha \text{ from } e_S(T*)) \quad (14)$$

leading to (*Courtier*, 1988)

$$d(ln(e_S(T))) = ((\beta-\gamma.T)/T^2).dT \quad (15)$$

α,β and γ being of course different between the liquid and ice cases.

Concerning the $q_S(T,p)$ and $dq_S(T,p)/dT$ functions, we had to cater for an additional difficulty, namely to avoid that q_S may become absurd if $e_S(T) \geq p$, a fact that would happen if the formula for $e_S(T) < p$ would be applied in a straightforward fashion. Choosing arbitrarily q_S to be equal to $e_S(T)/p$ in the case $e_S(T) \geq p$ allows to obtain an economical solution to this problem, e.g. :

$$q_S(T,p) = \Gamma/(1+(R_V/R_d-1).\max(0,1-\Gamma)) \quad (16a)$$
$$dq_S(T,p) = q_S.(1-q_S).d(ln(e_S(T)))/(1-\Gamma) \quad (16b)$$

with

$$\Gamma = e_S(T)/p \quad (17)$$

To further concretize all that rationalization effort, the prognostic equations for momentum, water vapour and enthalpy (or rather dry static energy at the left hand side) have been worked out as pure flux divergence equations. This requires to associate enthalpy fluxes to the falling precipitation fluxes. Taking into account all effects (no sink - no source) and applying the above-mentioned rules one obtains the following expression for those new fluxes :

$$Fhp_{l/i} = -Fwp_{l/i}.(L_{v/s}(T)-(C_{pv}-C_{pd})T) \quad (18)$$

An extension of this type of formula to the case when liquid and ice phases would not any more been instantaneously removed has been found in collaboration with H. Sundqvist (personal communication), in preparation for introduction of this new feature in the ARPEGE code.

All the above-mentioned has been coded with the possibility to use an option in which the total mass of the atmosphere varies with surface evaporation and precipitation fluxes, following an idea of M. Rochas. Up to now the only significant impact was detected in climate-type simulations for the summer monsoon that has a stronger intensity if this option is activated.

3. NON RADIATIVE CLOUD PROCESSES

3.1 **Stratiform precipitation**

This part of the parameterization package assumes either full coverage (generation of precipitation) or no cloud (evaporation in sub-saturated air). Melting/freezing of falling precipitations is also handled at the same time. The following rules and equations fully define the algorithm :

- as already mentioned, all supersaturation is removed via either liquid water or ice water fall;
- the distinction at that stage between liquid and ice water phases is twofold : (i) for thermodynamic effects of phase changes a brutal transition at the triple point is imposed in order to remain consistent with the saturation functions ; (ii) for "mechanic" properties, i.e. fall-speed depending rate of evaporation and/or melting-freezing, the distinction is made at the time of the generation and depends only on temperature, following

$$f_i(T) = 1.- \exp(-0.5((T-T*)/T_i)^2) \text{ if } T<T*, \tag{19a}$$

$$f_i(T) = 0. \text{ otherwise,} \tag{19b}$$

T_i corresponding to the difference between $T*$ and the temperature of maximum difference between the two saturation curves (T_i = 11.82 K), while track is kept of the two forms of precipitation, the pure ice one having rates of change E_{vap} and F_{ont} (see below) that are R_{apv} times those of liquid water (currently R_{apv} is equal to 80. , a value derived from the computations presented in *Clough and Franks* (1991)) ;

- the rate of evaporation in unsaturated air is computed from a Kessler-type formula (*Kessler*, 1969) that was obtained following the original rules but with a revisit of the basic data and some simplifications for a more compact presentation :

$$d(\sqrt{R})/d(1./p) = E_{vap}.(q_w-q_v) \tag{20}$$

with R rainfall rate, q_w wet bulb temperature (also used as target for removal calculations) and E_{vap} currently equal to $4.8\ 10^6$;

- a similar formula is derived for melting/freezing assuming the same proportionality between molecular fluxes and gradients for temperature than for humidity :

$$d(r_i)/d(1./p) = F_{ont}.(T-T*)/\sqrt{R} \tag{21}$$

with r_i the proportion of falling ice phase (in either of its two roles, see above, hence two such equations solved) and F_{ont} currently equal to 2.4 10^4 . The ratio of the constants E_{vap} and F_{ont} of 200 corresponds to half the ratio $(L_s-L_v)/C_p$ multiplied by the ratio of the molecular diffusivities for water and temperature respectively ; as such it should be kept constant during tuning processes.

3.2 **Convective precipitation**

The convective parameterization is the one described in *Bougeault* (1985) with a certain number of small modifications :

- addition of several securities to avoid numerical problems, including a vertical averaging of the sensible heat flux compensating term, the use for mass-flux type computations of the semi-implicit algorithm described in *Geleyn et al.* (1982), and a protection against non linear instability in the same computation as described in *Courtier et al.* (1991) ;

- introduction of a variation with height of the entrainment rate, starting from a maximum value E_{ntrx} at the bottom and relaxing exponentially towards a standard value E_{ntr} as one goes up the cloud ; the vertical scale of this relaxation has been chosen as $(E_{ntrx}^3.E_{ntr})^{1/4}$. This assumption of varying entrainment is similar to the one made by *Gregory and Rowntree* (1990) in the UKMO model.

- use of the thermodynamic framework for precipitation fluxes that has been introduced in the preceding part. One additional hypothesis is necessary here to go from integrated constraints to local fluxes. We thus elected to say that the diffusive flux of water vapour is the product of the mass flux by the difference between cloud and environmental (i.e. large scale) q_v. Furthermore, using the same hypothesis of a constant speed of adjustment towards saturation in and below the cloud than in *Geleyn* (1985) some account is being taken of sub-cloud evaporation ; this is done without any impact on the mass-flux and hence cannot be seen as an incorporation of downdrafts ;

- introduction of a distinction between ice and liquid phases for falling precipitations ; here, contrary to the large scale case where we used the "source distinction" approach, we choose the "target distinction" approach : the proportion of the two flux depends only on the temperature of the interface between levels (or of the surface) where it is computed. The ice proportion is one below $T*$ and decreases afterwards to zero following :

$$f'_i = 1.-\min(1.,M_{ltinv}.\max^2(0.,T-T*)/p) \qquad (22)$$

with M_{ltinv} currently equal to 1.25 10^4 . This value is in fact linked to those of the F_{ont} constant and of a constant relating convective cloudiness to convective precipitation (to be defined thereafter).

The only three other tuning constants of the scheme are currently set at 20000 m for the standard entrainment characteristic height (E_{ntr}^{-1}), at 2500 m for its maximum entrainment counterpart (E_{ntrx}^{-1}) and at 1000 m for the so called "depth of the non precipitating cloud" (in fact another length scale used in

an exponential type formulation, following the proposal of *Arakawa and Schubert* (1974) in the Appendix to their famous paper). The existence of the latter constant corresponds to the only "microphysics" we are putting in the cloud ascent model : the condensed/frozen water during the uplifting is converted into precipitation at a rate proportional to the steady-state remaining liquid/solid water content of the cloud. This may be the degree zero of convective cloud microphysics but the resulting tuning parameter is a very efficient regulating one.

The convection parameterization is the only one, in our choice of implementation, that has to know about some results of the dynamics, namely the total humidity advective tendency, that, combined with minus the divergence of the vertical turbulent flux of moisture will give the so-called "humidity convergence" used in the Kuo-type closure assumption of the Bougeault scheme. When running pre-operational tests of the ARPEGE model in stretched mode, it became obvious that the change of local resolution induced strong changes of the interaction between dynamics and convective forcing, with exaggerated activity in the area of maximum resolution. Following the results of *Piriou* (1991) the problem was partly cured when introducing a modulation of the sole dynamical part of the humidity convergence by a factor $1/(1+T_{loc}/T_{ref})$ where T_{loc} is the equivalent local triangular truncation at the considered grid point and T_{ref} a reference scaling triangular truncation value, currently tuned at 240.

3.3 **Turbulent effects**

The shallow convection parameterization is the one described in *Geleyn* (1986) with the additional security that it can only be activated if there exists some moist potential instability (to avoid "dry" shallow convection in case of very high stability, e.g. stratosphere under extreme conditions).

The modified Richardson number used in that parameterization is also used in the stability computations of the gravity wave drag parameterization (like in the FSU GCM), in order to have a consistent representation of clouds in all momentum dissipation processes.

4. COMPUTATION OF "RADIATIVE" CLOUD AMOUNTS

4.1 **Convective cloud amount**

This quantity is computed at the previous time step, after the completion of the convective calculations : one stores three informations, namely the top and bottom layers' indexes for this type of cloudiness and the total amount.

The latter is computed following *Tiedtke* (1988) :

$$N_{ctot} = \min(0.5, n_p.R_{csurf}) \tag{23}$$

with R_{csurf} the rate of convective precipitation at the surface and a slope n_p currently equal to 3000. .

At the time step when it is used, N_{ctot} is converted to a local N_c value at all levels between the two "marked" indexes, I_{ct} and I_{cb}. This conversion takes into account the forthcoming use of cloud cover information by the radiative code :

- if maximum overlap of adjacent cloud layers is to be used,

$$N_c = N_{ctot} \; ; \tag{24a}$$

- if random overlap of all cloud layers is to be used,

$$N_c = 1.-(1.-N_{ctot})^{(1./(I_{cb}-I_{ct}+1))} \; . \tag{24b}$$

Finally, after the "stratiform" cloud cover N_s has been locally defined (see next paragraph) the local total cloud amount is obtained following the usual combination rule :

$$N_t = 1.-(1.-N_s).(1.-N_c) \tag{25}$$

4.2 **Stratiform cloud amount**

The computation of N_s follows exactly the algorithm described by *Geleyn* (1980) :

$$N_s = \max^2(0., (q_v-Hu_c.q_s)/(q_s-Hu_c.q_s)) \tag{26}$$

with

$$Hu_c = 1. - Hu_{coe}.\sigma.(1.-\sigma).(1.+\sqrt{Hu_{til}}.(\sigma-0.5)) \tag{27}$$

and $\sigma = p/p_{surf}$, Hu_{coe} and Hu_{til} having currently (with a random overlap assumption) the values 1.5 and 4.5 .

Furthermore $N_s = 0.$ for all levels contiguous to the bottom and where the dry static energy is lower than that of the surface (cloud-free well mixed layer).

There is currently no cloudiness associated with the shallow convection computation and this is probably the main weakness of this "diagnostic-type" cloud amount scheme, otherwise apparently rather well tuned.

5. COMPUTATION OF "RADIATIVE" CLOUD LIQUID (AND ICE) WATER CONTENTS

The partition function between ice and liquid water phases is here exactly the same as the one used in the precipitation generation distinction for "mechanical" properties (see part III), a slight modification of the proposal of *Röckel et al.* (1991) from the data of *Matveev* (1984):

$$f_i(T) = 1.- \exp(-0.5((T-T*)/T_i)^2) \text{ if } T<T*, \tag{28a}$$

$$f_i(T) = 0. \text{ otherwise.} \tag{28b}$$

It will be used in the radiation code that knows different optical properties for the two phases, those being computed following the methods described in *Ritter and Geleyn* (1992) using the basic spectral data of *Stephens* (1984) for water and of *Röckel et al.* (1991) for ice clouds.

The total cloud water content is obtained in two steps. At first a "proportional" computation is done, following a modified version of the proposal of *Betts and Harshvardan* (1987) and *Somerville and Remer* (1984) :

$$q_{cw} = - (1./p*) \partial q_s/\partial (1/p) \Big|_{\Theta_e=\text{constant}} \tag{29}$$

i.e. the variation of the saturation value over a constant 1/p path for the moist adiabatic ascent ($p*$ is currently tuned at the value $8.\ 10^6$).

After that, following the work of *Tselioudis et al.* (1992), we wanted to impose an asymptotic maximum value to q_{cw}. Let us notice here that we did not follow these authors by going so far as to have a decrease of water content for the highest temperatures, for fear of introducing an uncertain and uncontrolable feed-back into the model. Furthermore, our system was yet unable to distinguish, for the same pressures and temperatures, between convective and large scale clouds in terms of water content. By assuming that the limiting asymptotic value would be inversely proportional to the previously computed total local cloud amount N_t we also somehow obtained this desirable feature for our parameterization :

$$q_{cw} = q_{cw}/(1.+N_t \cdot q_{cw} \cdot q_{cwcinv}) \qquad (30)$$

with q_{cwcinv} (the inverse of the limiting value for full coverage) having currently the value $1.5\ 10^4$.

From this final q_{cw} value and its partition function f_i, the radiation code will compute for each spectral interval and each layer the optical depths, single scattering albedos and back/up scattering fractions of the clouds that will occupy a fraction N_t of the mesh.

6. RADIATIVE CALCULATIONS

The radiation scheme considered here was primarily designed for short range fine-mesh NWP applications for which an immediate feed back between land-surface temperature, cloudiness and radiative forcing was judged essential. Hence the code was mainly designed for speed rather than absolute precision. Nevertheless it has proved rather robust and hardly biased in climate-type applications.

The main references are *Geleyn and Hollingsworth* (1979) for the algorithms and *Ritter and Geleyn* (1992) for the way to compute all "optical" constants as input to the code as well as for the use of a delta-two-stream approximation. In fact, contrary to what happens in the rest of the parameterization package of ARPEGE, all the optical constants are computed by a quasi-automatic process and are thereafter considered as non-tunable ones, owing to the enormous complexity of the programs necessary to go from laboratory/in-situ measurements to the constants appropriate for the simplified formulations used in the code. In fact, it was found necessary to keep one degree of freedom available for tuning : in first approximation its influence is limited to the control of the single scattering albedo of clouds in the solar part of the spectrum, and it was found that the latter parameter is a very efficient tuning mean for the overall tropospheric temperature. Nevertheless its adjustment was also done with respect to the literature, e.g. *Fouquart and Bonnel* (1980).

The additional information that is necessary to understand the code and that is present in neither of the two above-mentioned papers will be briefly mentioned thereafter :

- there are only two spectral intervals, one in the solar domain and one in the thermal one ;
- for gaseous absorption one assumes absolute random positioning between the lines of the three types of gases (H_2O, O_3 and all well mixed gases put together in a CO_2+ composite) ; for each of those three, the ratio of the optical depth to the equivalent width is parameterized by Padé functions of the latter which is itself obtained via a Malkmus "averaged" model with continuum (p-type and, in the case of thermal H_2O effects, e-type as well) ;
- all other absorptions are treated as "grey" with an off-line fit of all absorption and scattering coefficients for Rayleigh effect, clouds and aerosols ;

- full multiple scattering calculations are carried through, using delta-two-stream and adding methods ; for that we need local gaseous absorption optical depths that reflect the so-called "saturation effect" of previous absorptions ; the simplified way to deal with that problem that allows very fast calculations will be detailed in the next two paragraphs.

For solar absorption the gaseous absorption optical depth is taken exactly along the parallel downward beam and estimated for all diffuse beams along the directly reflected one at the surface (no atmospheric scattering is taken into account in that evaluation of the saturation effect).

For thermal radiation the situation is similar : "cooling to space" calculations are handled exactly while "exchange with surface" and "exchange between layers" terms are treated with an approximation. The latter consists in choosing as local optical depth the minimum of all such quantities under which the considered layer can be seen through a straight "diffuse beam" from any part of the atmosphere. In fact it is obviously sufficient to make that computation only from the top of the atmosphere (already done in fact for the cooling to space term) and from the surface and to choose the minimum between those two values. This method is not without drawbacks. Indeed, if there were exchanges between the surface and cold optically very thick layers, the results would be quite wrong (*Räisänen*, 1994). Fortunately the cold high clouds are sufficiently transparent for this deficiency to remain a minor one in the NWP context.

7. "DRY" VERTICAL TRANSPORT PROCESSES

7.1 **Gravity wave drag**

The gravity wave drag (currently only of orographic origin, an extension to convective sources being however considered (Cariolle, personal communication)) routine used in the ARPEGE physics comes from the superposition of three level of increasing sophistication which will be now detailed one by one :

*The linear basic model (closely following *Boer et al.* (1984)) computes a surface drag :

$$\tau_S = -\rho_S . K_{gwd} . N_S . \mid U_S \mid . \sigma_h \qquad (31)$$

with ρ_S, N_S and $\mid U_S \mid$ respectively being the density, Brunt-Vaisälä frequency and projection of the wind vector onto the stress direction at the surface ; σ_h is the subgrid scale variance of orography and K_{gwd} a dimensionless constant currently tuned at $3.5\ 10^{-3}$.

The linear deposition rate is computed under the assumption that the flux is just saturating the Lindzen criterium at the surface and being limited by the same during the upward propagation of the wave, every

time this leads to a reduction of the magnitude of the momentum flux vector. This gives the result :

$$\tau = -\rho.K_{gwd}.(N_s^2/N).(\mid U \mid^3/\mid U_s \mid^2).\sigma_h \qquad (32)$$

Like in all following refinements a system of numerical securities similar to the one used in the convection parameterization is introduced to avoid either acceleration of the wind or change of sign of its projection onto the stress direction. Furthermore the dependency of the surface wind evolution with respect to its own value is treated in a split-implicit way, like for instance in the vertical diffusion algorithm.

* Non linearities were introduced (*Mlynarz* (1990)) for two types of atmospheric behaviour : (i) trapping of the waves below an unstable layer ; the surface drag is then simply linearly removed between the surface and that level ; (ii) resonant damping or enhancement of the surface stress according to the interaction between the vertical wave length and the depth of the neutral layer from the surface up to the first layer where the Lindzen criterium acts to reduce the flux. The resonance has been parameterized with a very simple model where the only tuning parameter indicates the proportion in which maximum or minimum enhancement works ; it is currently set at $k_{res} = 0.6$. In case of decreased surface stress the non-acting layer simply extends a bit upwards. In case of increased surface stress the additional part is treated linearly like in the trapping case. This "resonant" effect has been introduced following the results of *Peltier and Clark* (1986).

* Surface anisotropy effects and a crude simulation of mountain blocking effects through an enhancement of the effective surface wind were combined to create a modified computation of $\mid U_s \mid$ (and of N_s for what concerns the second of the above-mentioned effects) :

+ Following the study made at ECMWF by Baines and Palmer of the suggestion of *Phillips* (1984) the introduction of orographic anisotropy effects has been recomputed and put in a formalism that allows to compute a fictitious wind (u',v') that would give the same surface stress than the one obtained by Phillips' method using (u,v) and a full algorithm.

$$u' = A.u + D.u^* \qquad (33a)$$
$$v' = A.v + D.v^* \qquad (33b)$$

with

$$u^* = u.\cos(2\alpha) + v.\sin(2\alpha) \qquad (34a)$$
$$v^* = v.\sin(2\alpha) - v.\cos(2\alpha) \qquad (34b)$$

α being the angle of the main gradient of the sub-grid scale orography with the "x" axis (the one of u,u^* and u') and

$$A = m_1^2+[4.(1-m_1).(1+\alpha_1.m_1) + m_1.\ln(1/m_1).(1+\alpha_2.m_1)]/(2.\pi) \quad (35a)$$
$$D = [4.(1-m_1).(1+\delta_1.m_1)-3.m_1.\ln(1/m_1).(1+\delta_2.m_1)]/(2.\pi) \quad (35b)$$

with

$$\alpha_1=(3+\ln(16))/4 \qquad \alpha_2=7.\pi/2-\ln(16)-8 \quad (36a/b)$$
$$\delta_1=(11-3\ln(16))/4 \qquad \delta_2=-\pi/4-\ln(16)+4 \quad (36c/d)$$

those last "odd-looking" coefficients resulting from "classical" approximations of elliptic integrals in the asymptotic cases $m_1 \rightarrow 0$ and $m_1 \rightarrow 1$. m_1, also called Γ^2, is the ratio of the orographic sub-grid scale variances between the direction characterised by α and its orthogonal one.

+ In order to simulate the blocking effect of mountain ranges with substantial sub-grid scale variance, one uses a vertical average of the wind and static stability (in the physical dimension of N^2, i.e. with possible negative contributions of unstable layers) to obtain a surface value for each of them. The wind and static stability profiles used for the computation above the surface level are linearly relaxed across the same depth starting from the mean value at the bottom to reach the local one at the top in order to avoid any jump at either of these two interfaces. The depth of the averaging layer is $H_{gwd} = H_{obst}.\sigma_h$ where the current value of the tuning constant H_{obst} is 2.

7.2 Surface fluxes and turbulent vertical diffusion

The surface and atmospheric turbulent exchange schemes are designed in continuity following the suggestion of *Louis* (1979). However the exact implementation of the algorithms is the one suggested in *Louis et al.* (1981) where a few additional "physical" constraints are put on the formulation and the interdependence of constants. Hence, apart from the rather arbitrary prescription of the asymptotic mixing length for momentum (the one for heat and moisture being obtained from the latter and the "d" parameter, see below) only four "universal" constants are controlling the intensity of vertical exchanges and its dependency on stability :

* the von Karman constant $x = 0.4$;

* the slope of the Monin-Obukov universal function for momentum at neutrality, slightly tuned to $b = 5$;

* the free-convection limit adimensionalised constant, slightly tuned to $c = 5$;

* the inverse of one and a half time the maximum flux Richardson number in the stable case, slightly tuned to $d = 5$.

The prescription of the asymptotic mixing length for momentum required several successive adjustments and it was found that a functional representation with three parameters, similar in spirit to the one of the HIRLAM group, was necessary. The three parameters are the asymptotic mixing lengths respectively at top and bottom of the atmosphere (or rather the bottom one λ and their ratio ß) and a "transition height" H_l, the formula for the mixing lengths themselves finally being :

$$l_{U/T} = (ß+(1-ß)/(1+((z+z_0)/H_l)^2))/(1/\kappa(z+z_0)+1/\lambda_{U/T}) \tag{37}$$

with

$$\lambda_T = \lambda_U.\sqrt{3d/2} \tag{38}$$

Currently used values for the parameters are λ_U = 75 m, λ_∞/λ = ß = 0.4 and H_l = 4000 m.

The roughness length over land is one of the geographical constants necessary to run the model, the production of which will not be mentioned here ; over sea however we are using the so-called Charnock formula with and additional "gustiness" stability dependent term. The latter has been introduced, following the same goals than *Miller et al.* (1992) but in a more simple framework, to ensure that, in low wind and unstable conditions the heat fluxes do not drop to zero, not for lack of a correct formula (the limit case is in principle taken into account by the Louis scheme), but simply for a vanishing roughness. Thus the formula now reads :

$$Z_0 = C_{hk}.u*^2/g + Z_{0cr}.C_D/C_N \tag{39}$$

where C_D and C_N are respectively the drag coefficient and its neutral counterpart. The currently used numerical values are 0.021 for the C_{hk} Charnock constant and 1.5 10^{-4} m for the critical "low wind" neutral roughness Z_{0cr}.

We are aware that the main weakness of the scheme just described is its lack of differentiation between roughness lengths for momentum and heat/moisture and this will be one of the next improvements we intend to prepare, test and eventually implement.

8. CONCLUSION

The all-purpose (NWP & Climate) parameterization package from which a good deal has been described above has evolved from the EMERAUDE code previously used at Météo-France (in two separate

versions) for large scale NWP and climate-type applications. The accent for this "ARPEGE conversion" was put on even more consistency and on additional physical representativity (ice phase cycle for example).

The most interesting feature is certainly the fact that only twenty-six tuning constants (with two interdependencies between them) are necessary to control all mentioned processes (a twenty-seventh and twenty-eighth one might be needed in the future for "two roughness lengths" and for "shallow convection cloudiness") in a physically sound and quite complete set of algorithms. Hence the tuning was made easier and it was even possible to find most of the time compromises that suited at the same time the NWP and climate-type applications (all parameters except the three of the stratiform precipitation have been retuned, while quasi-absolute faith was given to the "quasi-objective" procedure for computation of optical properties described in *Ritter and Geleyn* (1992)). Finally it should be mentioned that we elected to run the physics in parallel mode, each routine ignoring the tendencies computed by the other ones, as well for its input variables (all at t-δt) as for its algorithms, with the notable exception of the "humidity convergence" for the convective closure assumption.

The next actions to be undertaken are the above-mentioned differentiation of roughness lengths as well as search for a more specific "top of PBL radiative cloud amount parameterization" and the introduction of liquid and ice water as potential prognostic quantities (as transparent option for the rest of the code, following of course here also the underlying philosophy of all the Météo-France ARPEGE physical package!).

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