FULL-POS

SCIENTISTS' GUIDE

in ARPEGE/IFS cycle CY24T1

and

ALADIN cycle AL15 with ARPEGE/IFS cycle CY24T1

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Contents

1 INTRODUCTION

The so-called "Full-Pos" package is a post-processing package included inside Arpege/Aladin/Ifs ; it is replacing and extending the previous internal package ("Pos", from the subroutine having the same name).
Having an internal post-processing package enables to make the post-processed data during the model integration, as soon as a post-processing time step is reached. It also enables to have a full consistency between the model and the post-processing operators (spectral transforms, interpolators, ...), and to use the optimizations at disposal in the forecasting model.

The post-processing package provides an interface between the forecast (or assimilation) system and the databases for dissemination and archiving ; it is also used to elaborate historical files (given a historical file, by changing the geometrical parameters or the resolution) to be used in the (multi-) incremental variational analysis procedure, or to be used for running the limited area version Aladin.
In the former case it manipulates the model raw ouputs in order to present it in a more convenient and standard format for the users ; in the latter case, it manipulates the model raw ouputs and re-present it in another format, in order to fit the defined purpose.

Within the post-processing package, data can be interpolated vertically (from the model hybrid levels to other defined surfaces) and horizontally (from the model collocation grid to other defined grids) ; the data can also be filtered in the model spectral space (in order to remove numerical noise coming from the vertical interpolation, or to get smoother fields).
At the end, the data can be presented either as gridpoint fields, or as spectral fields (given the definition of the spectral geometry). But one should notice that all these operations (gridpoint vertical and horizontal interpolations, spectral filters and final representation) are not always exchangeable ; the next part will describe the actual possible combinations.

2 ORGANIZATION - ALGORITHMS

3 NOTATIONS

A, B : functions defining the vertical hybrid coordinate
c_p : constant pressure calorific capacity for moist air
c_{pd} : constant pressure calorific capacity for dry air
D : horizontal divergence
f : Coriolis parameter
g : acceleration of the gravity
L : number of model full levels
ℓ : index of a model full level
$\tilde{\ell}$: index of a model half level
p : pressure
p_0 : reference pressure equal to 1000 hPa
p_s : surface pressure
q : specific moisture
R : gas constant for moist air
R_d : gas constant for dry air
T : temperature
Γ_{st} : standard lapse rate
Φ : geopotential

Φ_s : surface geopotential
ζ : relative vorticity

4 ELABORATED FIELDS

The following fields, which are not primitive variables of the model, are calculated by specific operators.

4.1 Pressure

The upper air pressure at half levels is calculated from the surface pressure and the definition of the vertical hybrid coordinate :

$$p_{\tilde{\ell}} = A_{\tilde{\ell}} + B_{\tilde{\ell}} p_s$$

To compute the upper air pressure at full levels, we use :

$$\ln p = \frac{d}{dp}(p \ln p) - 1$$

In its discretized form :

$$\ln p_\ell = \frac{p_{\tilde{\ell}} \ln p_{\tilde{\ell}} - p_{\tilde{\ell}-1} \ln p_{\tilde{\ell}-1}}{p_{\tilde{\ell}} - p_{\tilde{\ell}-1}} - 1$$

Which simplifies into :

$$p_\ell = p_{\tilde{\ell}} e^{(-\alpha)}$$

where

$$\alpha_1 = 1 \qquad \text{and} \qquad \alpha_\ell = 1 - \frac{p_{\tilde{\ell}-1}}{\delta p_\ell} \ln \frac{p_{\tilde{\ell}}}{p_{\tilde{\ell}-1}} \tag{1}$$

The upper air pressure at half levels is computed in the subroutine `GPPREH`.

The coefficients α are computed in the subroutine `GPXYB`.
The upper air pressure at full levels is computed in the subroutine `GPPREF`.

4.2 Absolute vorticity

The absolute vorticity ζ_a is computed on the model levels as follows :

$$\zeta_a = \zeta + f$$

4.3 CAPE and CIEN

The calculation of the convective available potential energy (CAPE) or the convective inhibition energy (CIN or CIEN) is based on the following hypothesis :

- the parcel is assumed to ascent, following a dry adiabatic process, until it reaches the saturation ; then it is assumed to keep on ascending, following a moist adiabatic process
- the rising parcel does not mixt with the environment
- the parcel pressure is equal to the pressure of the environment at each level
- all the condensed water leaves the parcel as soon as it appears.

These assumptions can lead to significant errors when large displacement of parcels are involved, or at cloud edges, or for significant presence of water. However, measurements prove that they give good results in the core of a thunderstorm updraft.

CAPE and CIN are both computed in the subroutine `FPCAPE`.

4.4 Geopotential

The geopotential is calculated first on the model half levels, using the hydrostatic equation :

$$\Phi_{\tilde{\ell}-1} = \Phi_{\tilde{\ell}} - R_\ell T_\ell \ln \frac{p_{\tilde{\ell}-1}}{p_{\tilde{\ell}}}$$

Which leads, by summation, to :

$$\Phi_{\tilde{\ell}} = \Phi_s + \sum_{k=L}^{\ell+1} R_k T_k \ln \frac{p_{\tilde{k}}}{p_{\tilde{k}-1}}$$

Then the geopotential on full levels is computed as follows :

$$\Phi_\ell = \Phi_{\tilde{\ell}} + \alpha_\ell R_\ell T_\ell$$

where α is given by (1)

Remark : in the case of the non-hydrostatic model equations, the discretization of the logarithm of the layer depth $\ln \frac{p_{\tilde{\ell}}}{p_{\tilde{\ell}-1}}$ is replaced by $\frac{\delta p_\ell}{\sqrt{p_{\tilde{\ell}-1} \cdot p_{\tilde{\ell}}}}$

The geopotential on both half and full levels are computed in the subroutine `GPGEO`.

4.5 Gusts

4.6 ICAO jet pressure

4.7 ICAO Tropopause pressure

4.8 Iso-temperature surface

4.9 Mean sea level pressure

The calculation of mean sea-level pressure p_{msl} is based on assumption of a dry, hydrostatic subterranean atmosphere and a uniform lapse rate $\Gamma_{st} = -0.0065 K/m$. Actually this lapse rate is modified for warm surface temperatures to approximate the formulation developped at the US National Weather Service (Technical procedure Bulletin nr 57).
The calculation is invoked at points where the magnitude of the surface geopotential exceeds $10^{-3} J/kg$, otherwise p_{msl} is set equal to the surface pressure p_s.

Assuming a uniform lapse rate $\partial T / \partial \Phi = \Gamma_{st}/g$
and the the hydrostatic equation $\partial \Phi / \partial p = -R_d T/p$,
a surface temperature T_* is first computed :

$$T_* = T \left(1 + \Gamma_{st} \frac{R_d}{g} \left(1 - \frac{p_s}{p} \right) \right)$$

To ensure the independency between the model levels and the level from which the temperature is extrapolated, we first choose the height from which the temperature should be extrapolated, then we convert this height into the nearest model level.
This height is the variable `HEXTRAP` which is tunable. The default value is 150. m for ECMWF and 100. m for METEO-FRANCE.

We also compute a standard surface temperature : $T_0 = T_* - (\Gamma_{st} \Phi_s / g)$

The surface temperatures T_* and T_0 are computed in the subroutine `CTSTAR`.

From the previous formulae and with the convention : $\alpha = R_d \left(T_0 - T_*\right)/\Phi_s$, we obtain :

$$T = T_* \left(\frac{p}{p_s}\right)^\alpha$$

$$\Phi = \Phi_s - \frac{R_d T_*}{\alpha}\left[\left(\frac{p}{p_s}\right)^\alpha - 1\right]$$

Applying the last formula to $\Phi = 0$, we obtain :

$$p_{msl} = p_s \left(1 + \frac{\alpha \Phi_s}{R_d T_*}\right)^{1/\alpha}$$

The actual calculation of p_{msl} uses an approximation valid for small α :

$$p_{msl} = p_s . \exp\left[\frac{\Phi_s}{R_d T_*}\left(1 - \frac{y}{2} + \frac{y^2}{3}\right)\right] \qquad with \quad y = \frac{\alpha \Phi_s}{R_d T_*}$$

Notice :
To inhibit low extrapolated pressures under hot elevated terrain T_* and T_0 (and consequently α) are modified as follows :
if $T_* < 290.5$ K and $T_0 > 290.5$ K then $T_0 = 290.5$ K
if $T_* \geq 290.5$ K and $T_0 > 290.5$ K then T_* is replaced by $\left(T_* + 290.5\right)/2$ and $T_0 = T_*$

In a similar spirit, to inhibit high pressures under cold terrain,
if $T_* < 255$ K then T_* is replaced by $\left(T_* + 255\right)/2$

The medium sea level pressure is computed in the subroutine `PPPMER`.

4.10 Potential temperature

The potential temperature is defined as follows :

$$\theta = T \left(\frac{p}{p_0}\right)^{-\kappa} \qquad \text{where} \qquad \kappa = \frac{R}{c_p}$$

Which leads naturally to the discretized form :

$$\theta_\ell = T_\ell \left(\frac{p_\ell}{p_0}\right)^{-\kappa_\ell}$$

The potential temperature is computed in the subroutine `GPPVO`.

4.11 Equivalent potential temperature

4.12 Potential Vorticity

4.13 Montgomery potential

The Montgomery potential is computed by neglecting the variations of R and c_p :

$$\Upsilon = \Phi + c_{pd} T = \Phi + c_{pd} \theta \left(\frac{p}{p_0}\right)^{\kappa_d} \qquad \text{where} \qquad \kappa_d = \frac{R_d}{c_p}$$

This formulation is specially adapted to post-process the Montgomery potential on θ surfaces ; therefore one should notice that the code is restricted to the computation of Υ on θ surfaces (see the subroutine `POS`).
Therefore the final formulation will read :

$$\Upsilon_{\theta=cste} = \Phi_{\theta=cste} + c_{pd}\theta \left(\frac{p_{\theta=cste}}{p_0}\right)^{\kappa_d}$$

4.14 Non-hydrostatic pseudo-vertical Divergence

After Bubnova et al, and following the same notations, the non-hydrostatic pseudo-vertical Divergence is defined as follows :

$$d = -\frac{\partial \phi}{\partial \eta}\left[-\frac{C_v}{C_p}\frac{\dot{\pi}}{\pi} - \nabla \bullet \mathbf{V} + \nabla \phi \bullet \left(\frac{\partial \mathbf{V}}{\partial \phi}\right)\right]$$

4.15 Relative humidity

4.16 Shearing Deformation

4.17 Stretching Deformation

4.18 Surface Vertical Velocity

4.19 Total water vapour

The total water vapour above a given surface point is expressed as follows :

$$Q = \frac{1}{g}\int_{p_s}^{0} q\,dp$$

Which leads to the discretized expression :

$$Q = \frac{1}{g}\sum_{k=1}^{L} q_\ell \delta p_\ell$$

4.20 Tropo. Folding Indicator

4.21 True Vertical NH Velocity

After Bubnova et al, and following the same notations, the non-hydrostatic true vertical velocity w is defined by :

$$d = -g\left(\frac{\partial w}{\partial \eta}\right)$$

And on the surface :

$$w_{\tilde{L}} = \frac{1}{g}(\mathbf{V} \bullet \nabla \phi)_{\tilde{L}}$$

The discretization leads to :

$$w_{\tilde{\ell}-1} = w_{\tilde{\ell}} + \frac{d_\ell}{g} \qquad \text{and} \qquad w_{\tilde{L}} = \frac{1}{g}\left[u_{\tilde{L}}\left(\frac{\partial \phi}{\partial x}\right)_{\tilde{L}} + v_{\tilde{L}}\left(\frac{\partial \phi}{\partial y}\right)_{\tilde{L}}\right]$$

4.22 Vertical velocity

4.23 Wet bulb potential temperature

4.24 Wet bulb temperature

4.25 Wind velocity

5 VERTICAL INTERPOLATIONS

6 SPECTRAL FILTERING

7 HORIZONTAL INTERPOLATIONS