# Whole Atmospheric Global Balanced Wind and Thermodynamic States from the Ground to the Lower Thermosphere

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## 1. Introduction

Korea Polar Research Institute has been operating instruments to observe and estimate atmospheric state variables in the mesosphere and thermosphere in the polar regions (Kim et al. 2012). It is also found that those observations can give insights on gravity wave activities in the upper atmosphere (Lee et al. 2013). To understand more precisely how the observed upper atmospheric phenomena are generated or propagated from the other regions, numerical modeling and its comparison with observations are necessary.

Comparison between model results and observations for specific events requires numerical model to start from a time prior to specific observation date and time. For this, a proper initialization of model is essential. Without appropriate initialization, spurious initial high-frequency noises due to unbalanced atmospheric states contaminate model results even at a later time (Daley 1991).

Purpose of the present research is to generate balanced atmospheric state variables from the ground to the lower thermosphere at specific date and time. Since our major interest is in simulating already observed phenomena rather than in forecasting, the state variables are obtained using pre-existing analysis data for the troposphere and stratosphere. Above the stratopause, empirical horizontal wind and temperature models are used.

The separate data sets and empirical model results are combined using data-fusion method. Finally, a balanced atmosphere state is derived from the combined data at a specific date and time using a global primitive equation dynamical model. Using global model is preferred to regional model since upper atmospheric phenomena such as gravity waves observed at a certain location can be generated in the other regions far from the observation site (Liu et al. 2014).

## 2. Data and Methodology

Variables required for global modeling should contain at least zonal and meridional wind components ($U$ and $V$), temperature ($T$), specific humidity ($Q$), and surface pressure ($P_s$) at given date and time. Below the lower mesosphere, those variables are obtained from the ERA-Interim (Dee et al. 2011) and MERRA (Rienecker et al. 2011) analysis data. The standard pressure-level versions of those analysis data extend from the ground up to 1 hPa and 0.1 hPa for ERA-Interim (ERA hereafter) and MERRA, respectively. For MERRA analysis, variables above 400 hPa are only to avoid topographic mask.

Above 1 hPa, empirical wind and temperature models are also used to specify state variables. The horizontal wind model is composed of two parts: Horizontal wind for geo-magnetically quiet time (HWM14: Drob et al. 2015) and disturbed wind (DWM07: Emmert et al. 2008). The quiet-time wind represents zonal-mean, stationary planetary waves and three harmonics of the diurnal tides. Empirical temperature model (NRL's MSISE-00; Picone et al. 2002) provides temperature that depends on solar flux indicated by 10.7 cm radio signals from the sun, but HWM14 does not include the solar-flux effects.

A contiguous whole atmospheric global data set is made through the following two procedures: (i) the MERRA analysis data are horizontally interpolated into the ERA 0.75°x0.75° grid, and two empirical model results are generated at the identical horizontal grid. Empirical models are computed at pre-defined pressure levels from 1hPa to above. (ii) Data and model results with their own vertical coverages and some overlapped layers between 400-1 hPa and 1-0.1 hPa are fit to a smooth curve represented by a linear combination of the 3rd order B-spline basis functions (Piegl and Tiller 1997).

## 3. Global Dynamical Model

A dynamically balanced atmospheric state is obtained using a global dynamical model initialized with the whole atmospheric data set described in the previous section. The dynamical model is based on the spectral element (SE) hydrostatic dynamical core implemented in NCAR Community Atmospheric Model (Dennis et al. 2012). In this research, the SE core is run on the cubed-sphere grid of the ne16np4 (~2°) horizontal resolution. Firstly, the whole atmospheric global data set is horizontally remapped to the model horizontal grid, and then the remapped data are vertically interpolated into the model vertical grid.

The SE core is formulated based on the mass-based (i.e., $P_s$) hybrid vertical coordinate. Therefore, model surface pressure $P_{sm}$ should be given in advance to define model vertical coordinate. The $P_{sm}$ is obtained using the ERA's $P_s$ after the correction of discrepancies in the topographic heights between the dynamical model and ERA. In case that model topography is below the ERA's ground, the tropospheric mean temperature lapse rate is assumed and then $P_{sm}$ is computed using the hypsometric equation. Finally, to prevent initial instability, negative static stability is removed. The static stability is computed using the potential temperature for unsaturated moist air.

To achieve a balanced state using the global primitive equation model, we follow an approach similar to Kasahara (1982). That is, atmospheric total mass ($P_s$) and temperature ($T$) are assumed given by values determined above (as a result, the geopotential height is also given), and then only the horizontal momentum equations are numerically integrated in time along with the diagnostic computation of the vertical velocity ($\dot{\eta}$):

$$\frac{\partial \mathbf{v}}{\partial t} + (\zeta + f)\hat{\boldsymbol{k}} \times \mathbf{v} + \boldsymbol{\nabla}_{\eta}\left(\frac{1}{2}\mathbf{v}\cdot\mathbf{v} + \Phi\right) + \dot{\eta}\frac{\partial \mathbf{v}}{\partial t} + \frac{R_d T_v}{p}\boldsymbol{\nabla}_{\eta} p = D_h + D_v - \left(\frac{\partial \mathbf{v}}{\partial t}\right)_{\text{specified}} \quad (1)$$

$$\frac{\partial p}{\partial \eta}\dot{\eta} + \int_{\eta_t}^{\eta} \boldsymbol{\nabla}_{\eta}\cdot\left(\frac{\partial p}{\partial \eta'}\mathbf{v}\right) d\eta' = 0 \quad (2)$$

where $(\partial \mathbf{v}/\partial t)_{\text{specified}}$ is pre-computed using the whole atmospheric global atmospheric data set, described in section 2, at three different times. The time interval between two data sets is set equal to 5 min to resolve properly the terdiurnal wind variation of the HWM14.

The term $D_v$ in (1) is the vertical diffusive forcing, and is simply parameterized as follows:

$$D_v = -\frac{\mathbf{v} - \mathbf{v}_{ini}}{\tau} \quad (3)$$

where $\mathbf{v}_{ini}$ is the whole atmospheric initial wind; and $\tau$ is the relaxation time scale.

By integrating (1) in time, we obtain steady-state (balanced) wind components. Depending on $\tau$, the final balanced state approaches the initial condition quickly (for small $\tau$) or approaches a new wind state (for $\tau$ = 1-2 hr) within 1 day. For very large $\tau$, a balanced state is not obtained (model system diverges).

## 4. Results

Figure 1 shows the initial and two balanced zonal wind at z = 90 km computed for two different $\tau$s.

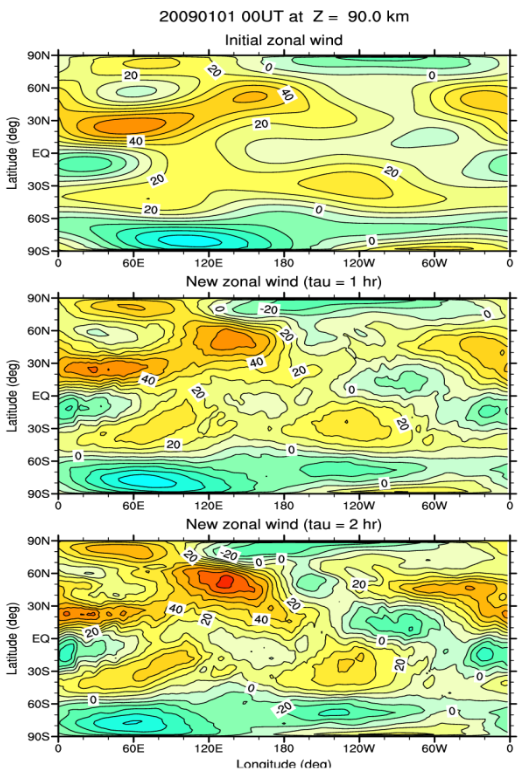


Figure 1: Initial and Balanced wind at z = 90 km

Balanced wind is generally a little stronger than the initial wind but seems acceptable. For the balanced flows shown in Fig. 1, the magnitude of the parameterized vertical diffusive forcing term ranges roughly from 50 to 120 m $s^{-1}$ $day^{-1}$ above the middle mesosphere, which is reasonable and strong enough to reverse the vertical wind shear in the upper mesosphere (not shown).

Difference between initial and balanced states is large especially in the summer polar lower thermosphere (i.e., southern polar regions). This difference may be accounted for by the solar flux effects included in the MSIS temperature alone. Further results will be presented and discussed at conference.

## 5. Acknowledgement


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