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Alain Hauchecorne, Alexandre Maillard

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A 2-D DYNAMICAL MODEL OF MESOSPHERIC TEMPERATURE INVERSIONS IN WINTER

Alain Hauchecorne and Alexandre Maillard

Service d'Aéronomie du CNRS, Verrières-le Buisson, France

Abstract : A 2-D stratospheric and mesospheric dynamical model including drag and diffusion due to gravity wave breaking is used to simulate winter mesospheric temperature inversions similar to those observed by Rayleigh lidar. It is shown that adiabatic heating associated to descending velocities in the mesosphere is the main mechanism involved in the formation of such inversions. Sensitivity tests are performed with the model and confirm this assumption. It is also explained why other previous similar studies with 2-D models did not show mesospheric inversion layers.

Introduction

A strong inversion layer is very often observed in mesospheric temperature profiles with a relative minimum of temperature around 70 km, and a positive temperature gradient in the few kilometers above this minimum. This feature has been first observed by rocket measurements and reported by Schmidlin [1976] but without explanation of the phenomenon. More recently local deficiencies of the density, associated with temperature inversion layers, were observed from the reentry data of the US space shuttle [Champion, 1986]. The first statistical study of these inversion layers has been made with a large data base of more than 1000 temperature profiles obtained by two Rayleigh lidars located in the south of France [Hauchecorne et al., 1987, hereafter HCW]. This study showed that the probability of occurrence of an inversion presents a maximum of 70 % in winter. These inversions are observed at the same altitude during several days and usually simultaneously above the two lidar sites (distance between the two sites: 550 km). The similarity between the characteristics of the inversions and of the MST radar echoes in the mesosphere, leads HCW to conclude that both phenomena have the same origin, that is to say the breaking of upward propagating internal gravity waves. Using a crude estimate of the amplitude growth with height of a gravity wave, HCW showed that gravity waves break preferably inside the inversion layer and then maintain this layer. More recently Clancy and Rusch [1989] showed from the Solar Mesosphere Explorer data that these inversions exist at middle latitude in both hemispheres in winter with a higher amplitude in the southern hemisphere.

Proposed mechanism

The usual mesospheric temperature field is far from the radiative equilibrium. Above about 65 km the mesosphere is colder in summer than in winter and a reversal of the wind is observed near the mesopause level with easterlies in winter and westerlies in summer. The role of gravity waves dissipation to maintain the observed mesospheric circulation is now well recognized [Lindzen, 1981; Matsuno, 1982; Holton, 1982]. The wave drag due to gravity wave breaking induces a meridional circulation from the summer pole to the winter pole and an upward (resp. downward) motion at the summer (resp. winter) pole with an adiabatic cooling (resp. heating). Observations of gravity waves by lidars and radars [HCW; Manson and Meek, 1987] indicate that wave breaking occurs in relatively well defined layer, in a region of decreasing wind above the mesospheric jet, where the probability for the waves to reach a breaking level is high. The resulting adiabatic heating provides most of the energy necessary to the development of a temperature inversion in the winter hemisphere. The so-formed temperature inversion may be enhanced by the heating due to the dissipation of turbulent kinetic energy in the warm layer. As shown in HCW, the persistence of inversion layers during a few days is also favoured by the preference of the waves to break in a layer of increasing stability (dN/dz >0).

In all the previous simulations of gravity waves breaking in 2-D or 3-D dynamical models, the deposition of wave momentum gradually occurs in the whole mesosphere and leads to a smooth decrease of the temperature from the stratopause to the mesopause, but without inversion layer. We now present a 2-D model which generates this feature and we estimate the role of different physical mechanisms involved in mesospheric inversion layers.

Simulation of the inversion layer

A 2-D latitude-altitude dynamical model is used in order to simulate a mesospheric inversion layer in the winter hemisphere and to test the ideas presented in the previous section. The model is in log-pressure coordinates, and it is driven by the set of equations (1-5):

$$\frac{\partial\bar{u}}{\partial t} - f\bar{v} = F_x + \frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0 D\frac{\partial\bar{u}}{\partial z}\right) \tag{1}$$

$$\frac{\partial\bar{v}}{\partial t} + f\bar{u} + \frac{\partial\bar{\Phi}}{\partial y} = \frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0 D\frac{\partial\bar{v}}{\partial z}\right) \tag{2}$$

$$\frac{\partial\bar{T}}{\partial t} = -N^2\frac{H}{R}\bar{w} - \alpha\,(\bar{T} - \bar{T}_R) + \frac{e^{\kappa z/H}}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0 D\frac{\partial\bar{\theta}}{\partial z}\right) + \varepsilon \tag{3}$$

$$\frac{1}{\cos\phi}\frac{\partial}{\partial y}(\bar{v}\cos\phi) + \frac{1}{\rho_0}\frac{\partial}{\partial z}(\rho_0\,\bar{w}) = 0 \tag{4}$$

$$\frac{\partial\bar{\Phi}}{\partial z} = \frac{R\bar{T}}{H} \tag{5}$$

Bars denote zonal means, u, v and w are the horizontal, meridional and vertical velocities, T is the temperature, T_R is the initial temperature, θ is the potential temperature, ε is a term due to turbulent heating, ρ_0 is the density, Φ is the geopotential height, ϕ is the latitude, α is a Newtonian cooling rate coefficient, f is the Coriolis parameter, H is the scale height, R is the gas constant, N is the log-pressure

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buoyancy frequency; y is equal to $a\phi$ where a is the radius of the earth, z is equal to $-H\ln(p_0/p)$. Equation (3) assumes that the thermodynamical quantity which is diffused is potential temperature. Details concerning these equations are given for example in Schoeberl et al. [1983].

We use a slight extension of the Lindzen [1981] parameterization of the zonal drag and diffusion due to gravity wave breaking. In the notation of Holton [1982, 1983], we consider a zonally propagating wave, in terms of the zonal mean flow $\bar{u}$, the buoyancy frequency N, the scale height H and the wave parameters c, k and B; c and k are the zonal phase speed and wavenumber, B is the vertical perturbation velocity amplitude of the gravity wave at the tropopause. The breaking level of such a wave is given by:

$$z_b = 2\,H\ln\left\{\frac{(\bar{u}-c)^{3/2}(u_0-c)^{1/2}\,k}{B\,N}\right\} \qquad (6)$$

or:

$$z_b = 3\,H\ln\left\{\frac{|\bar{u}-c|}{\tilde{u}}\right\}$$

where:

$$\tilde{u}^{3/2} = B\,N/\left(k\,|u_0-c|^{1/2}\right)$$

$\tilde{u}$ is therefore depending on N at the power 2/3. We assume that :

$$\tilde{u} = K\,(N/N_0)^{2/3}$$

The scale height for the dissipation of wave energy is given by [McIntyre, 1989]:

$$H_{diss} = \left(\frac{1}{H} - 3\,\frac{d\bar{u}/dz}{(\bar{u}-c)} + \frac{1}{N}\frac{dN}{dz}\right)^{-1} \qquad (7)$$

The turbulent heating is given by:

$$\varepsilon = \frac{k\,(\bar{u}-c)^4}{2\,N\,H_{diss}} \qquad (8)$$

The turbulent diffusion coefficient is therefore related to ε by the relation [Ebel, 1984]:

$$D = \beta\varepsilon/N^2 \qquad (9)$$

where β is a dimensionless number smaller than 1.

The drag coefficient due to vertical momentum transfer can be expressed as:

$$F_x = \frac{\varepsilon}{\bar{u}-c} \qquad (10)$$

or:

$$F_x = -A\,(\bar{u}-c)^3/H_{diss}$$

In the reference simulation, the model has been run for 15 days of simulation with three waves of respective phase speeds -20, 0 and +20 m/s. For the three waves, A is a constant having the numerical value 0.8 10^{-8} $m^{-2}s$, β has the numerical value 0.3 (this choice is discussed in the conclusion), K is equal to 0.6 ms^{-1} and N_0 is equal to 2 10^{-2} s^{-1}. The initial wind field is supposed as linearly increasing with altitude, due to the fact that at the beginning of the simulation there are no gravity waves; a temperature profile is prescribed at the equator, thus giving the wind and temperature shown in figure 1 and figure 2 (we assume that we are under winter

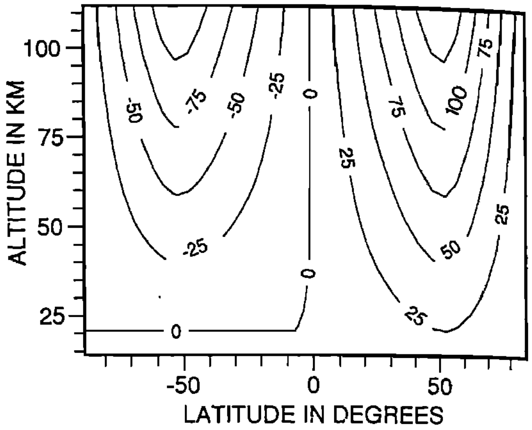

Fig.1 - Initial zonal wind field (m/s).

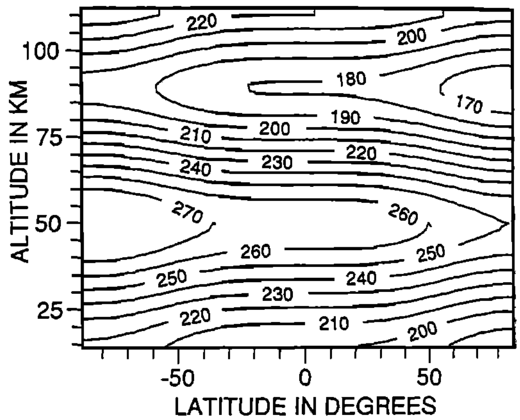

Fig. 2 - Initial temperature field (K).

conditions in the Northern Hemisphere); note the winter polar mesopause, colder than the summer one.

As the model runs the wave breaking progressively produces a realistic inversion of the meridional gradient of temperature between the summer and winter polar mesopause (figure 3), associated with the formation of realistic jets in the mesosphere, i.e. westerly winds in the Northern Hemisphere (winter) and easterly winds in the Southern Hemisphere (summer) (figure.4). Furthermore a temperature inversion

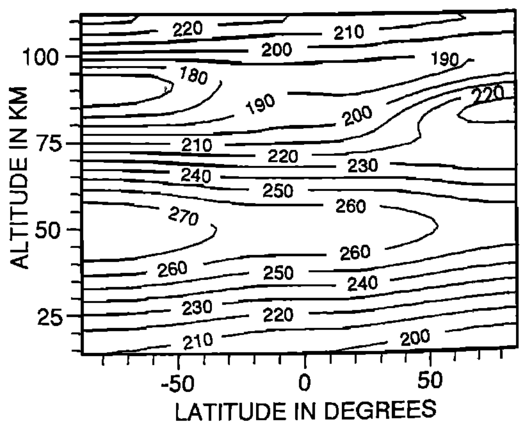

Fig. 3 - Temperature field (K) after 15 days in the reference simulation.

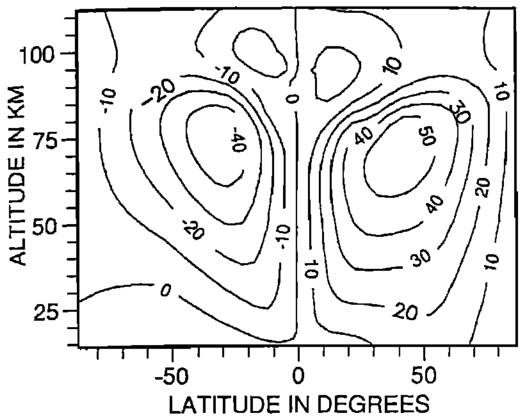

Fig. 4 - Zonal wind field (m/s) after 15 days in the reference simulation.

layer is generated in the altitude range 75-85 km between 50 °N latitude and the North Pole. The evolution of the temperature profile at 60 °N latitude (figure 5) shows the development of a stable warm layer of temperature (the so-called mesospheric temperature inversion) below the mesopause, with a relative maximum at about 80 km. The wave drag associated with this inversion (figure 6) is localized in the mesospheric region above 75 km altitude, which is 15 km higher than in Holton [1983]; this difference is discussed below.

In order to understand which are the most important physical processes in the formation of temperature inversions, a series of tests has been made.

First the model was tested with only one wave. The only kind of wave able to produce heating in the Northern Hemisphere must have a zero or negative phase speed. Of course there is no inversion of the mesospheric gradient of temperature between the poles, since such waves cannot propagate in the summer hemisphere, but mesospheric inversions occur in the Northern Hemisphere with phase speeds c equal to 0 or to -20. The amplitude of the inversion is weaker with only one wave, and is found stronger with c=-20 than with c=0. The inversion also occurs when five waves are included with phase speeds c=-30, -15, 0, 15 and 30 ms^{-1}.

Another kind of sensitivity tests was made: in the thermodynamical equation (3) each term was suppressed alternatively. The inversion layer is slightly weaker when the energy

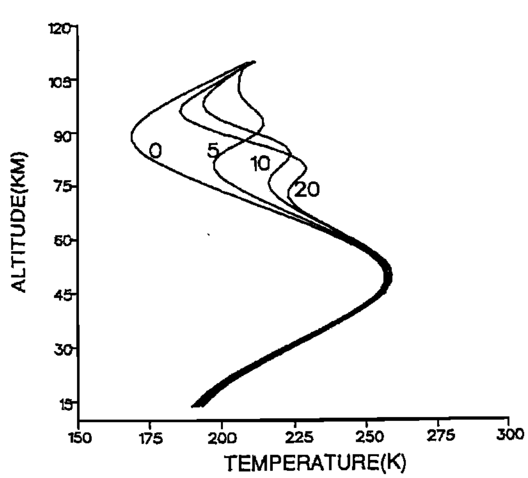

Fig. 5 - Evolution of the temperature profile at 60°N latitude during 20 days of simulation. The labels indicate the day of the corresponding profile.

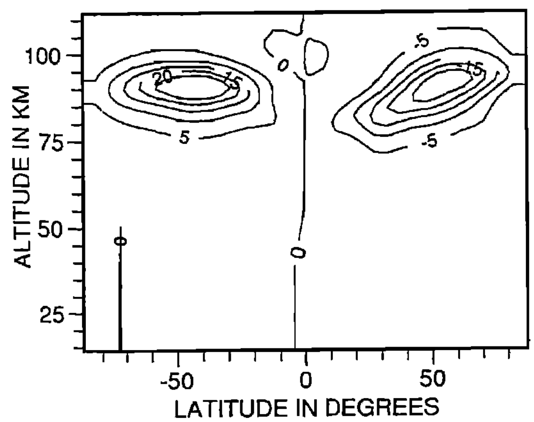

Fig. 6 - Drag (in m/s/day) after 10 days of simulation.

dissipation is suppressed and not significantly affected when the turbulent diffusion is cut. If the Newtonian cooling is suppressed, the inversion also remains, but of course the mean temperature slowly increases at each altitude level. In contrast the inversion disappears completely when the adiabatic warming due to the vertical velocity is suppressed.

The thermal budget of the atmosphere has been made at 60 °N for the day 15 (figure 7) and confirms the previous analysis: the adiabatic heating is the most important term for the formation of the warm layer and is partially compensated by the Newtonian cooling in the altitude range where the inversion occurs. The diffusion and the turbulent heating are clearly less important in this process.

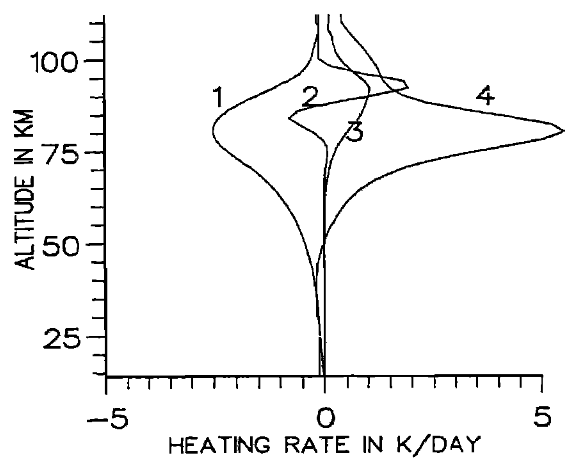

Fig. 7 - Thermal budget at 60°N latitude on day 15.
curve 1: Newtonian cooling.
curve 2: Potential temperature diffusion.
curve 3: turbulent heating.
curve 4: adiabatic heating.

Another important factor in the formation of the inversion is the narrowness of the layer of wave breaking. This narrowness is depending on the value of K (or $\tilde{u}$ if $N=N_0$) which determines the amplitude of the wave u' for which a breaking level is reached ($u'=\bar{u}-c$). A simulation has been made with a value of K equal to 1.2 ms^{-1} instead of 0.6 ms^{-1}, similar to the $\tilde{u}$ values used in Holton [1982, 1983] ranging from 1.2 to 2 ms^{-1}. The resulting temperature field obtained after 15 days (figure 8) shows a result similar to previous 2D models, that is to say a cold mesopause at summer pole and

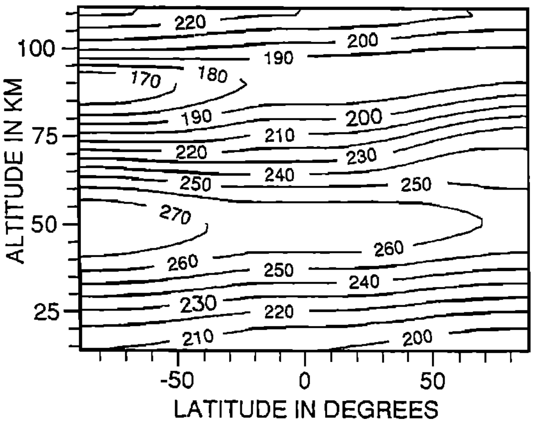

Fig. 8 - Temperature field (K) obtained with K=1.2 ms^{-1} after 15 days of simulation.

warm at winter pole but without temperature inversion. Note that a value of 1.6 ms^{-1} for K produces an amplitude for the wave of the order of 45 ms^{-1} if this wave breaks at 70 km, which is probably too large. A smaller value for K is therefore certainly more realistic.

Discussion and conclusion

This study shows that a 2-D dynamical model, taking into account the drag, diffusion and dissipation due to gravity wave breaking, is able to produce the formation of a stable mesospheric inversion of temperature in winter. The main mechanism that produces this layer is an adiabatic warming associated to downward mean vertical velocities of the order of 1 to 2 cms^{-1} in the region 75-85 km. This vertical velocity is in the same order than in previous models [e.g. Holton, 1983]. The only difference is that in our case the wave breaking and then the strong downward motion are limited to the upper mesosphere. The narrowness of the layer of wave breaking is a necessary condition for the formation of a temperature inversion.

Recent studies [McIntyre, 1989] pointed out that the concept of eddy diffusivity might suffer some shortcomings in certain conditions and that some precautions should be taken in future models including wave drag parameterizations. A gravity wave parameterization such as the one presented in this paper probably overestimates the values of the diffusion, even with the value we choose for β (0.3 instead of 1 in the original Lindzen parameterization). Strobel et al. [1987] have shown from measurements of vertical constituent transport in the mesosphere that the eddy Prandtl number Pr used to describe the constituent and potential temperature diffusion must be of order 10 to explain the data, which corresponds to β=1/Pr=0.1. Nevertheless we have shown that even if we neglect the potential temperature diffusion, the inversion occurs.

Though this first modelling of mesospheric inversions is encouraging, two important questions remain. First, further studies will have to explain how mesospheric inversions are able to occur at middle latitudes in the real atmosphere, which is not the case in this model. The use of a 3-D model might partially answer this question. One idea is that a displacement of the jet, due for example to planetary waves, could influence the location of mesospheric inversions. A second problem is that summer mesospheric inversions are not simulated with this model; these inversions are more sporadic in the real atmosphere and not very stable. A 2-D mechanistic model with a constant source of gravity waves is probably not able to reproduce such summer inversions.

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A. Hauchecorne and A. Maillard, Service d'Aéronomie du CNRS, B.P.3, 91371 Verrières-le-Buisson cédex, France.

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