Mausam, (1987), **38**, 1, 89-92

551.513.1 : 551.511

The effect of diabatic heating on the change of surface pressure

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(*Received 29 November 1984*)

सार — वायुमंडलीय डायाबेटिक तापन के प्रभाव पर विचार करने के लिए दाब प्रवृति के समीकरण को संशोधित किया गया। विभिन्न तुंगताओं के लिए अनुमानित दबाव में तापन का समावेशन अच्छा खासा अंतर बताता है। अंशत: स्वच्छ आकाश की दशाओं के 32 मामलों का परीक्षण किया गया। ऐसा प्रतीत होता है कि 300/100 मि. बार के स्तरों का प्रभाव दबाव विभिन्नताओं पर सबसे अधिक पड़ता है।

ABSTRACT. The equation of pressure tendency has been modified to consider the effect of atmospheric diabatic heating. The inclusion of heating shows a marked difference in estimated pressure for different altitudes. Thirtytwo cases of partially clear sky conditions were examined. The layer 300/100 mb seems to have the greatest effect on pressure variation.

1. Introduction

Diabatic heating is the forcing term of the equation of thermodynamics. It can be presented as :

$$\dot{Q} = \frac{\partial T}{\partial t} + v.\nabla T + w\left(\frac{\partial T}{\partial p} - \frac{RT}{c_p p}\right)$$

where the notation is in a p-coordinate system and $\dot{Q}$ is the heating (cooling) rate per unit mass.

The term $\dot{Q}$ can usually be divided into three parts :

$$\dot{Q} = Q_r + Q_d + Q_L \qquad (1)$$

where,

Q_r is heating (cooling) due to radiation, Q_d is heating (cooling) due to eddy diffusivity, and Q_L is due to heating released by condensation.

$$Q_r = q_1 + q_2$$

q_1 is the rate due to long wave radiation and

q_2 is due to absorbed solar radiation.

$$q_1 = -(g/c_p)\, \partial F/\partial p$$

and

$$q_2 = -(g/c_p)\, \partial Q/\partial p$$

where,

F is the infrared radiation flux and Q is solar radiation flux. The method of calculation follows Danard (1969).

The downward short wave radiation at level i was calculated at each level (i) using the equation :

$$Q_i = S\,(1 - A\,[W_i \sec(Z)]) - S\,(1 - A\,[W_0 \sec(Z)]) \times \alpha_s\,(1 - A\,[1{\cdot}66\,(W_0 - W_i)]) \qquad (2)$$

where,

W is the optical path, 0 refers to the surface, Z is the zenith angle, A is the absorptivity due to water vapour and α_s is the surface albedo. This formula was given by Joseph (1966). Only clear sky hours were considered in order to disregard the effect of cloud on the absorption and reflection of incoming radiation.

For sake of simplicity, the effect of eddy diffusivity was disregarded in this study. The heat flux due to effect of moisture content can be considered as

$$Q_L = H_{ce} + H_{sc}$$

TABLE 1

Pressure tendency at different layers

	Layers						
	1000/850	850/700	700/500	500/300	300/200	200/100	100/50
	(a) Average of 32 cases						
$\triangle P$	2.4	2.3	2.5	2.3	2.2	.9	.3
$\triangle P^*$	2.47	2.37	2.59	2.41	2.4	.98	.32
	(b) Case of 27 March 1980						
$\triangle P$	3.1	5.2	7.3	9.1	—9	—7.3	—2.1
$\triangle P^*$	2.7	4.4	6.4	8	—6.7	—6.6	—1.8

*indicates the inclusion of diabatic effects.

$\triangle P$ is the mean of absolute change of pressure (mb) during 24 hr

$P_s = 7$ mb; $P_s^* = 8.3$ mb, P_s indicate change of surface pressure

H_{cc} is the latent heat released by convection and was calculated at each level using :

$$H_{cc} = R_t\,[(P_0/P)^{R/c_p}\,(T_c - T/\triangle t) + w\,\partial\theta/\partial p].$$

The vertical velocity w was calculated by method proposed by Abdel-Wahab (1981).

The method of computation was based upon Kanamitsu (1975) where R_t can be defined as

$$R_t = \frac{g\,(1-B)\,I}{\int_{P_{ct}}^{P_{cb}} \frac{c_p}{L}\left[\frac{T_c - T}{\triangle t} + \left(\frac{p}{p_0}\right)^{R/c_p} w\,\frac{\partial\theta}{\partial p}\right] dp}$$

where,

I is the net convergence of moisture in column of air and B is defined as

$$B = \frac{\int_{P_{ct}}^{P_{cb}} \left[-\frac{c_p}{L}\left(\frac{p}{p_0}\right)^{R/c_p} \times \frac{\partial\theta}{\partial p}\,w - w\,\frac{\partial q}{\partial p}\right] dp}{gI}$$

where, $I = \frac{1}{g}\int_{P_{ct}}^{P_{cb}} w\,\frac{\partial q}{\partial p}\,dp$ and

T_c is the temperature at cloud level. The stable heating rate (H_{sc}) was computed using a simple expression because of its smaller relative magnitude in comparison with other terms :

$$H_{sc} = -\,(L\,\theta/c_p\,T)\,(\partial\,q_s/\partial p)$$

where q_s specific humidity (Chang 1980). This term shows significance in moist atmosphere conditions.

The effect of the Q_d term will be disregarded in calculating the diabatic term. A discussion of the contributions of heating terms and other terms will follow.

2. Analysis of pressure tendency equation

Assuming that P_s is the surface pressure, then the rate of change of P_s on a flat surface may be written as :

$$\frac{\partial P_s}{\partial t} = -\int_0 g\,\nabla\,.\,\rho\,v\,dz \tag{3}$$

This equation is used in the diagnosis of tropical disturbances as Gray (1968).

Assuming a parcel of air with density (ρ), mass (M) and volume (V) then,

$$\partial\rho/\partial t = (1/V)\,\partial M/\partial t - (M/V^2)\,\partial V/\partial t \tag{4}$$

and using the equation of state, we can simply get

$$\frac{\partial\rho}{\partial t} = (1/V)\,\partial M/\partial t + (1/RT)\,\partial p/\partial t - (P/RT^2)\,\partial T/\partial t \tag{5}$$

It is easy to show that variation of mass with time for volume V is :

$$(1/V)\ \partial M/\partial t = -1/V_s \iint_s \rho V_n\ ds \qquad (6)$$

where, s is the corresponding area an V_n is the normal wind component. When $V \to 0$, we get :

$$\lim_{V\to 0} (1/V)\ \partial M/\partial t = -\operatorname{div} \rho V \qquad (7)$$

From Eqn. (5) we can have

$$\frac{\partial \rho}{\partial t} = -\operatorname{div} \rho V + \frac{1}{RT}\frac{\partial p}{\partial t} - \frac{p}{RT^2}\frac{\partial T}{\partial t} \qquad (8)$$

using the hydrostatic equation, we can get

$$\frac{\partial^2 p}{\partial t\, \partial z} + \frac{g}{RT}\ \frac{\partial p}{\partial t} - \frac{gp}{RT^2}\ \frac{\partial T}{\partial t} - g \operatorname{div} V\rho = 0 \qquad (9)$$

The term $(\partial T/\partial t)$ can be replaced by

$$\frac{\partial T}{\partial t} = -v \cdot VT + w(\gamma - \gamma_a) + \frac{\gamma_a}{g\rho}\frac{\partial p}{\partial t} + \frac{\gamma_a}{g\rho} v \cdot VP + \frac{Q}{c_p} \qquad (10)$$

where γ and γ_a are the actual and dry adiabatic lapse rates respectively.

Then :

$$\frac{\partial^2 p}{\partial t\, \partial z} + \frac{\partial p}{\partial t}\left(\frac{g}{RT} - \frac{\gamma_a p}{R\rho T^2}\right) - \frac{gp}{RT^2}\left(-v \cdot VT + w\gamma' + \frac{\gamma_a}{g\rho}\ v \cdot \nabla p + \frac{Q}{c_p}\right) - g \operatorname{div} \rho v = 0 \qquad (11)$$

where, $\gamma' = (\gamma - \gamma_a)$

Eqn. (11) can be represented in simple form as :

$$\partial X/\partial Z + AX + B = 0 \qquad (12)$$

where, $X = \partial P/\partial t$

$$A = \frac{g}{RT} - \frac{\gamma_a p}{\rho R T^2}$$

$$B = -\frac{g}{RT^2}\left[-v \cdot VT + w\gamma' + \frac{\gamma_a}{g\rho}\ v \cdot VP + \frac{Q}{c_p}\right] - g \operatorname{div} \rho V$$

Solution of this equation is

$$X = e^{-\int A dZ}\left[\int B\, e^{\int A dZ}\ dz + c\right] \qquad (13)$$

where, c is the integration constant,

$$B = B_0 - g\nabla \cdot V\rho$$

and

$$B_0 = -\frac{gp}{RT^2}\left[-v \cdot VT + w\gamma' + \frac{\gamma_a}{\rho g} v \cdot VP \quad \frac{Q}{c_p}\right]$$

The conditions for (13) to be the same as (3) are

$$B_0 = 0$$

$$\text{and } \rho = \frac{\gamma_a p}{g\ T}$$

3. Discussion

Accounting only for the diabatic term, the term B can be expressed

$$B = -g\left[Q/c_p T + V \cdot \nabla \rho\right]$$

using the fact that 95% of the atmospheric mass is in the layer 20-25 km above the surface, then Eqn. (1) can be integrated as :

$$\frac{\partial p_s}{\partial t} = \int_0^\infty g\ \frac{\partial \rho}{\partial t}\ dz = \int_{P_0}^{50} \left(\frac{1}{p}\frac{\partial p}{\partial t} - \frac{1}{T}\frac{\partial T}{\partial t}\right) dp \qquad (14)$$

The pressure levels used in computation are at standard levels.

Computations were carried out using 32 cases of radiosonde data at Buffalo N.Y. to examine the change in surface pressure after 24 hours. The variation of surface pressure was found to be close to the algebraic sum of change due to individual layers. This simply can be explained from Eqn. (14) when

$$\int_{P_0}^{50} \frac{1}{T}\frac{\partial T}{\partial t}\ dp \text{ is very negligible.}$$

In Table 1(a) the average conditions of change of surface pressure and individual layers are tabulated. Generally, change in surface pressure is around 1.2-2.6 mb/24 hours. The effect of diabatic heating seems to contribute around 1.3 mb/24 hr at the surface.

Eqn. (13) can be more generally used in the diagnosis of surface pressure variation. Also the inclusion of the heating effect may be of special importance in studying circulation problems at low latitudes, where the heating effect has some influence on low level circulation.

An individual case for 27 March 1980 is shown in Table 1(b) and shows different results than the mean cases listed in Table 1(a). In this case, the greatest effect for every 100 mb is due to the layer (300/200) mb where the change is surface pressure was calculated from the change of pressure at individual levels using Eqn. (14). In this particular case the term $(1/T)\ \partial T/\partial t$ shows a significant effect.

The two terms $\partial p/\partial t$ and $\partial T/\partial t$ were simply calculated from Eqns. (13) and (10) respectively.

In conclusion, Eqn. (13) can be used for diagnostic pressure tendency in more complete form. The change in pressure at layer (300/200) is the most significant in the air column. Finally, the inclusion of nonadiabatic effects may explain some dynamical processes in the pressure system movements.

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