Dynamical Estimation of Atmospheric Diabatic Heating Over the Northern Hemisphere in Winter

By Kooiti Masuda

Geophysical Institute, University of Tokyo, Yayoi, Bunkyo-ku, Tokyo 113, Japan
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Abstract

Atmospheric diabatic heating rate in the lower troposphere in January of nine years is estimated as the residual in the thermodynamic equation from the data of observed temperature and geopotential height. The heating rate shows marked year-to-year variation. Particularly, the heat source over the North Atlantic has two preferred locations and exhibits some seesaw feature.

1. Introduction

Atmospheric circulation exhibits some variation whose time scale is the order of a month (so-called “index cycle”, etc.) and also interannual variation. It has been emphasized that thermal forcing is important as well as topographic forcing in the mechanism that controls them. It is one of indispensable tasks to know how the forcings actually vary in order to answer whether the variations of circulation pattern are “forced” or “free”.

Diabatic heating is the forcing term of the equation of first law of thermodynamics used in atmospheric dynamics which may be written as

$$c_p \left( \frac{\partial T}{\partial t} + V_H \cdot \nabla H T + \omega \left( \frac{\partial T}{\partial p} - \frac{RT}{cp} \frac{\partial p}{\partial T} \right) \right) = \dot{Q}.$$  

(1)

Here the notation is usual one in (x, y, p)-coordinate system, and $\dot{Q}$ is defined as the heating rate per unit mass.

We can estimate directly the amount of $\dot{Q}$ by estimating convergence of visible and infra-red radiative fluxes, heat released by condensation of water vapor, and convergence of sensible heat flux by sub-grid scale motions. Usually, sensible heat is estimated from conditions at the earth’s surface such as sea surface temperature; condensation is estimated from precipitation. Then only vertically integrated value of $\dot{Q}$ can be obtained.

On the other hand, we can estimate the amount of $\dot{Q}$ as the residual by estimating all terms in the left-hand side of (1), from data of temperature and wind. Such studies has been done in hemispheric scale by Chu (1957), Asa-kura and Katayama (1964), and Brown (1964). There are also recent studies such as Lau (1979). However, these studies give either the normal (many-year average) field of heating rate, or the heating rate in several selected months (as in Brown’s). Now there are sufficient aerological data to discuss interannual variation and vertical distribution of heating rate.

2. Data

I use the data of geopotential height and temperature, analyzed by U.S. National Meteorological Center (NMC). Data are given twice daily (00 GMT and 12 GMT). The grid covers 90°N to approximately 15°N. Gridpoints are arranged as equally spaced on polar stereographic projection. Grid distance is 381 km at 60°N. Vertically, 10 levels (1000, 850, 700, 500, 400, 300, 250, 200, 150, 100 mb level) are used. The period used in this study is every January from 1971 to 1979.

For topography, I use the height averaged in each 5° longitude x 5° latitude mesh arranged by Berkofsky and Bertoni (1955), and interpolate it into NMC gridpoints.

3. Method

I compute the left-hand side of (1) directly
replacing differentials by finite differences on the NMC grid. Space interval for differences is about 700 km, since centered differences are used. Time interval is 12 hours. Averages of values at two time steps are used for quantities other than time derivatives. Horizontal wind \( V_H \) is assumed as geostrophic and is computed by finite differences from geopotential height. (In the computation of geostrophic wind and vorticity, Coriolis parameter is given as function of latitude.)

I estimate vertical \( p \)-velocity \( \omega \) in (1) using the quasi-geostrophic vorticity equation:

\[
\frac{\partial \zeta}{\partial t} + V_H \cdot \nabla \zeta + \beta v - f \frac{\partial \omega}{\partial p} = 0. \tag{2}
\]

Notation is usual one in \((x, y, p)\)-coordinate system. Each term of (2) indicates local change of vorticity, advection of relative vorticity, conversion from planetary vorticity to relative vorticity due to movement and stretching of vortex tube, respectively. We can obtain \( \omega \) by vertically integrating the last term of (2), which is given as residual when the other three terms are estimated (cf. Gambo, 1982). The computation is done in the finite difference form of

\[
\omega = \omega_T + \omega_E - \frac{1}{f_0} \int_P \left( \frac{\partial \zeta}{\partial t} + V_H \cdot \nabla \zeta + \beta v \right) dp \tag{3}
\]

where \( f_0 \) is Coriolis parameter at 45°N. Surface pressure \( P_s \) is obtained by vertically interpolating standard level pressure assuming that height is linear to \( \log(p) \) in each interval.

For the boundary condition of \( \omega \) at the earth's surface, we have to consider two effects. One is the topographic effect, such that air is forced to blow upward when it is encountered by mountains. In this study, it is expressed as \( \omega_T = k V_H P_h \), where \( V_H \) is the geostrophic wind on the standard level near the surface at that location, and \( k \) is an empirical constant (0.5) indicating that wind in the planetary boundary layer is weaker than that of free atmosphere. The value is comparable to the ratio of the observed wind speed \( V \) to the geostrophic wind speed \( V_g \) over the Tibetan Plateau reported by Murakami (1981): \( V/V_g = 0.40 \) for samples at 0.0-0.5 km from the surface, 0.74 for 0.5-1.0 km.

The other effect is the Ekman pumping, such that air moves inside the surface lows and converges, and is forced to move upward. It can be formulated as follows:

\[
\omega_E = -\frac{\sin 2\theta}{2} \rho g \sqrt{\frac{2K}{f}} \zeta \tag{4}
\]

where \( \rho \) is density of air, \( g \) is acceleration of gravity, \( K \) is eddy viscosity, \( f \) is Coriolis parameter, and \( \theta \) is the angle between geostrophic wind and surface wind. Since \( K \) and \( \theta \) are uncertain, I assume all these parameters as constants: \( \rho = 1.29 \text{ kg m}^{-3}, \ K = 10 \text{ m}^2/\text{s}, \ \theta = 22.5^\circ, \ f = 10^{-4} \text{ s}^{-1} \); thus \( \omega_E \) is set proportional to \( \zeta \), geostrophic vorticity at the standard level near the surface with the coefficient of proportionality 20 mb.

To assure that \( \omega \) is zero at the top of the atmosphere, I adjust \( \omega \) by assuming that the error in divergence is linear in pressure difference from the surface, similarly to Eq.(19) in O'Brien (1970). It is formulated here as

\[
\omega = \omega^* - \omega^*_{\text{top}} (p - p_t)^2/(P_{\text{top}} - p_t)^2 \tag{5}
\]

where \( \omega^* \) is the value of \( \omega \) computed from (3), and 100 mb level is assumed as "top".

In the course of computation by finite differences, small scale variations appear. They seem to be due to the error in advective terms in (3) and (1). To extract large-scale phenomena where quasi-geostrophic approximation is valid, I filter the results of (3) and (1) with a kind of running mean operator with weighting shown in Fig. 1. It extinguishes spatial variations whose wavelength is 2d or 4d, where \( d \) is the grid interval. Thus most of variations whose scale is less than 1500 km are eliminated.

The domain of computation includes the entire

![Fig. 1 Weights used for filtering.](image-url)
NMC grid except the points at the edge of the grid. The levels substantially lower than earth's surface at each point are excluded.

4. Result

In Fig. 2, the value of $\omega$ at 700 mb computed by (3), averaged for 9 Januarys from 1971 to 1979, is shown. We see upward motion as large as $2 \times 10^{-4}$ mb/s, i.e. 2 mm/s, at the locations of Icelandic and Aleutian lows. Downward motion exists over continental area, as China and Canada.

In Fig. 3, $\dot{Q}$ integrated vertically from the surface to 500 mb is shown. The amount of heating at the peak over Japan is 230 W/m² (450 langleys/day), corresponding to warming of 3.5 K per day. Heating rate at the peak over North Atlantic is about 2 K/day. These heat sources lie where cold air flowing out of continents gets heat from warm sea below (Kuroshio and Gulf Stream). Heat sinks exist over continents such as Siberia and Canada, where air gives heat to the surface which is cooled by radiation.

The result is in good agreement with heating rate of atmospheric columns for normal January, which is estimated from sensible heat, condensation and radiation by Kubota (1970). However, heat sinks over continents are much stronger than in his result. My result also agrees with Lau's dynamical estimation (1979).

There is another heat source over western North America. Though I have not examined its structure well, it seems to be due to transient disturbances since it does not appear in the result of the steady state approximation mentioned below.

In Fig. 5, vertical distribution of $\dot{Q}$ at several regions is shown. The value is averaged for gridpoints included in each 15° longitude × 15°
latitude mesh. The letters on lines in Fig. 5 denote the area as shown in Fig. 4. Roughly speaking, (i) is west of Japan, (j) is east of Japan, (t) is eastern coast of the United States, (v') is south of Greenland. As may be seen in Fig. 5, all these heat sources decrease in intensity upwards up to 500 mb level.

The result in the upper troposphere (above 500 mb) seems to be unrealistic. It indicates strong heating near Japan and strong cooling over central North Pacific (see (m) in Fig. 5). This seems to be due to insufficient capability of the finite difference scheme over regions of strong jet stream. The ratio of spatial interval to time interval \( \Delta s / \Delta t \) is about 17 m/s so that this scheme cannot correctly treat balances of vorticity and heat where characteristic wind speed exceeds this value.

For comparison, \( \dot{Q} \) is estimated from temperature and height averaged through the period of analysis, assuming steady state (Experiment S). The result (not shown) is roughly similar to Fig. 3, so that S is reasonable for January in the sense of first approximation. However, two peaks over the North Atlantic in Fig. 3 cannot be resolved in S. In the difference between the result in Fig. 3 and S, which indicates the effect of transient disturbances, heating near 30\(^\circ\)N and cooling over high latitude continents are found.

Fig. 6 shows the variation of heating rate in time series for regions (t) and (v') where year-to-year variation is the most marked. Dots indicate 5-day averages and short horizontal bars indicate monthly averages. Heating rate is negatively correlated to the temperature at 850 mb level as shown in Fig. 7. The relation is almost linear, so that a straight line is manually fitted to the data in each region. The relation may be regarded as reasonable in the sense of bulk method, since the effect of variation of sea surface temperature and wind speed is less than that of variation of air temperature in these regions.

Heating rate of the two regions are negatively correlated to each other as may be seen from Fig. 8. Monthly averages indicated by squares in the figure show good correlation. But 5-day averages show less distinct one. These results suggest that the representative time scale of this pattern is longer than 5 days and is the order of a month.

Significant simultaneous correlations between temporal fluctuations in meteorological parameters at widely separated regions is referred to as "teleconnection". The relation shown in Fig. 8 may be regarded as one example of teleconnec-

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**Fig. 5** Vertical profile of \( \dot{Q} \) for several regions. (The value shown at 1,000 mb is rough estimation using surface temperature for 7 Januarys from 1973 to 1979. It is not used in the vertical integration.)

**Fig. 6** \( \dot{Q} \) integrated from 500 mb to surface in time sequence in regions (t) and (v').
- : 5-day averages.
- : monthly averages.
N : average through the 9 Januarys.

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\[ \frac{1}{\rho} \int \dot{Q} \, dp \]

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\[ W \, m^{-2} \]
Fig. 7 Relation between \( \dot{Q} \) integrated from 500 mb to surface (ordinate) and temperature at 850 mb (abscissa) in regions (t) and (\( v' \)).

- : 5-day averages.
\( \Box \) : monthly averages.
N : average through the 9 Januaries.

Fig. 8 Relation between \( \dot{Q} \) integrated from 500 mb to surface in region (\( v' \)) (ordinate) and that in region (t) (abscissa).

- : 5-day averages.
\( \Box \) : monthly averages.
N : average through the 9 Januaries.

5. Discussion

The year-to-year variation of heating rate is closely correlated to that of hemispheric circulation pattern. When zonal wave-number 2 (3) dominates the circulation as in January 1972, 1974 (1977, 1976), the center of North Atlantic heat source is located to the south of Greenland (over the east coast of the United States) to be distant approximately 180° (130°) longitude from the heat source over Japan.

It can be explained in the following way: the circulation pattern in the atmosphere controls the strength of continental air masses and the location where the air flows out (particularly on the east coast of North America). Thus, diabatic heating cannot be regarded as pure external forcing to the atmospheric circulation.

Then it is the question whether heating rate is essentially passively determined by the circulation or it forms some feedback system with the circulation to maintain them together. I wish to contribute to solving the problem by examining time series of heating rate and other variables. The period of analysis should be extended to whole winter.

However, there are technical problems to be solved before adapting the method to long period. It is necessary to improve the computational scheme for estimating advective terms in vor-
ticity and thermodynamic equations to obtain information about $Q$ in the upper troposphere. Moreover, this study does not include the heat sources over the tropics nor the Southern Hemisphere mainly due to limitation of data. To extend the study to tropics, the data of observed wind are necessary, because the quasi-geostrophic approximation is not adequate.

Since I am checking the method using the FGGE 3B data, the accuracy of $\omega$ and heating rate estimated in this note will be discussed in a coming paper.

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Computation is performed on M-200H at the Computer Centre, University of Tokyo. Contour maps are drawn using graphic programs of NCAR software support library.

References


冬の北半球大気中の非断熱加熱率の力学的見積もり

増 田 耕一
東京大学理学部地球物理学教室

対流圈下部の非断熱加熱率を，1971年から79年までの9年間の1月について，熱力学の式の残差として，等圧面高度と気温の実測値から見積もった。加熱率には明らかな年々の変動がある。ことに，北大西洋の熱源は出やすいところが2か所あり，シーソーのようなふるまいを示す。