Estimates of the Generation of Available Potential Energy by Infrared Radiation

ANTHONY R. HANSEN[1] AND RICHARD L. NAGLE

Department of Geology and Geophysics, Yale University, New Haven, CT 06511

(Manuscript received 23 December 1983, in final form 20 April 1984)

ABSTRACT

Data from the National Meteorological Center and net outgoing infrared radiation (IR) data measured by NOAA satellites for January 1977 are used to compute estimates of the spectral and spatial contributions to the net generation of available potential energy in the Northern Hemisphere due to infrared radiation. Although these estimates are necessarily crude, the results obtained indicate that IR causes destruction of both zonal and eddy available potential energy. The contributions from midlatitudes to the zonal and eddy generation are $G_Z \approx -5.0$ W m^{-2} and $G_E \approx -0.6$ W m^{-2}, respectively. The eddy generation is due almost entirely to stationary wavenumbers 1 and 2. Comparison with earlier studies and computation of Newtonian cooling coefficients are discussed.

1. Introduction and background

Diabatic heating in the atmosphere is composed of four principal contributions. These are: 1) absorption of solar radiation; 2) emission and absorption of longwave (infrared) radiation; 3) heating due to the release of latent heat of condensation; and 4) transfer of sensible heat from the surface of the earth to the atmosphere. The characteristics of atmospheric diabatic heating are of crucial importance because it is the nonuniform distribution of diabatic heating that is the driving force behind the general circulation. A measure of this effect in the atmospheric energy cycle is given by the generation of available potential energy (APE), which is determined by the spatial correlation of the diabatic heating with the atmospheric temperature field (Lorenz, 1955). The rate of APE generation is one measure of the intensity of the general circulation (Wiin-Nielsen, 1968).

Diabatic heating and APE generation are difficult quantities to measure and are conventionally estimated as budget residuals. Two approaches have been used traditionally to compute the diabatic heating. First, the heating is computed as a residual in a heat budget (Wiin-Nielsen and Brown, 1962; Brown, 1964; Geller and Avery, 1978) or potential vorticity budget (Lawniczak, 1969). In principle, the residual methods can give the spatial and temporal distributions of the heating field. However, they are limited by data quality and the need to compute vertical velocities in the heat budget calculation. The second approach is to use detailed radiation models to compute the time mean and (often) zonal mean radiational heating fields (Katayama, 1967; Newell *et al.*, 1974; Freeman and Liou, 1979). Latent heat release and surface sensible heat flux terms must be estimated independently. Unfortunately, this approach must necessarily deal with the time mean heating because the radiation models require assumptions about the mean temperature, cloudiness (amount and level) and atmospheric composition (water vapor, aerosols, etc.). Therefore, these studies by their very nature cannot deal with the transient part of the diabatic heating and APE generation. Newell *et al.* (1974) were further constrained to the time averaged, zonal mean APE generation, although Katayama (1967) made estimates of the standing eddy APE generation.

In this report, we use an alternative approach to estimate the contribution of infrared cooling to the APE generation. We use NOAA satellite measurements of outgoing infrared radiation (Gruber and Winston, 1978) combined with NMC height field data to determine temperature, both for January 1977, to estimate the spectral and geographic contributions to this component of the generation. Suomi and Shen (1963) were the first to use this approach to estimate IR-induced APE generation on selected dates and for selected areas during 1959 and 1960. They found the generation due to IR to be comparable in magnitude with other estimates of the total APE generation and to exhibit significant temporal variability.

2. Data and approach

The total rate of APE generation G can be expressed as the sum of various spatial and temporal contributions:

$$G = G_Z + G_E = G_{ZS} + G_{ZT} + G_{ES} + G_{ET},$$

[1] Present affiliation: Meteorology Research Center, Control Data Corporation, Minneapolis, MN 55440.

where subscripts Z refer to zonally averaged quantities

$$(\ \ )_Z = \frac{1}{2\pi}\int_0^{2\pi} (\ \ )d\lambda$$

and E to eddy quantities $(\ \ )_E = (\ \ ) - (\ \ )_Z$, and the subscripts S and T refer to the stationary (time mean) mode

$$(\ \ )_S = \frac{1}{t_2 - t_1}\int_{t_1}^{t_2} (\ \ )dt$$

and transient mode $(\ \ )_T = (\ \ ) - (\ \ )_S$, respectively. Using Lorenz' (1955) approximate form, these contributions can be written as

$$G_{ZS} = [\gamma q'_{ZS}T'_{ZS}], \tag{1}$$

$$G_{ZT} = [\gamma(q'_{ZT}T'_{ZT})_S], \tag{2}$$

$$G_{ES} = [\gamma(q_{ES}T_{ES})_Z], \tag{3}$$

$$G_{ET} = [\gamma(q_{ET}T_{ET})_{ZS}], \tag{4}$$

where the brackets denote an average over the mass of the atmosphere normalized per unit area, q is the diabatic heating rate per unit mass, T the temperature, $\gamma = R^2/(p^2c_p\bar{\sigma})$,
and

$$\bar{\sigma} = -\frac{\alpha}{\theta}\frac{\partial\theta}{\partial p}$$

is a measure of the static stability. A value of $\bar{\sigma}$ for the 850–300 mb layer was adapted from Tomatsu (1979). Departures of the zonal mean heating and temperature from their hemispheric mean values are denoted by q'_Z and T'_Z, respectively. (We will ignore departures of the hemispheric means from the global means.) Dutton and Johnson (1967) and Newell *et al.* (1974) discuss the validity of these approximate forms of G_{ZS} and G_{ZT}. In addition, the eddy components of the generation can be broken down into zonal harmonic spectra

$$G_E = \sum_m G_m = \sum_m [\gamma(q_mT_m^* + q_m^*T_m)],$$

where the subscript m denotes the contribution for wavenumber m, and an asterisk denotes a complex conjugate of the coefficients q_m or T_m. Thus, relative cooling of relatively warm air or relative heating of relatively cold air will result in destruction of APE and vice versa.

The infrared radiation data that we use are those measured by the scanning radiometers aboard NOAA operational polar-orbiting satellites. Descriptions of the instruments and data reduction procedures are given by Gruber and Winston (1978). The satellite's equator-crossings occur at 0900 and 2100 LST daily. We average the twice-daily observations to get daily average values at each grid point of a 2.5° latitude–longitude array. The NOAA satellite observes IR in the atmospheric "window region" (10.5–12.5 μm wavelengths). In the analysis procedure used by NOAA, estimates are made of the total outgoing IR flux from these window region radiance measurements by using a nonlinear regression model derived from radiation calculations for 99 different atmospheres covering a wide variety of temperature and moisture conditions as well as clear and cloudy skies (Gruber and Winston, 1978).

Let us briefly discuss the sources of the net outgoing IR. The surface of the earth or of a cloud radiates essentially as a blackbody. A portion of this radiation is transmitted by the atmosphere to space and the remainder is absorbed by atmospheric gases, primarily water vapor, carbon dioxide and ozone, or by clouds. These constituents then reradiate IR at an intensity appropriate to their temperature and emissivity. For a long-term global mean, most of the IR leaving the top of the atmosphere comes from the atmosphere itself. However, the distribution of the total outgoing IR is particularly dependent upon the distribution of cloudiness and water vapor in the atmosphere (Weinstein and Suomi, 1961; Paltridge and Platt, 1976; Freeman and Liou, 1979; Ohring and Gruber, 1983).

Our interest is in estimating the tropospheric IR cooling rate from the satellite measurements. Earlier studies (e.g., Kuhn and Suomi, 1960; Sabatini and Suomi, 1962) indicate that satellite measurements of outgoing IR give a good estimate of the net IR flux divergence from the troposphere compared to direct radiometersonde measurements. However, some unknown contribution to the satellite-derived IR flux will be provided by emission from the earth's surface. For a long-term, global mean this fraction will be roughly 20% of the total (Paltridge and Platt, 1976). In cloud-free conditions it will be closer to 50% (or greater in regions of low water vapor content, i.e., near the pole), while in regions of overcast it will be near zero. In winter, average daytime cloudiness typically exceeds 75% over midlatitude oceans and 50% over midlatitude continents (except in desert regions) (Clapp, 1964), suggesting a generally smaller contribution from the surface to the satellite IR observations. However, the vertical distribution of the cloudiness is crucial in determining its effect on the tropospheric cooling rate for a given temperature and moisture profile (Cox, 1969). Unfortunately, the available satellite observations cannot determine the vertical structure of the cooling rate. As an approximation, we shall assume that a linear relationship exists between measured outgoing IR and the resultant tropospheric cooling rate due to this radiation.

Following Suomi and Shen (1963), we will assume that the satellite-observed IR is representative of the vertical mean value for the troposphere (1000–100 mb layer) and will ignore correlations between q and T

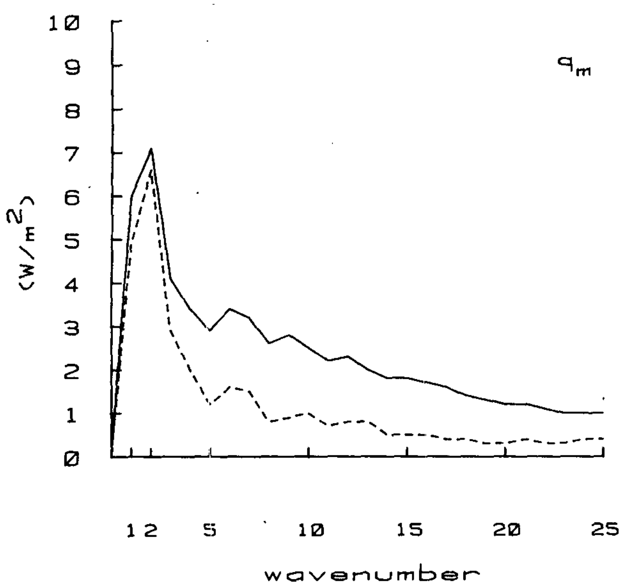

FIG. 1. The amplitude of $[|q_m|]_s$ for the total IR (solid line) and time-mean contribution ($[|q_{ms}|]$, dotted line) for January 1977 as the function of the zonal wavenumber m. In this case, q_m is averaged over the zone from 30 to 80°N.

due to variations about their vertical average. This is a crude but acceptable first approximation. Freeman and Liou's (1979) model, for example, indicates that variations in q_{zs} about its vertical mean are not excessively large for January conditions, particularly in extratropical latitudes. By making this assumption, we will overestimate the tropospheric cooling rate by ignoring the contribution to the satellite-observed IR flux given by radiation from the earth's surface transmitted by the atmosphere and by radiation from layers above the troposphere. The former contribution could account for up to 50% of the total observed irradiance in certain locations and the latter effect is probably small.

Tropospheric mean temperatures are computed hydrostatically from the NMC operational analyses of the 1000 and 100 mb geopotential heights. These data are available twice daily at 0000 and 1200 GMT and were averaged to give daily mean temperatures. Certain inaccuracies may be introduced because the satellite data and temperature data are not synoptic, leading to misinterpretation of the transient part of G. However, for slowly evolving, large spatial scales, this problem should not be serious.

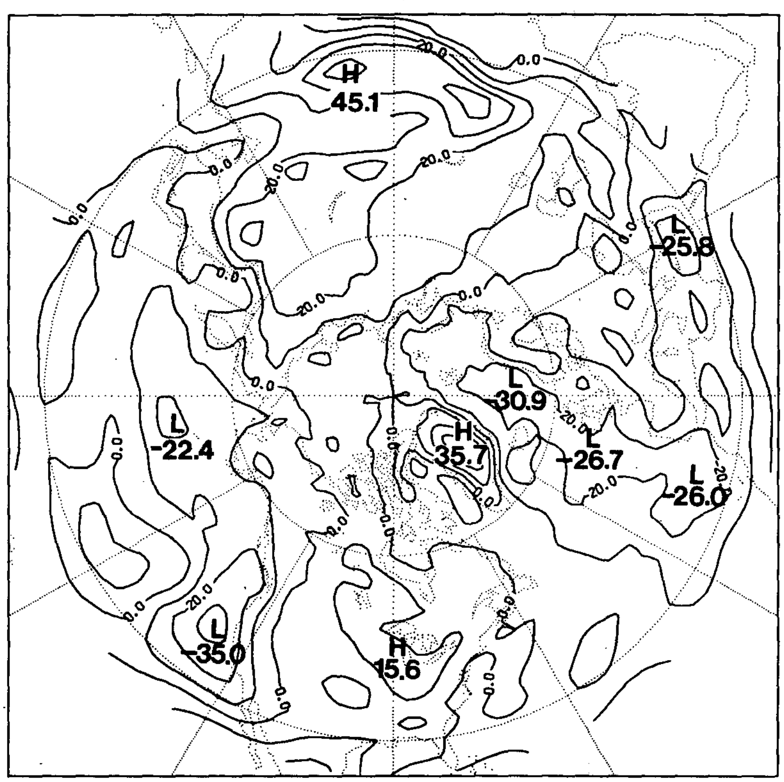

FIG. 2. The geographic distribution of q_{ES} in the zone from 25 to 80°N. The field plotted is the IR *heating rate* (W m^{-2}), so negative values indicate strong cooling and positive values indicate relative warming. Contour interval is 10 W m^{-2}. Contributions to q_{ES} from wavenumbers higher than 25 have been removed.

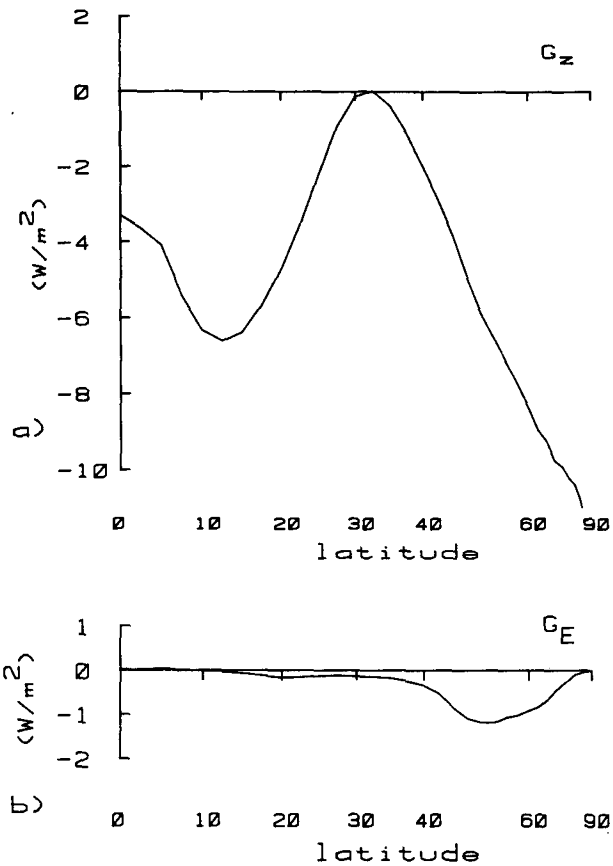

FIG. 3. The integrands of G_Z (a) and G_E (b) as functions of latitude.

Although our admittedly crude assumptions may introduce certain systematic errors into our calculation, if at least the phase of the net cooling rate is correct, we can proceed with our estimate of the APE generation. However, we must keep in mind that our estimates of G will at best achieve the correct sign and the correct order of magnitude.

TABLE 1. Average values during January 1977 for the contributions to the IR-produced generation of available potential energy (W m^{-2}) in the zones noted. For comparison, Katayama's (1967) values from a radiation model calculation are included.

Zone	G_Z	G_{ZS}	G_E	G_{ES}
30–80°N	−4.98	−4.96	−0.61	−0.61
0–90°N	−4.74	−4.73	−0.33	−0.31
Katayama (1967) IR (0–90°N)	—	−3.443	—	−0.242
Katayama (1967) SR* (0–90°N)	—	2.914	—	−0.005

* Solar radiation.

3. Results

We will present results for January 1977 only. Calculations for January 1978 showed qualitatively similar features. First, consider the amplitude spectrum of the eddy infrared cooling rate $[|q_m|]_s$ (Fig. 1). Notice that maximum values are obtained at planetary-scale wavenumbers, suggesting the important sensitivity of q_m to the large-scale temperature field. Examination of the geographic distribution of q_{ES} (Fig. 2) shows that certain geographic features leave a strong signature in the radiation field. Features such as the Tibetan

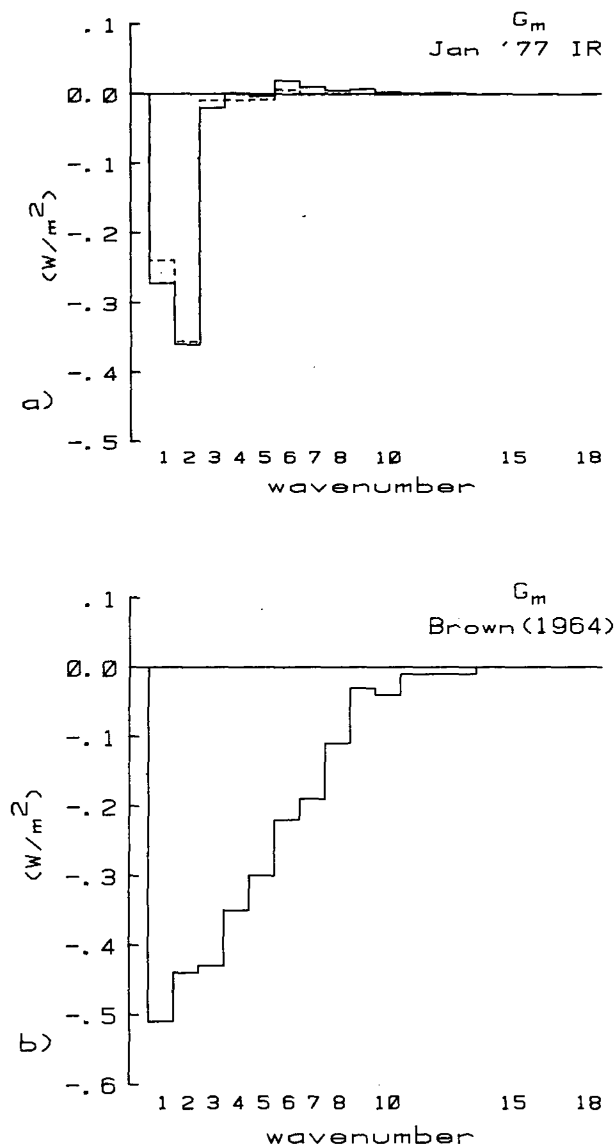

FIG. 4. The spectrum of G_E for (a) our results for the IR contribution in January 1977 where the solid line is the total G_m and the dashed line is G_{ms}, and (b) Brown's (1964) estimates of the total G_m averaged for three Januaries (1959, 1962, 1963.) Our values are averaged for 30–80°N and Brown's values are for 20–87.5°N.

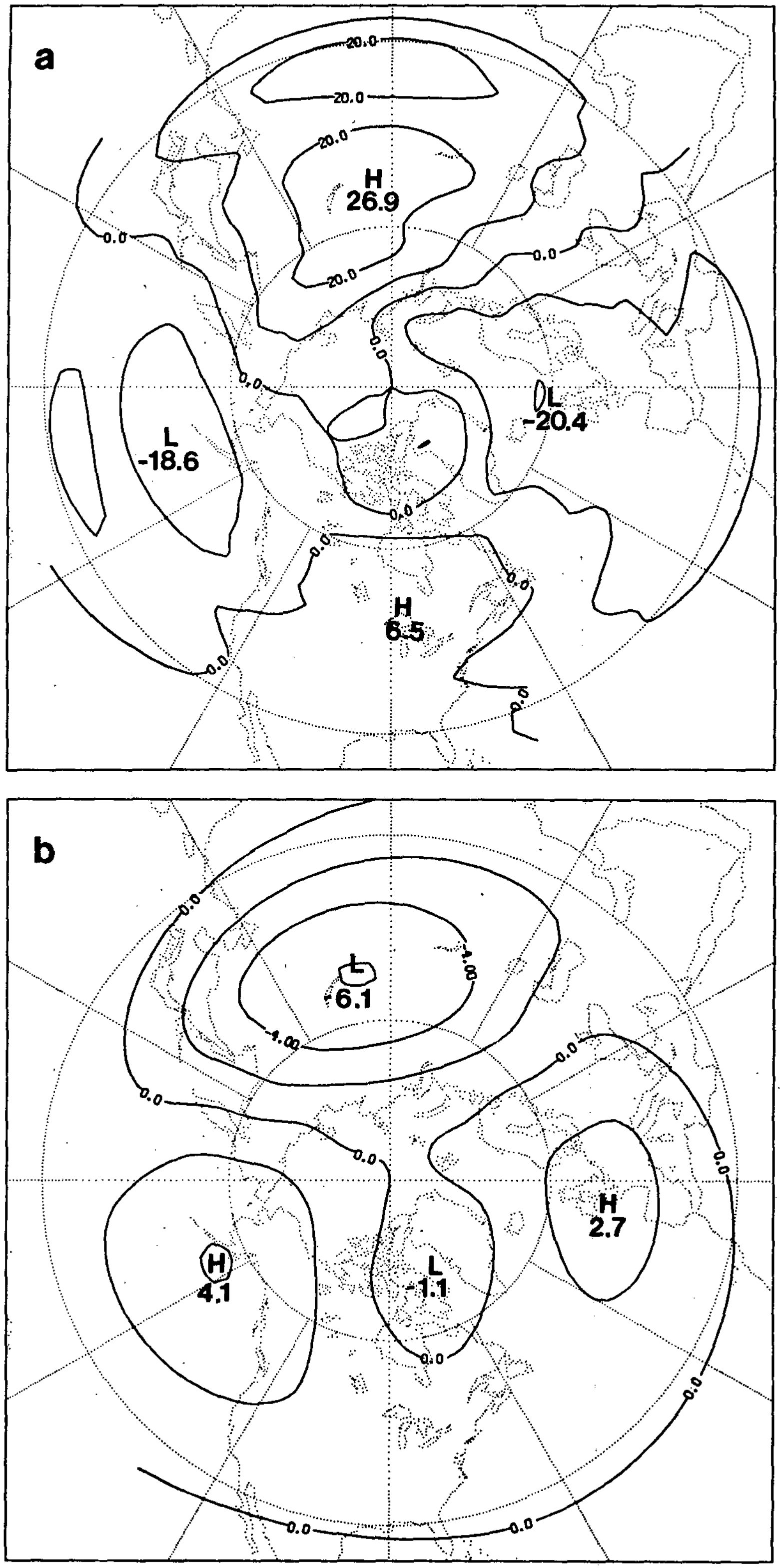

FIG. 5. The geographic distribution of the ultralong-wave fields (zonal wavenumbers 1–2) of (a) $\sum_{m=1}^{2} q_{ms}$ (W m^{-2}), (b) $\sum_{m=1}^{2} T_{ms}$ (°C) and (c) the integrand of $\sum_{m=1}^{2} G_{ms}$ (W m^{-2}) for January 1977. Contour intervals are 10 W m^{-2}, 2°C and 1 W m^{-2}, respectively.

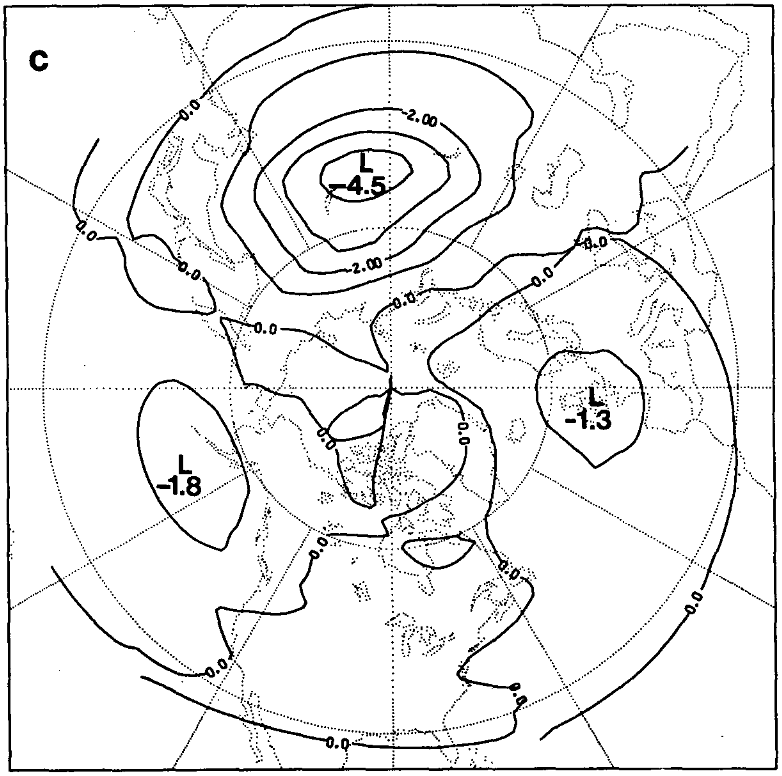

FIG. 5. (*Continued*)

plateau and Greenland are clearly evident. Most of these features are represented by wavenumbers higher than 3, however. At the planetary scale, the IR distribution mirrors the temperature field, as we will show shortly. The monthly mean field of q_{ES} for January 1977 does show certain large departures from the mean for the eight Januaries from 1975 through 1982 (not shown). In particular, the strong cooling off the California coast in January 1977 has an amplitude nearly twice the eight-January mean value.

In Fig. 3 the integrands of G_Z and G_E are presented as functions of latitude. The monthly mean values are contributed almost entirely by the stationary mode (See Table 1). The relative maximum in the G_Z integrand (Fig. 3a) near the equator is due to persistent high cloudiness near the ITCZ and the resultant reduction in radiation from the troposphere (during summer, preliminary estimates indicate that this effect can result in a positive contribution to G_Z in the tropics due to IR). A minimum in the integrand of G_Z occurs at ~12.5°N due to relatively cloud-free conditions and large IR cooling rates in this zone (Gruber and Winston, 1978). From a zero value near 30°N, the G_Z integrand decreases monotonically toward polar regions. Contributions to G_E (Fig. 3b) are only important between 35 and 75°N, reaching values of approximately -1 W m^{-2} between 45 and 60°N. Katayama (1967) estimated G_{ZS} and G_{ES} for the Northern Hemisphere based upon mean heating rates from a detailed radiation model. His results are quite similar to ours (Table 1), suggesting that the approximations we have made are acceptable. In addition, Katayama's results for G_{ES} due to absorption of solar radiation indicate that this contribution is extremely small (<0.01 W m^{-2}). Thus, the infrared contribution to G_{ES} may be equivalent to the total radiational contribution.

The spectrum of G_E (Fig. 4a) reveals that the total G_E is contributed almost entirely by the standing component of wavenumbers 1 and 2. This contribution by the ultralong waves is competitive with the total G_m at wavenumbers 1 and 2 computed by Brown (1964; January values reproduced in Fig. 4b). The small positive values of G_m at intermediate wavenumbers may be the result of a cumulative effect due to positive APE generation in the developing stage of cyclones (Vincent *et al.*, 1977), but it is too small to be of importance to the mean energetics. We should add, of course, that the spectrum of G_E is not linearly related to the spectrum of the APE (not shown).

The spectral energetics for the winter of 1976–77 have been presented by Chen (1982). Comparison with his calculations of the zonal to eddy APE conversion $C(A_Z, A_E)$, the eddy APE to eddy kinetic energy conversion $C(A_E, K_E)$, and wave–wave interactions for the ultralong waves shows that the IR-induced APE destruction is a significant term in the APE budget of the standing ultralong waves, accounting for 70–80% of the required balancing of the APE budget residual at these length scales.

Because the standing ultralong waves make the major contribution to the IR-induced G_E, it is of interest to examine the geographic distribution of the ultralong-wave IR heating and temperature fields, as well as their correlation (Fig. 5). Notice the tendency for the IR heating and temperature fields to be out of phase, leading to large negative contributions to G_{ES} over interior Asia with lesser contributions over the northeast Pacific and over western Europe and the eastern Atlantic (Fig. 5c).

The destruction of eddy APE in midlatitudes measured by Brown (1964) may be largely due to air mass modification over the Gulf Stream and Kuroshio. This process would presumably take place at primarily intermediate scales. If Brown's and our results are representative and comparable, we could infer that a significant amount of the total destruction of eddy APE at wavenumbers 1 and 2 is due to radiational processes, while radiational processes are energetically unimportant at intermediate scales.

Although the transient mode makes no contribution to the total monthly mean G_Z and G_E, this does not mean that it is zero on any given day. The time series for G_Z and G_E averaged over 30–80°N (Figs. 6a and b) show significant departures from the monthly mean that appear systematic rather than random. The departures of G_Z and G_E from their monthly mean values are nearly always directly out of phase with the temporal variations of A_Z and A_E (not shown), a not altogether surprising result which is consistent with the strong dependence of the cooling rate on temperature.

One source of systematic error in our calculation is probably the lack of vertical resolution in our computations. For instance, we have compared the APE calculated from the 1000–100 mb thickness with that calculated from the NMC analyses of observed temperatures at the ten mandatory levels in the troposphere. Although the APE calculated in the present study is fairly accurate for wavenumbers 1, 2 and 3, it is consistently underestimated by a factor of 3 or more for wavenumbers 4 and higher. A phase reversal of the intermediate-scale temperature waves between the lower and upper troposphere most likely explains this discrepancy. This implies that G_m for these wavenumbers is also underestimated. Although the intermediate scale G_m would then assume larger positive values, it would still be small compared to other APE budget terms.

Finally, we can use our results to compute a Newtonian cooling coefficient α in a manner analogous to that used by Saltzman (1973). This coefficient, which is a time constant for the destruction of eddy temperature variance due to longwave radiation, can be defined as

$$\alpha = \frac{(q_E T_E)_{zs}}{[C_p (T_E^2)_{zs}]}.$$

We get $\alpha \approx 3 \times 10^{-7}\ \mathrm{s}^{-1}$ ($\alpha^{-1} \approx 38$ days).

As already mentioned, Katayama (1967) indicates that the contribution of solar radiation to G_{ES} is negligible and so its contribution to the total G_E is probably small also. Our α may therefore be representative of the total radiational damping coefficient and would apply only to the ultralong waves, since the radiational contribution to G_E for smaller scales is negligible. Wiin-Nielsen *et al.* (1967) obtained a somewhat larger value for α ($4 \times 10^{-7}\ \mathrm{s}^{-1}$) based on Brown's (1964) total diabatic heating. Saltzman (1973) obtained a much larger value, $\alpha = 2 \times 10^{-6}\ \mathrm{s}^{-1}$, appropriate to the total diabatic heating at 850 mb. This seems plausible due to the large influence of boundary-layer effects at that level.

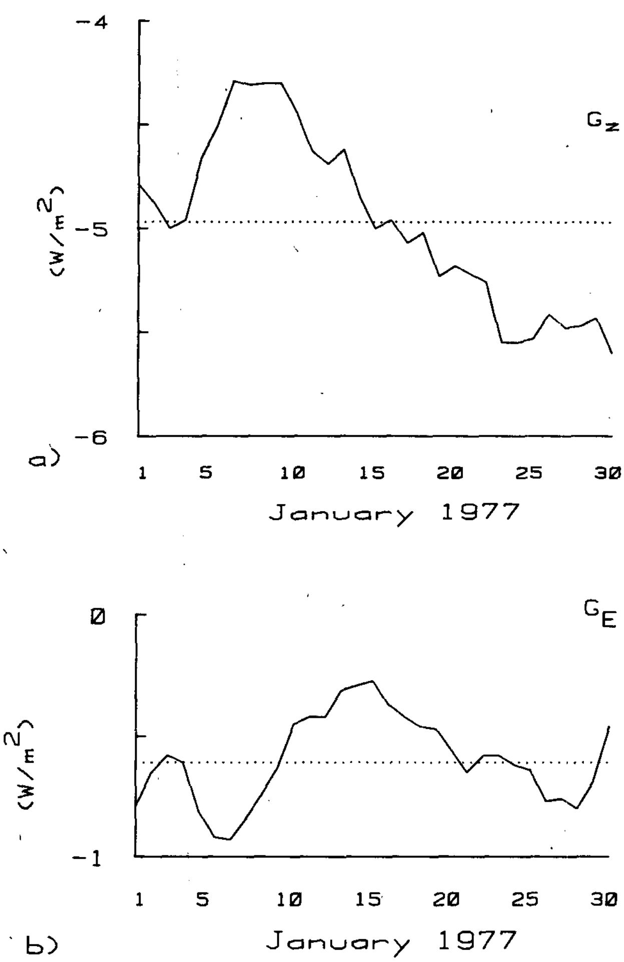

FIG. 6. The time series of the IR-produced G_Z (a) and G_E (b). The values shown are the contributions to G_Z and G_E for the zone 30–80°N. The dotted line indicates the value of G_{ZS} and G_{ES} in (a) and (b), respectively.

4. Conclusions

The infrared cooling contribution to G_Z and G_E in winter based on satellite observations of the infrared cooling rate is negative. Further, the satellite-based estimates compare well with available radiation-model-based estimates despite the assumptions about the source and vertical uniformity of the observed cooling rate. The contribution of G_Z in winter is negative at all Northern Hemisphere latitudes while contributions to G_E are only significant in the 35–70°N zone and are due almost entirely to standing wavenumbers 1 and 2. A Newtonian cooling coefficient of $\sim 3 \times 10^{-7}$ s^{-1} applicable to the ultralong waves was calculated, corresponding to a radiative damping time scale of $\sim$38 days. A similar value was obtained for January in both 1977 and 1978.

Based upon a comparison of our results with those of Wiin-Nielsen and Brown (1962) and Brown (1964), it appears that the contribution to the destruction of eddy APE in midlatitudes during winter by radiational processes is significant for the ultralong waves, while destruction of eddy APE by heat flux from the earth's surface may be predominant at higher wavenumbers ($m \geqslant 3$). The role of latent heat release in midlatitude G_E is still an open question (Hayashi and Golder, 1981).

Acknowledgments. We would like to thank Professor Barry Saltzman and Professor Ronald Smith for useful discussions. This study was supported by the National Aeronautics and Space Administration under Grant NAS 8-34903 at Yale University.

REFERENCES

Brown, J. A., 1964: A diagnostic study of tropospheric diabatic heating and the generation of available potential energy. *Tellus,* **16,** 371–388.

Chen, T.-C., 1982: A further study of spectral energetics in the winter atmosphere. *Mon. Wea. Rev.,* **110,** 947–961.

Clapp, P. F., 1964: Global cloud cover for seasons using Tiros nephanalysis. *Mon. Wea. Rev.,* **92,** 495–507.

Cox, S. K., 1969: Radiation models of midlatitude synoptic features. *Mon. Wea. Rev.,* **97,** 637–651.

Dutton, J. A., and D. R. Johnson, 1967: The theory of available potential energy and a variational approach to atmospheric energetics. *Advances in Geophysics,* Vol. 12, Academic Press, 333–436.

Freeman, K. P., and K.-N. Liou, 1979: Climatic effects of cirrus clouds. *Advances in Geophysics,* Vol. 21, Academic Press, 231–287.

Geller, M. A., and S. K. Avery, 1978: Northern Hemisphere distributions of diabatic heating in the troposphere derived from general circulation data. *Mon. Wea. Rev.,* **106,** 629–636.

Gruber, A., and J. S. Winston, 1978: Earth–atmosphere radiative heating based on NOAA scanning radiometer measurements. *Bull. Amer. Meteor. Soc.,* **59,** 1570–1573.

Hayashi, Y., and I. Golder, 1981: The effects of condensational heating on mid-latitude transient waves in their mature stage: Control experiments with a GFDL general circulation model. *J. Atmos. Sci.,* **38,** 2532–2539.

Katayama, A., 1967: On the radiation budget of the troposphere over the Northern Hemisphere (III)—Zonal cross-sections and energy consideration. *J. Meteor. Soc. Japan,* **45,** 26–38.

Kuhn, P. M., and V. E. Suomi, 1960: Infrared radiometer soundings on a synoptic scale. *J. Geophys. Res.,* **65,** 3669–3677.

Lawniczak, G. E., Jr., 1969: On a multi-layer analysis of atmospheric diabatic processes and the generation of available potential energy. Ph.D. dissertation, University of Michigan, Ann Arbor, 111 pp.

Lorenz, E. N., 1955: Available potential energy and the maintenance of the general circulation. *Tellus,* **7,** 157–167.

Newell, R. E., J. W. Kidson, D. G. Vincent and G. J. Boer, 1974: *The General Circulation of the Tropical Atmosphere and Interactions with Extra-Tropical Latitudes, Vol. 2.* The MIT Press, 371 pp.

Ohring, G., and A. Gruber, 1983: Satellite radiation observations and climate theory. *Advances in Geophysics,* Vol. 25, Academic Press, 237–304.

Paltridge, G. W., and C. M. R. Platt, 1976: *Radiative Processes in Meteorology and Climatology.* Elsevier, 318 pp.

Sabatini, R. R., and V. E. Suomi, 1962: On the possibility of atmospheric infra-red cooling estimates from satellite observations. *J. Atmos. Sci.,* **19,** 349–350.

Saltzman, B., 1973: Parameterization of hemispheric heating and temperature variance fields in the lower troposphere. *Pure Appl. Geophys.,* **105,** 890–899.

Suomi, V. E., and W. C. Shen, 1963: Horizontal variation of infrared cooling and the generation of eddy available potential energy. *J. Atmos. Sci.,* **20,** 62–65.

Tomatsu, K., 1979: Spectral energetics of the troposphere and lower stratosphere. *Advances in Geophysics,* Vol. 21, Academic Press, 289–405.

Vincent, D. G., G. B. Pant and H. J. Edmon, Jr., 1977: Generation of available potential energy of an extratropical cyclone system. *Mon. Wea. Rev.,* **105,** 1252–1265.

Weinstein, M., and V. E. Suomi, 1961: Analysis of satellite infrared radiation measurement on a synoptic scale. *Mon. Wea. Rev.,* **89,** 419–428.

Wiin-Nielsen, A., 1968: On the intensity of the general circulation of the atmosphere. *Rev. Geophys.,* **6,** 559–579.

——, and J. A. Brown, 1962: On diagnostic computations of atmospheric heat sources and sinks and the generation of available potential energy. *Proc. Int. Symp. on Numerical Weather Prediction,* Tokyo, Meteor. Soc. Japan, 593–613.

——, A. Vernekar and C. H. Yang, 1967: On the development of baroclinic waves influenced by friction and heating. *Pure Appl. Geophys.,* **68,** 131–161.