THE RADIATION BUDGET OF THE STRATOSPHERE

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ABSTRACT

The radiation budget of the northern-hemisphere stratosphere, as a function of the mean thermal structure and composition of the stratosphere, is determined for the months of January, April, July, and October. Emission of infrared radiation by carbon dioxide, water vapor, and ozone is calculated by means of a simple numerical method derived from the differential equations of radiative transfer. Absorption of solar radiation by ozone is taken from published results; absorption of solar energy by water vapor is computed with the aid of an empirical formula.

It is found that, in general, radiative equilibrium is not obtained at any latitude. Low latitudes constitute a heat source and high latitudes a heat sink in the stratospheric energy budget. It is shown that carbon dioxide is more important than water vapor in cooling the stratosphere and that infrared transfer in the 9.6μ ozone band normally results in a convergence of energy in the stratosphere.

Some features of the stratospheric temperature distribution and circulation pattern are inferred from the computed radiation budget and its seasonal variations.

1. Introduction

The mean general circulation and meridional temperature distribution in the stratosphere, as in the troposphere, are consequences of the mean latitudinal distribution of radiant energy. In the troposphere, the mean vertical temperature distribution is a result of convective, condensation-evaporation, and radiative processes, whereas in the stratosphere the shape of the vertical temperature profile is primarily determined by radiation. Therefore, a knowledge of the radiation budget and its latitudinal, vertical, and seasonal variations is basic for an adequate understanding of the mean circulation and thermal state of the stratosphere.

Early investigations of the radiative properties of the stratosphere (Humphreys, 1909; Gold, 1909; Gowan, 1947) were based upon the assumption of radiative equilibrium. Radiative equilibrium temperature distributions were derived which maintained a balance between absorption and emission of radiation. Although this approach eventually led to increased knowledge of the vertical temperature structure, it is now realized that the assumption of radiative equilibrium at each point in the stratosphere is unrealistic.

In recent years the absorption of solar energy and/or the emission of infrared radiation have been computed independently for selected stratospheric models. The latitudinal, seasonal, and vertical variations of the absorption of solar energy by ozone have been studied by a number of researchers (Karandikar, 1946; London, 1952; Johnson, 1953; Pressman, 1954, 1955) and are fairly well known at the present time. The other half of the radiation budget—the distribution of infrared emission by carbon-dioxide, water vapor, and ozone—is not known except for the vertical variation, which has been studied only for idealized models (Craig, 1949; Kaplan, 1952; Plass, 1956a, 1956b).

In the present study, the distribution of infrared emission is computed and is compared to the distribution of solar absorption in order to obtain the mean latitudinal and seasonal variations of the stratospheric radiation budget. (The stratosphere is assumed to extend from the tropopause to 55 km.)

2. Methods of calculation

Theory and equations for infrared transfer.—Atmospheric infrared fluxes and flux divergences are usually computed with the aid of radiation charts (for example, Elsasser, 1942) or tabular methods (for example, Bruinenberg, 1946), which are based upon a graphical or numerical integration of the differential equations of radiative transfer. Since these charts and tables have been constructed for the range of absorber concentrations found in the troposphere, they are not applicable for radiation computations in the stratosphere where gas concentrations are extremely low. In this section we shall derive a simple, numerical procedure for calculating the infrared fluxes.

The basic equations for monochromatic radiative transfer can be written as

$$dI_\nu\downarrow = -k_\nu \sec\theta (I_\nu\downarrow - B_\nu)du \tag{1}$$

$$dI_\nu\uparrow = k_\nu \sec\theta (I_\nu\uparrow - B_\nu)du, \tag{2}$$

where I_ν is the energy in a monochromatic radiation beam of frequency ν crossing a unit horizontal area per unit time, k_ν is the absorption coefficient at frequency ν, θ is the angle between the vertical and the direction of the radiation beam, B_ν is the intensity of black-body radiation at frequency ν, and u is the optical path length of the radiating gas—defined as $u = \int_z^\infty \rho_g dz$, where ρ_g is the density of the radiating gas—which is equal to zero at the top of the atmosphere and increases downwards (see fig. 1). Equation (1) is for a downward-directed beam and equation (2) is for an upward-directed beam.

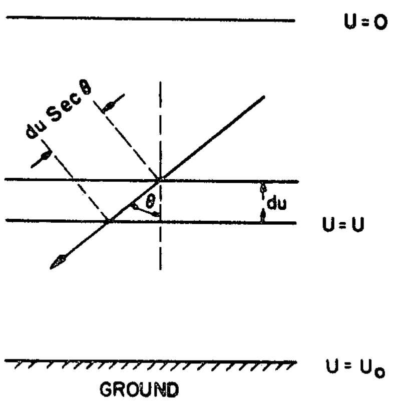

FIG. 1. Geometry of a radiation beam.

In order to obtain the flux of radiation at a particular level in the atmosphere, these equations must be integrated over: (a) path length, (b) frequency of radiation, and (c) the hemispheric solid angle, for each of the absorbants in the atmosphere. The integration of equation (1) from the top of the atmosphere down to a level u yields (see, for instance, Elsasser and King, 1952)

$$I_\nu\downarrow(u) = I_\nu\downarrow(0)\tau_\nu(0, u) + \int_0^u B_\nu(v)\frac{\partial \tau_\nu(v, u)}{\partial v} dv \quad (3)$$

where τ_ν, the monochromatic beam transmissivity, is defined as

$$\tau_\nu(v, u) = \exp\left(-\int_v^u k_\nu \sec\theta du\right) \quad (4)$$

with v and u representing any two levels in the atmosphere. The dummy variable v varies from zero to u in the integral of (3). It can be seen from (4) that v must always be less than or equal to u, since v greater than u would yield a transmissivity larger than unity, which is physically impossible.

To evaluate the integral on the right hand side of equation (3), consider small layers of optical path Δv. Assume that each such layer can be approximated by an isothermal layer whose temperature is the mean temperature of the layer. We then have for this integral the summation $\sum_{i=1}^n \bar{B}_{\nu i}\Delta\tau_{\nu i}$ where n represents the total number of layers considered, and $\bar{B}_{\nu i}$ is the monochromatic black-body intensity at the mean temperature of each layer. Then

$$I_\nu\downarrow(u) = I_\nu\downarrow(0)\tau_\nu(0, u) + \sum_{i=1}^n \bar{B}_{\nu i}\Delta\tau_{\nu i}. \quad (5)$$

To determine the total intensity of beam radiation in a particular band, we must integrate over the width of the band. The beam transmissivity of the band can be obtained from laboratory measurements or from theory, and the black-body energy of the band can be determined from tables. To change from beam radiation to slab radiation (*i.e.*, to integrate over the hemispheric solid angle) we will assume that the transmission of a beam through a column of path length 1.5 u is equivalent to transmission through a slab of path length u. (The choice of the value 1.5, which corresponds to the secant of an angle of 48 deg, is compatible with other simplifications made in the present investigation.) With these assumptions and simplifications, we get for the downward flux

$$F\downarrow(u) = F\downarrow(0)\tau_F(0, u) + \sum_{i=1}^n \bar{B}_i\Delta\tau_{Fi} \quad (6)$$

and similarly for the upward flux

$$F\uparrow(u) = F\uparrow(u_0)\tau_F(u, u_0) + \sum_{j=1}^m \bar{B}_j\Delta\tau_{Fj}, \quad (7)$$

where $F\downarrow(u)$ and $F\uparrow(u)$ are the downward and upward fluxes of diffuse radiation at the level u, $F\downarrow(0)$ is the downward flux at the top of the atmosphere, $F\uparrow(u_0)$ is the upward flux at the ground, τ_F is the slab transmissivity of the band, and $\bar{B}$ is the total black-body energy radiated into the hemisphere at the mean temperature of each layer. The summation index i runs through n layers from the top of atmosphere to the level u, and the summation index j runs through m layers from the ground to the level u.

Equations (6) and (7) state that the radiation flux at a level u is equal to the part of the initial flux transmitted through the entire layer plus the sum of the contributions to the flux from each sub-layer within the entire layer.

The net flux at the level u is given by

$$F_{\text{net}}(u) = F\uparrow(u) - F\downarrow(u). \quad (8)$$

Infrared transmissivities.—The transmissivity of a gas is a function of the optical path length and the pressure. The infrared transmissivities used in the present study are based upon experimental determinations which, in many cases, had to be extrapolated to apply to the extremely low gas concentrations and low pressures found in the stratosphere. The uncertainty in the transmissivities at low gas concentrations is probably the greatest source of error in the infrared flux computations.

The 15μ band, which was assumed to extend from 12.5μ to 17.5μ, is the most important carbon dioxide band for atmospheric radiation computations. There have been a number of measurements and theoretical determinations of the transmissivities of the various parts of the band and of the mean transmissivity of the entire band (Callender, 1941; Kaplan, 1950, 1952; Elsasser and King, 1953; Howard *et al*, 1955). Preliminary computations indicated that a mean transmissivity for the entire 15μ band did not adequately represent the transmission of carbon dioxide at low concentrations and pressures and, in fact, led to over-estimation of the carbon dioxide flux divergence (London *et al*, 1956). The band was therefore broken up into smaller intervals and the empirical formulas of Callender, which represent the transmissivities of intervals within the band, were used in the final computations. These formulas can be written as follows:

for 14–16μ, $$\tau_F = [1 + 0.9(1.5u^*)^{0.84}]^{-1}; \tag{9}$$

for 13–14μ and 16–17μ,
$$\tau_F = [1 + 0.8(1.5u^*)^{0.67}]^{-1}; \tag{10}$$

for 12.5–13μ and 17–17.5μ,
$$\tau_F = [1 + 0.0055(1.5u^*)^{0.67}]^{-1}; \tag{11}$$

where u^* is the pressure-corrected carbon dioxide path length in cm STP (the pressure correction is discussed below), and the factor 1.5 is introduced to correct for slab transmission. It is the center of the band, 14 to 16μ, that is most important for radiative transfer in the stratosphere, since the edges of the band are nearly transparent for the short carbon dioxide path lengths found in the stratosphere.

The major ozone band for atmospheric radiation is the 9.6μ band which extends from 8.85μ to 10.3μ. Laboratory measurements of the transmissivities of this band were made by Summerfield in 1941 and were processed by Elsasser and King (1953) and others. The generalized transmission function which Elsasser and King derived from these measurements was used in the present study.

Recent water-vapor transmissivity measurements at short path lengths and pressures (Daw, 1956) are readily applicable to stratospheric computations and were used in this investigation.

Pressure broadening of absorption lines was taken into account by simply applying a linear pressure correction ($p/1000$, where p is the mean pressure of the layer, and 1000 mb is taken as standard pressure) to the path lengths of carbon dioxide, ozone, and water vapor. Doppler broadening becomes as important as pressure broadening for these gases at about 30 km and was taken into account by keeping the pressure correction constant above 30 km.

Boundary conditions.—The downward flux of infrared radiation at 55 km was neglected because of the small concentrations of absorbing gases above this level. The upward water-vapor radiation flux at the tropopause was taken from values given by London (1957) which were computed for average cloudiness. The upward flux of infrared radiation due to carbon dioxide at the tropopause was assumed to be independent of tropospheric cloudiness and was computed by applying equation (7) to the troposphere. For ozone, the effect of tropospheric cloudiness was considered by computing a weighted upward infrared flux at the tropopause (weighted according to mean total cloudiness at each latitude belt).

The atmosphere was divided into 2-km layers for calculating the infrared fluxes and computations

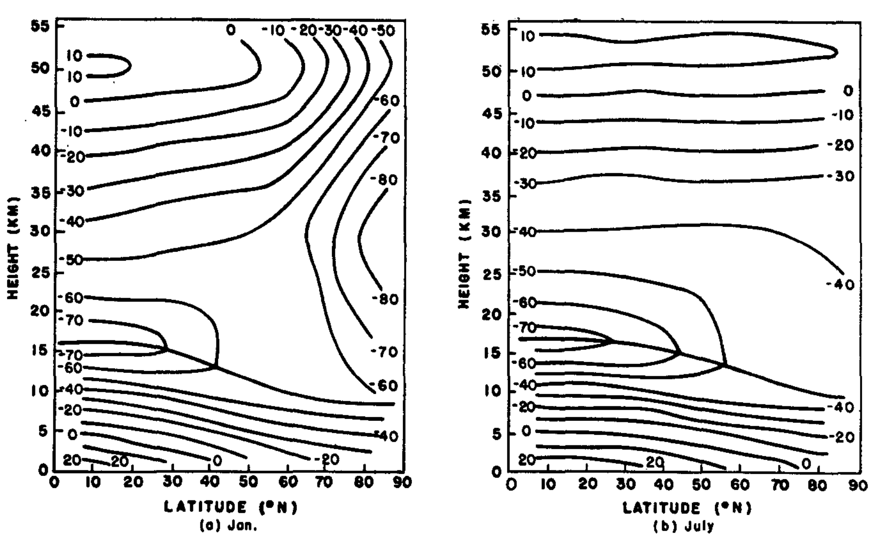

FIG. 2. Average temperature (C) cross sections for (a) Jan., (b) July.

were made for 10-deg-latitude belts in the northern hemisphere for the months of January, April, July, and October.

3. Composition and structure of the stratosphere

In order to compute atmospheric infrared fluxes, distributions of the following parameters are needed: (1) temperature—upon which the black-body energies depend, and (2) gas concentrations and atmospheric pressure—upon which the transmissivities depend. Mean models of the northern hemisphere were developed for four months—January, April, July, and October—representing the four seasons. The distributions are based upon direct and indirect observations and theoretical considerations and represent, especially for water vapor and ozone, only educated guesses.

The meridional-temperature cross sections shown in figs. 2 and 3 were derived from the following sources: (1) London (1957)—tropospheric temperatures, (2) Kochanski (1955)—temperatures in the layer from the tropopause to 32 km, and (3) extrapolation, miscellaneous measurements, and theoretical reasoning—temperatures above 32 km. The latitudinal and seasonal variations of the vertical pressure distribution are relatively unimportant as far as radiation calculations are concerned. It was assumed that the vertical pressure variation remains constant with latitude and season, and the National Advisory Committee for Aeronautics tentative upper-atmosphere pressures (Smithsonian Tables, 1951) were used in the computations.

In the few measurements of stratospheric water vapor that have been made (see summary by Goody, 1954), two general features seem to stand out: (1) the mixing ratio drops sharply in the first few kilometers above the tropopause, and (2) the mixing ratio does not vary greatly with height in the remainder of the stratosphere. The two features were incorporated into the present study by assuming that the specific humidity decreases exponentially with height in the layer from the tropopause to 21 km and that above 21 km the specific humidity remains constant with height, latitude, and season. For simplicity, this constant specific humidity in the upper stratosphere (above 21 km) was assumed to be the same as the specific humidity at the tropopause at 0N to 10N latitude. The values of specific humidity at the tropopause were taken from London (1957).

Measurements of the carbon dioxide content of the stratosphere (see, for example, Explorer II, 1938; Paneth, 1952) indicate that the volume percentage of carbon dioxide is about 0.03 per cent. As far as is known, carbon dioxide is thoroughly mixed in the stratosphere so that this value was assumed constant with latitude, season, and height.

The ozone cross sections (fig. 4) were constructed under the assumption that the vertical distribution depended upon the total amount of ozone present. Total amounts were taken from Götz (1951) and vertical profiles were then determined from Umkehr measurements summarized by Craig (1950).

4. Infrared emission of the stratosphere

The total infrared flux divergence (F net at 55 km minus F net at tropopause) represents the rate at which the stratosphere loses heat by infrared radiational processes. This heat loss varies with latitude and season since it depends upon the distributions of temperature and absorber concentration. Each one of the three emitting gases—carbon dioxide, water vapor, and ozone—contributes to the total flux divergence.

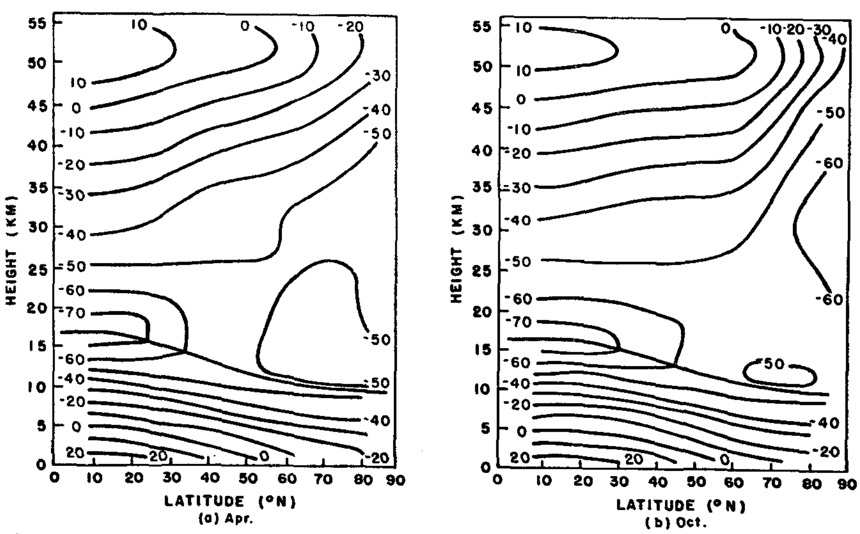

Fig. 3. Average temperature (C) cross sections for (a) April, (b) Oct.

Distributions of computed net infrared flux divergence for carbon dioxide, water vapor, and ozone, as functions of latitude and season, are shown in figs. 5, 6, and 7. The average stratospheric carbon dioxide flux divergence (16×10^{-3} ly min^{-1}) is much larger than the average water-vapor divergence (6×10^{-3} ly min^{-1}). This is due to the fact that carbon dioxide amounts decrease slowly with height compared to the rapid decrease of water vapor with height. In the troposphere, carbon dioxide is present in nearly black-body amounts so that the net flux changes very little with height and the net flux divergence in the troposphere is comparatively small (*i.e.*, compared to water vapor). In the stratosphere, carbon dioxide "opens up"—that is, the transmissivities are higher and the net flux increases rapidly with height, resulting

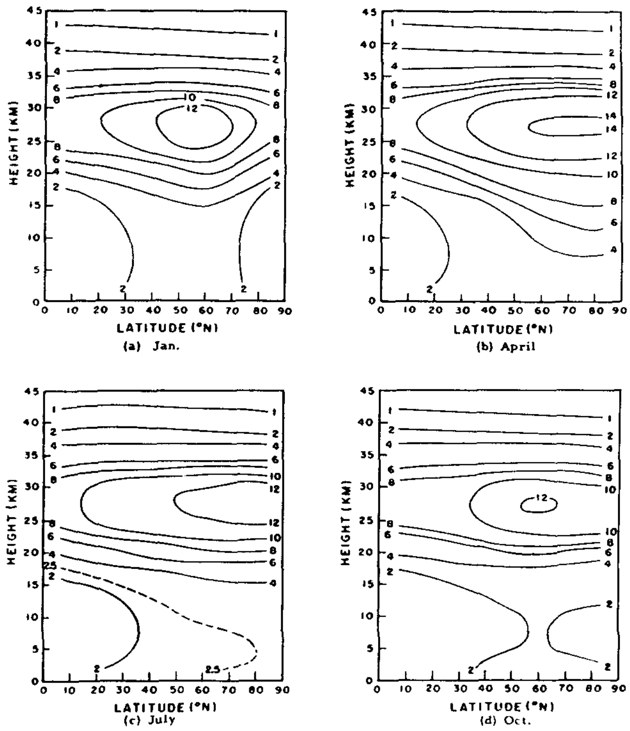

FIG. 4. The distribution of ozone concentration. (a) Jan., (b) April, (c) July, (d) Oct. Units: 10^{-3} cm STP km^{-1}.

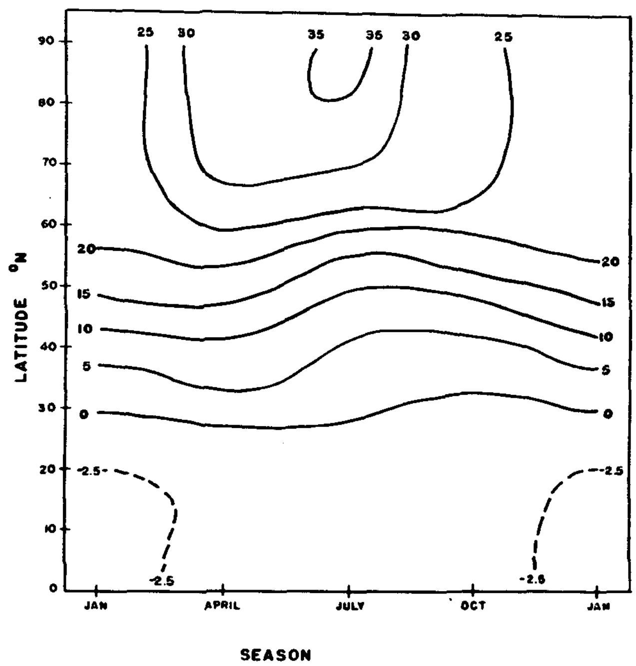

FIG. 6. Distribution of water vapor net flux divergence. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

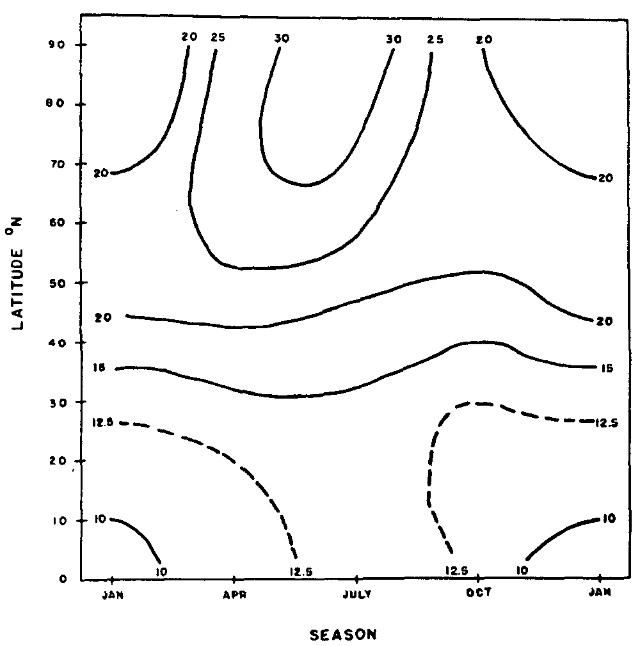

FIG. 5. Distribution of carbon dioxide net flux divergence. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

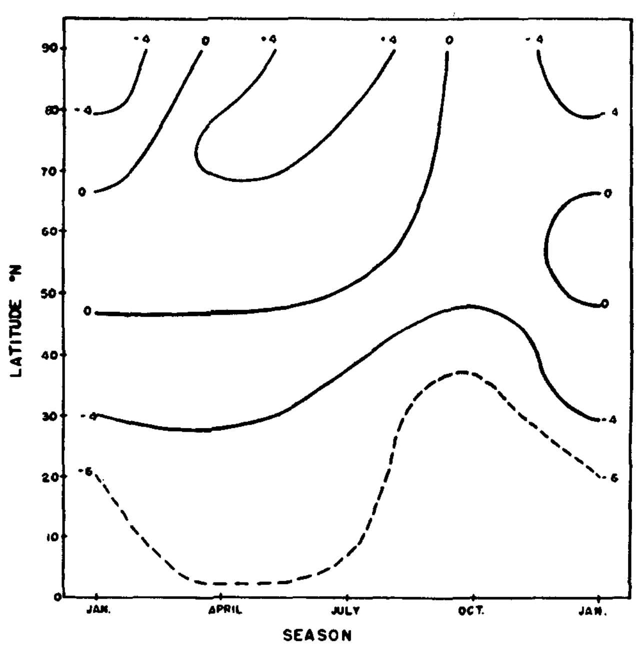

FIG. 7. Distribution of ozone net flux divergence. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

in a large divergence. On the other hand, water-vapor amounts decrease so rapidly with height that although the gas is present in nearly black-body proportions at the ground it is almost transparent at the tropopause. Thus, the water-vapor divergence is largely confined to the troposphere where it averages about 0.2 ly min^{-1}, over 30 times the water-vapor divergence in the stratosphere.

Latitudinally, the net flux divergences of carbon dioxide and water-vapor radiation generally increase toward polar regions. This is essentially due to the lower tropopause and thicker stratosphere at high latitudes which enable the net flux to increase considerably from its initial value at the tropopause. At low latitudes, the higher tropopause is at an elevation where the gases are more tenuous and more transparent. Thus, the net flux at the tropopause is increased very little as it passes through the stratosphere.

Infrared radiative transfer in the 9.6μ ozone band generally results in a warming (convergence of energy) of the stratosphere as can be seen from fig. 7. This effect was suggested by Dobson *et al* (1946) and is a consequence of the vertical ozone distribution. Since the troposphere contains very little ozone, the upward flux of energy in the 9.6μ band entering the stratosphere comes from an effective temperature near the ground. A large part of this energy is absorbed by the ozone layer and re-emitted upwards and downwards, but at a much lower temperature. The result is a net convergence of energy in the stratosphere. The magnitude of the convergence increases with larger differences in effective temperatures—that is, with

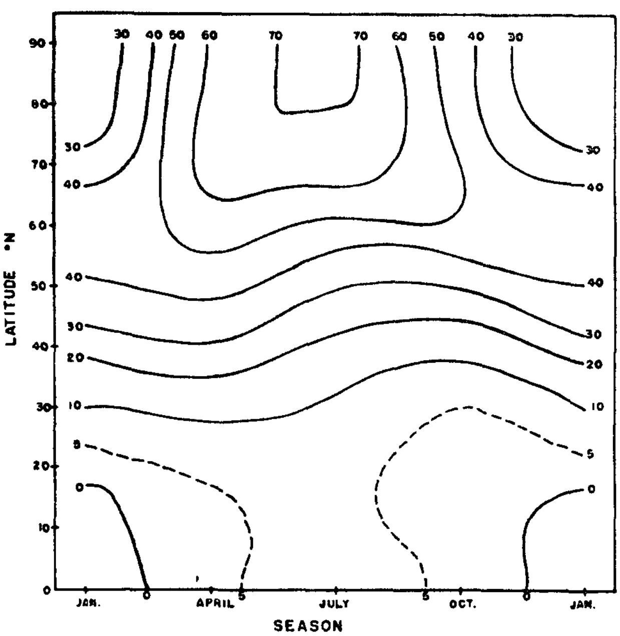

FIG. 8. Distribution of *total* infrared flux divergence. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

increasing instability. Averaged over latitude and season, the ozone infrared convergence amounts to 4×10^{-3} ly min^{-1}.

If the effects of all three gases are added together, the total infrared divergence of the stratosphere, which is equivalent to the net energy lost by infrared radiation, is obtained. Fig. 8 combines the characteristics of the preceding three diagrams:

(1) Increasing energy loss with increasing latitude.
(2) Maximum energy loss in July, except at middle latitudes where the maximum energy loss occurs in April.
(3) Minimum energy loss in January, except at middle latitudes where the minimum is displaced toward October.

The overall average loss of the stratosphere due to infrared radiational processes is about 18×10^{-3} ly min^{-1} or about 9 per cent of the average infrared loss of the troposphere.

5. Absorption of solar radiation by the stratosphere

Methods.—The major absorber of solar radiation in the stratosphere is ozone. This gas absorbs solar ultraviolet radiation in the Huggins (3200–3600 A) and Hartley (2000–3200 A) bands and visible radiation in the Chappuis (4500–6500 A) band. The results of Pressman (1954) which indicate the vertical, latitudinal, and seasonal variations of ozone absorption were integrated vertically in order to obtain the absorption of the entire stratosphere for each latitude belt and for each of the months considered in the present study.

Water vapor absorbs energy in the near infrared region of the solar spectrum. Empirical relationships for the absorption of solar radiation by water vapor have been derived from laboratory measurements and atmospheric observations (see, for instance, London, 1952). It has been found that the following formula can be used to estimate the absorption of solar radiation by water vapor:

$$a_z = 0.172[u^*(z)\,\overline{\sec\zeta}]^{0.3}\,\overline{\cos\zeta}\times P, \qquad (12)$$

where a_z is the absorption of insolation by water vapor in a vertical column extending from the height z to the top of the atmosphere (ly min^{-1}), $u^*(z)$ is the pressure-corrected water-vapor path length from the height z to the top of the atmosphere (cm STP), $\overline{\sec\zeta}$ and $\overline{\cos\zeta}$ are the average values of the secant and cosine of the zenith angle for the fifteenth day of each month considered, and P is the fraction of the day that the sun is shinning.

Distribution of absorption of solar radiation.—In the troposphere, the infrared energy loss is largely compensated for by the transport of latent and sensible heat from the lower boundary, the ground. In the stratosphere, however, the infrared emission is counterbalanced by direct absorption of solar radiation. The distributions of computed ozone and water-vapor absorption are presented in figs. 9 and 10. Because of the small amounts of water vapor in the stratosphere and the limited amount of energy in the near infrared region of the solar spectrum, the water-vapor absorption is quite small, averaging about one fourth of the ozone absorption. As a result of increasing duration of insolation with increasing latitude during spring and summer, both gases show maximum absorption at high latitudes during these seasons.

By adding the absorptions of ozone and water vapor, the energy gain of the stratosphere is obtained. The distribution of total absorption (fig. 11) indicates:

(1) Maximum absorption during spring and summer.
(2) Minimum absorption during fall and winter.
(3) Increasing absorption with increasing latitude during spring and summer, and decreasing absorption with increasing latitude during fall and winter.
(4) Little seasonal variation in equatorial latitudes.

6. The radiation budget of the stratosphere

The radiation budget is obtained by subtracting the infrared emission from the solar absorption at each latitude belt for each of the four months considered. The computed radiation budget is presented in fig. 12; positive values indicate a net excess of energy, and negative values indicate a net deficit. In general, local radiative equilibrium is not found in the stratosphere. Radiation excesses are found south of middle latitudes, and radiation deficits are found north of middle latitudes. This distribution of radiative energy implies that advection must play an important role in the redistribution of heat energy, for otherwise temperatures would continually increase with time at low latitudes and continually decrease with time at high latitudes. There must be, on the average, a net

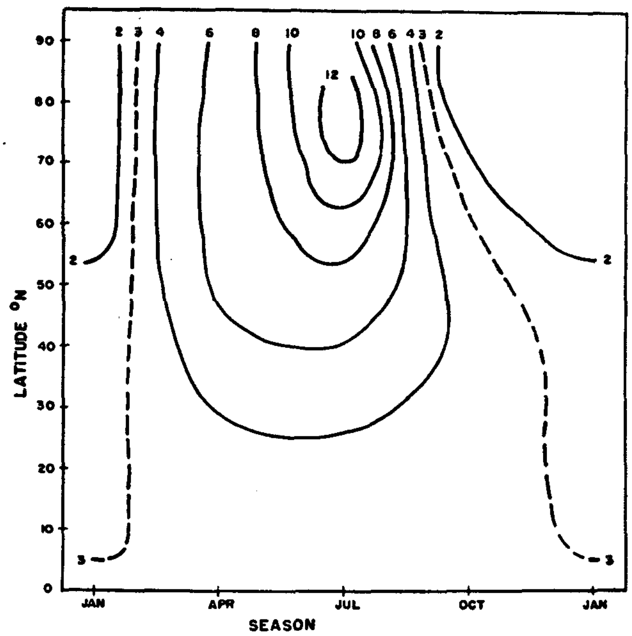

FIG. 10. Distribution of absorption of solar radiation by water vapor. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

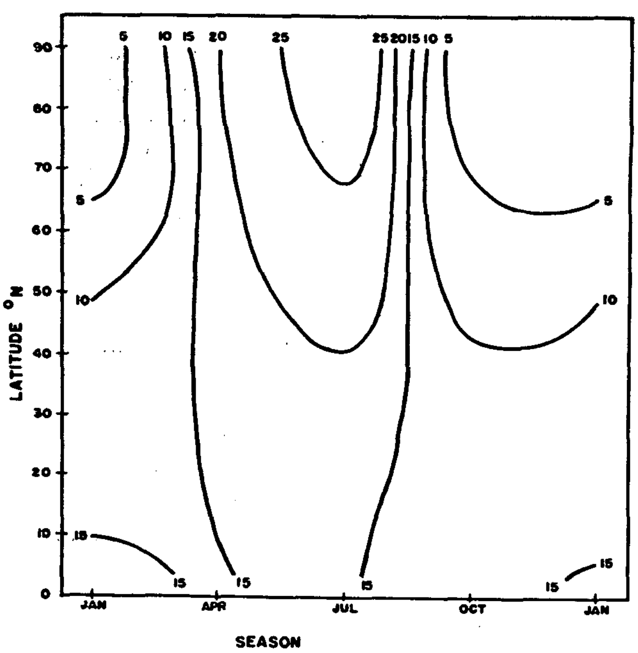

FIG. 9. Distribution of absorption of solar radiation by ozone. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

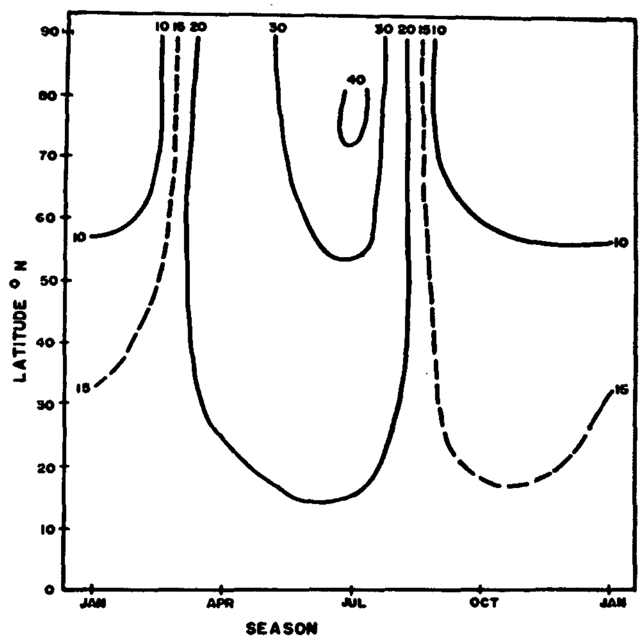

FIG. 11. Distribution of total absorption of solar radiation. (Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

transport of the excess energy of low latitudes to the deficit regions of high latitudes.

Observations indicate that, except for the lower stratosphere, the temperature decreases polewards in the stratosphere. Such a temperature variation is a consequence of the mean distribution of excess and deficit. In the lower stratosphere the mean latitudinal temperature variation is probably related to the meridional variation of the height of the tropopause and to non-radiational transfer processes through the tropopause.

If the excesses and deficits are averaged over the hemisphere, the average net excess (or net deficit) is obtained for each season (table 1).

TABLE 1. Hemispheric averages of net excess in the stratosphere (tropopause to 55 km). (Units: 10^{-3} ly min^{-1})

Month	Jan.	April	July	Oct.	Annual
Average net excess (+) or deficit (−)	−1	−2	+5	−2	0

If radiation is the only process transferring heat into or out of the stratosphere, one would expect an average hemispheric net excess in April when the stratosphere is warming up, an average net deficit in October when the stratosphere is cooling off, zero excess in January and July when stratospheric temperatures are at their minimum and maximum, respectively. The computed averages do not agree with the theoretical values because of two factors: (1) the assumption upon which the theoretical values are based—that radiation is the only process transferring heat into and out of the stratosphere—is probably not valid, and (2) the accuracy of the calculations is not sufficient for detecting the small average seasonal variations. The factor (1) would introduce some non-radiational heat transfer through the tropopause, through zero degrees latitude and through 55 km. This heat transfer would, of course, affect the temperatures and their seasonal variations. From the present calculations, it is impossible to determine just how large the non-radiational transfer might be.

The effect of one of the boundaries, the tropopause, can be partially eliminated if one studies the radiation budget of the 21 to 55 km region rather than the radiation budget of the entire stratosphere. By taking the lower boundary at 21 km, the effect of advection through the sloping tropopause of middle latitudes is eliminated and the effect of vertical transfer through the lower boundary is suppressed. A radiation budget for the upper stratosphere (21 to 55 km) is presented in fig. 13. The meridional distribution reveals radiation excesses in equatorial regions and deficits in polar regions, except during the late spring and summer. During the summer, the entire upper stratosphere is a region of radiation excess. The latitudinal and seasonal variations of net excess indicate that the temperature should decrease poleward except during the summer when a reversed temperature gradient could exist. Poleward transport of heat during winter and equatorward transport of

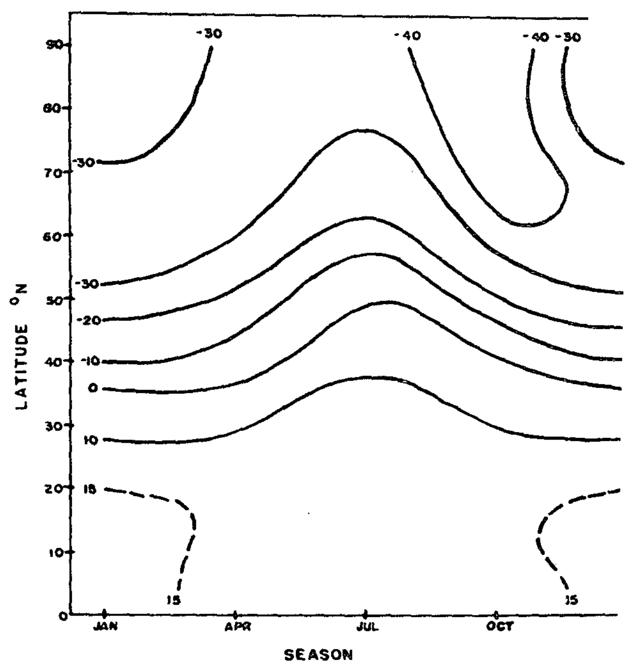

FIG. 12. The radiation budget of the stratosphere. (Absorption of solar radiation minus emission of infrared radiation—Tropopause to 55 km.) Units: 10^{-3} ly min^{-1}.

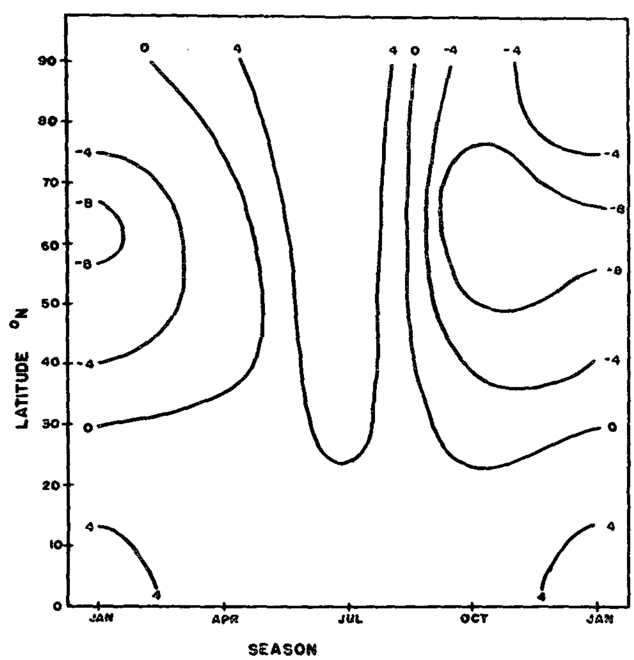

FIG. 13. The radiation budget of the *upper* stratosphere (Absorption of solar radiation minus emission of infrared radiation—21 km to 55 km.) Units: 10^{-3} ly min^{-1}.

heat during the summer are also suggested by the radiation budget. Mean meridional-wind cross sections published by Kellog and Schilling (1951) and Pant (1956) indicate a reversal of circulation from summer to winter in the upper stratosphere. This is consistent with the present radiation findings.

The relative differences between the April and October values at high latitudes are in the correct direction to explain the extremely large summer-to-winter temperature changes in the upper stratosphere at these latitudes. In October there are very large deficits, whereas in April there is an approximate balance.

If the radiation excesses and deficits in the upper stratosphere (21 to 55 km) are averaged over the hemisphere, the values shown in table 2 are obtained.

TABLE 2. Hemispheric averages of net excess in the upper stratosphere (21 to 55 km). (Units: 10^{-3} ly min^{-1})

Month	Jan.	April	July	Oct.	Annual
Average net excess (+) or deficit (−)	−1	−1	+3	−4	−0.8

As in the case of the entire stratosphere, the computed averages shown in table 2 do not agree with the theoretical values. Once again the discrepancy is due to insufficient accuracy in the computed results, and the invalid assumption of no non-radiational transfer through the boundaries. Although the effect of transport through the lower boundary has been minimized by taking it at 21 km rather than at the tropopause, the effects of transport through the upper boundary and through zero degrees latitude are still present. However, the relative difference between the April and October averages appears real and can be used to explain the temperature changes occurring in the upper stratosphere during these months.

Since stratospheric temperatures are at their maximum in July, the net excess of heat energy in this month is not being used to raise the temperature. Therefore, it appears likely that in the upper stratosphere there is a transport of excess heat energy from the northern hemisphere to the southern hemisphere during the summer months. There may also be a reverse transport of heat, from southern hemisphere to northern hemisphere, during winter and early spring.

7. Radiational heating and cooling rates in the stratosphere

The vertical distribution of the radiation budget can be presented in the form of radiational heating and cooling rates. Infrared cooling rates were computed from the vertical divergence of the net infrared fluxes for 2-km layers. Except for some modifications, solar-heating rates are based upon those given by Pressman (1955). Pressman's heating rates indicate a maximum at about 45 km whereas other calculations (London, 1951; Johnson, 1953), based upon observed ozone distributions, show that the maximum heating occurs higher up at about 50 km. Thus, Pressman's values were modified to indicate a heating maximum at 50 km and to include water-vapor absorption.

The July, 0N to 10N lat curves (fig. 14), are representative of southern latitudes where there are very small seasonal changes in the radiation budget. The 60N to 70N lat curves, for April and October (fig. 15 and 16), are representative of high latitudes and indicate the seasonal variations at these latitudes. The net-temperature-change curve is simply the difference between the heating and cooling values. All heating, cooling, and net-temperature-change curves were smoothed in order to iron out irregularities and obtain a reasonable pattern.

In general, radiative equilibrium is not obtained at any particular level. Instead, there is a vertical pattern of net radiational temperature change which varies with latitude and season. At lower latitudes there is a convergence of infrared energy in the first few kilometers above the tropopause, which shows up in July, 0N to 10N lat, as a net heating in this region.

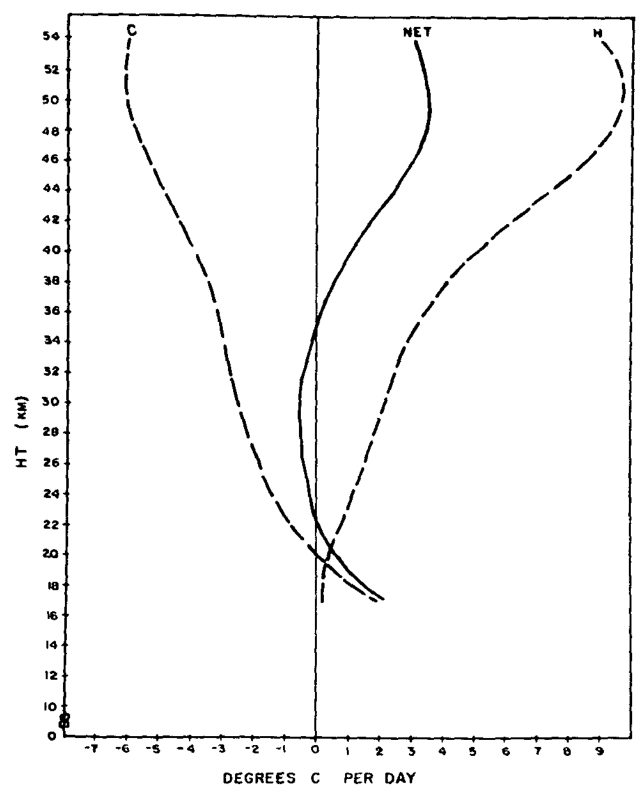

FIG. 14. Vertical distribution of solar heating rates (H), infrared cooling rates (C), and net radiational temperature change rates (Net). July 0N to 10N lat.

A heating of this layer due to infrared radiational transfer is reasonable since there is a pronounced minimum in the vertical temperature field at the equatorial tropopause and infrared transfer acts to smooth out such irregularities in the temperature structure. Thus, infrared radiation tends to destroy the equatorial tropopause.

Another interesting feature is the net radiational heating at the top of the stratosphere. This net radiational heating represents an excess of energy which is used to heat the 50-km region until an adiabatic lapse rate is formed in the 50- to 80-km region. Once an adiabatic lapse rate is established, there is a transport of heat energy upwards. The 50- to 80-km region is characterized by instability and turbulence (see, for example, Kellog and Schilling, 1951), and the heat source at 50 km is probably a major cause of these conditions.

Although the general shape of the heating and cooling curves is probably correct, the absolute magnitudes are somewhat uncertain, especially at higher levels. Thus, in discussing the net-temperature-change curve, the author has tried in a qualitative way to point out only those features which are apparently real.

8. Accuracy of results

The greatest source of error in the infrared calculations is the uncertainty in the transmission functions at short path lengths. Possible errors in the physical model and simplifications in the method are much less critical for the infrared divergence than are the uncertainties in the transmissivities. On the basis of sample calculations, the average hemispheric infrared divergence (18×10^{-3} ly min^{-1}) is estimated to have a possible error of 4×10^{-3} ly min^{-1}. The patterns of computed divergence are more likely to be correct than the absolute values at any particular latitude.

The values for the absorption of solar radiation could be in error by as much as 3×10^{-3} ly min^{-1} largely due to uncertainties in the ozone and water-vapor concentrations. The net radiation excesses and deficits, and especially their hemispheric averages, are more uncertain than the values of infrared emission or solar absorption since they are determined by the difference between absorption and emission. However, the pattern of excess and deficit should be fairly reliable and it is the pattern, rather than any individual value, that is most important for a general picture of radiative conditions in the stratosphere.

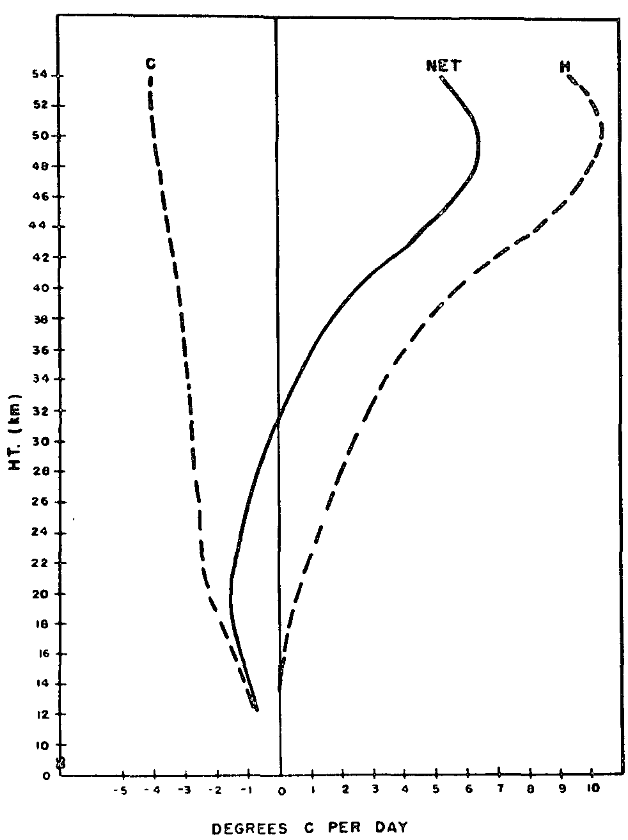

FIG. 15. Vertical distribution of solar heating rates (H), infrared cooling rates (C), and net radiational temperature change rates (Net). April 60N to 70N lat.

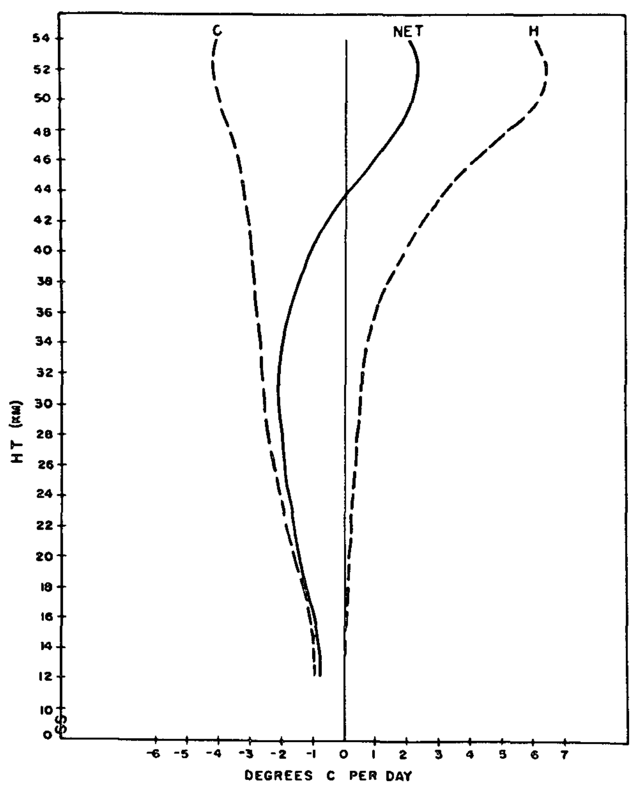

FIG. 16. Vertical distribution of solar heating rates (H), infrared cooling rates (C), and net radiational temperature change rates (Net). October 60N to 70N lat.

9. Summary and conclusions

The major results of the present investigation can be summarized as follows:

(1) The radiation budget of the entire stratosphere (tropopause to 55 km) indicates that low latitudes act as heat source and high latitudes act as a heat sink.

(2) The radiation budget of the upper stratosphere (21 to 55 km) indicates a similar latitudinal distribution of radiational energy. Seasonally, spring and summer are periods of radiation excess and fall and winter are periods of radiation deficit.

(3) Carbon dioxide is the most important gas in the stratosphere as far as infrared cooling is concerned; ozone acts to produce a convergence of infrared energy in the stratosphere.

(4) The tropopause region at low latitudes (0N to 40N lat) is heated by the action of infrared transfer. The top of the stratosphere (about 50 km) is, in general, a region of net radiational heating.

A redistribution of radiative energy in the stratosphere is presumably accomplished by means of a meridional heat flux. There is probably a poleward transport of heat during fall and winter and, at least in the upper stratosphere, an equatorward transport of heat during late spring and summer. Advection of heat energy into and out of the northern hemisphere stratosphere apparently occurs at the lower boundary, the tropopause, and at the southern boundary, the equator. As a consequence of the heat source at about 50 km, convection of heat energy out of the stratosphere takes place.

The latitudinal distribution of radiation excess and deficit suggests that the temperature decreases northward in the stratosphere. The large seasonal temperature changes in the upper stratosphere at high latitudes are due to the differences in radiative properties of the upper stratosphere between April and October.

Radiative equilibrium is not obtained at any point in the stratosphere. Nor is it obtained for the entire stratosphere in any season. The uncertainties in the results do not permit any definite statement on whether the stratosphere is in radiative equilibrium on an annual basis. However, the average annual absorption and emission computed in the present study do approximately balance each other.

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