RADIATIVE VARIATIONS OF THE AIR TEMPERATURE

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Some formulas derived using the data of actinometric sounding for use in calculation of radiative cooling of air when processing the data of airborne and radiosonde sounding are verified in this paper. It is shown that they make it possible to precalculate vertical distribution of air temperature in the lower part of the boundary layer under the cloudless conditions and are useful for estimating the radiative cooling of the upper troposphere at the downwelling adiabatic air motion.

To solve some problems of atmospheric physics and optics, it is necessary to precalculate the air temperature distribution in a separate layer or a volume. In particular, one of such problems is the prediction of vertical distribution of temperature in the mixing layer, that governs the regime of aerosol accumulation in this layer, balance of the adiabatic heating of air at the dounwelling motion and its simultaneous cooling due to radiative cooling. Some other tasks can also be mentioned.

According to Ref. 1, if the gradient of effective atmospheric radiation Φ is determined, the temporal variation of temperature at the altitude Z_0 is unambiguously related to it:

$$dT/dt = -(1/C_p\,\rho)\,(d\Phi/dZ), \qquad (1)$$

where T is the air temperature, t is time, ρ is the air density, and C_p is the specific heat at constant pressure.

The effective flux of atmospheric radiation is determined by the difference between the downwelling $U(Z)$ and upwelling $Q(Z)$ fluxes:

$$\Phi(Z) = U(Z) - Q(Z). \qquad (2)$$

The problem on calculating the effective radiation flux $\Phi(Z)$ is in the fact that the atmosphere radiates not like the absolutely black body. It is difficult to represent this deviation in the analytical form, so the empirical and semiempirical formulas are usually used to calculate $\Phi(Z)$.

Some of such formulas are given in Ref. 1. The accuracy of calculations using the majority of them is tested in Ref. 2–7. These tests show that in some cases the error in calculating $\Phi(Z)$ and determining the value dT/dt from $d\Phi/dZ$ does not fit the accuracy required in practice. So the search for new techniques for calculating $\Phi(Z)$ is being continued.

Creation of actinometric radiosondes[8] and the network of actinometric stations[9] in the former USSR made it possible to study the spatiotemporal variability of the effective atmospheric radiation and its downwelling and upwelling components in many physico-geographical regions in different seasons in the height range up to 30 km and sometimes higher.

The radiative fluxes averaged over all sites and years are most closely connected with the average temperature at the level where the flux has been measured. The statistical analysis made it possible to obtain the following relationships:

$$U(Z) = (2.650 - 0.360\,x - 0.061\,x^2)\,\sigma T^4,$$
$$P_Z = 1000 - 300 \text{ hPa}; \qquad (3)$$
$$Q(Z) = (0.158 + 0.260\,x)\,\sigma T^4;$$
$$U(Z) = (0.675 - 0.333\,x)\,\sigma T^4;$$
$$P_Z = 300 - 150 \text{ hPa}; \qquad (4)$$
$$Q(Z) = (-0.350 + 0.380\,x + 0.003\,x^2)\,\sigma T^4.$$

Here $x = \log P_Z$, where P_Z is the air pressure at the level Z, for which the fluxes are calculated, T is air temperature, σ is the Stefan–Boltzmann constant.

The aim of this paper is to compare the air temperature values calculated by formulas (1)–(4) with the measurement data.

To precalculate the vertical distribution of temperature in the mixing layer, let us use the data of airborne sounding over the city of Tomsk obtained during the dark periods under the anticyclone conditions when the advection of heat is absent. The values of temperature were measured each 0.1 km altitude, the measurement error did not exceed 0.5°C. Then the vertical profile obtained at the ascending flight is the principal one, and the profile obtained at landing 6 –8 hours after was used for a control.

Calculations have been made for 100 m thick layers. The value obtained is referred to the middle of this layer. Time interval was 0.5 or 1 hour. The formulas from Refs. 2–7 were used for calculations of the air temperature near the ground surface for the clear sky conditions. On the whole 28 events were tested for the period since 1981 till 1985. All they were observed under clear sky conditions.

An example of such calculations is shown in Fig. 1. As is seen, good agreement, within the measurement error, is observed in the layer up to 0.7 km, i.e. in the lower part of the inversion. Near the inversion level the calculation data and measurement differ by 2 to 3°C that is not satisfactory.

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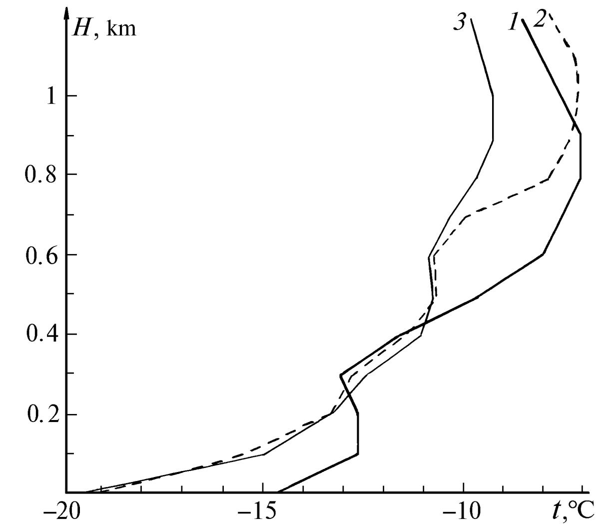

FIG. 1. Vertical distribution of temperature over the city of Tomsk, December 20–21, 1981: 1) profile measured at the ascending flight (6 pm, local time); 2) profile calculated by Eqs. (1)–(3); 3) profile measured at landing (1^{50} am, local time)

The average characteristics, over all 28 events, used in calculations are given in Table I. It follows from these data that the calculations by Eqs. (1)–(3) provide quite good accuracy in the lower layer up to 0.8 km, then the error increases in the layer where the inversions are often observed. Above 1 km the error in calculations decreases again.

Such differences between the calculated and measured values of temperature are possibly caused by the fact that aerosol and gas pollutants of air are accumulated in the inversion layer, what leads to the deviation of the effective radiation fluxes described by Eq. 3.

Many authors, explaining the sharp increase in the ozone concentration near the ground surface, relate this process to the ozone transfer from stratosphere to the troposphere through the tropopause by the system of descending motions that appear in the jet streams.[10,11]. However, the authors of Ref. 12 argued them and noted that if the motion has occurred from the stratosphere down to the ground surface, it should increase the air temperature up to 40°C. The observed temperature increase is only 13°C in comparison with the neighbor air. In addition, the ozone transfer in such a system exists, and the results obtained in Ref. 13 evidence this fact. This transfer is accompanied by the adiabatic heating of air, but not so intense.[14]

TABLE I. Root mean square deviation (RMSD) of calculated air temperature and its measured value in the boundary layer.

Height, km	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
RMSD, °q	0.28	0.31	0.3	0.32	0.35	0.34	0.34	0.48	0.58	1.75	2.52
Height, km	1.10	1.20	1.30	1.40	1.50	1.60	1.70	1.80	1.90	2.00	2.10
RMSD, °q	2.85	2.54	2.12	1.78	1.12	1.09	0.94	0.87	0.78	0.70	0.68

One can remove the contradiction by taking into account the fact that adiabatic heating of air due to descending motion should be accompanied by the radiative cooling, increasing as the intensity of the descending motion increases. The relationships (4) make it possible to calculate such cooling and estimate the balance between the values of adiabatic and radiative variations of the air temperature.

The calculations by Eqs. (1), (2), and (4) are shown in Fig. 2. The calculations used the velocity of vertical motion of air as an argument. The function was the rate of the air temperature variation. This caused by the fact that the value of adiabatic heating is proportional to the descending motion velocity, and the radiative cooling is proportional to the fourth power of the heating. It was assumed, for calculations, that first the heating of the descending air occurs and then the radiative cooling begins. Temperature variation was calculated for the 1 hour interval. The layer thickness was determined as the product of the descending motion velocity and time. The temperature variation rate was recalculated for the layer thickness of 1 km. The reference altitude was 12 km. Vertical gradient of temperature for the calculation of adiabatic heating was taken to be equal to zero.

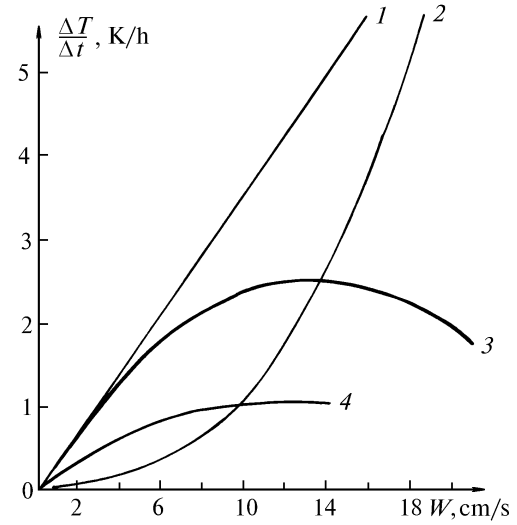

FIG. 2. The rate of adiabatic (1), radiative (2), total (3), and real (4) variation of air temperature in the tropopause layer as a function of the vertical motion speed.

It is seen from Fig. 2 that the rate of adiabatic heating for the small values of the descending motion velocity is greater than the radiative cooling rate. Their differences are maximum at the velocity of 12–14 cm/s. The rates of adiabatic heating and radiative cooling become equal at the velocity of the descending motion of 30 cm/s. Thus, the data obtained evidence the fact that the process of air heating by the descending motions should saturate at the velocity of 12–14 cm/s.

To check this conclusion, we have considered the air temperature variations in the jet streams at different intensity of vertical motion. Air temperature variations were determined from the pressure field maps at the levels closest to the jet stream axis using the technique of fixed parameters.[15] The velocity of vertical motion was calculated using a single point technique[16] from the data of radiosounding. Comparison of the data is shown in Fig. 3 involving 68 events.

It is shown in Fig. 3 that actual variation of air temperature is nonlinear. One can approximate this variation by the function of the following form

$$\Delta T/\Delta t = w^{0.9} \pm 1.06.$$

Then the curve constructed from the field of points shows the tendency towards saturation near the level of 8 K/6 hours.

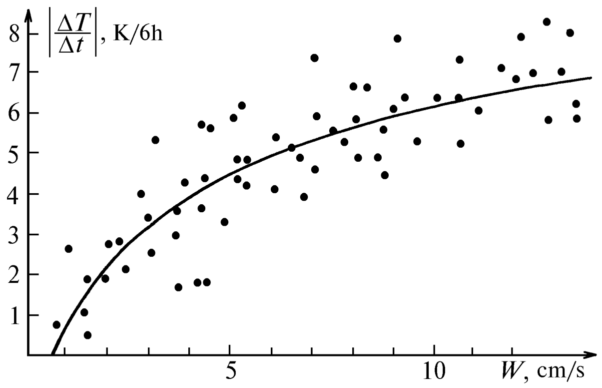

FIG. 3. Air temperature variation in the jet stream for different vertical motion velocities in a spiral circulation system

The same curve drawn in Fig. 2 (*4*) follows the behavior of the resulting curve (*3*), within the limits of the vertical velocity range under study, but there are differences in the absolute values. On the average, the values of actual rate of the air temperature variation are two times lower than the rates of the calculated variations.

Such differences can appear due to the following reasons. According to Refs. 13 and 14, the zones of elevated ozone content and elevated temperature in comparison with the surrounding air have the shape of a band elongated along the jet stream axis. In this case, the zone of enhanced temperature emits energy not only in the upward and downward directions, as it has been assumed in this paper, but also to the sides that was not taken into account in calculations. There are no reasons to suppose that the intensity of radiative cooling due to the side (horizontal) emission is less than the vertical one.

Summing up the results, let us note that the empirical formulas, obtained from the data of actinometric sounding, make it possible to precalculate the vertical distribution of air temperature in the lower part of the boundary layer and they are useful for estimating the rate of radiative cooling of the air layer that is adiabatically heated by the descending motions.

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