Quantitative investigation of stratospheric mixing processes by means of long lived radon decay products

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(Manuscript received October 15, 1965)

ABSTRACT

A model of tracer transport process in the meridional plane of lower stratosphere is briefly outlined and the corresponding partial differential equation for tracer concentration is deduced. For the long lived radon daughters RaD(Pb^{210}) and RaF(Po^{210}) in the Northern Hemisphere stratosphere such equations can be simplified to one dimensional stationary form, describing the quasivertical transport of these isotopes. The mean values of quasivertical velocity and turbulent diffusion coefficient in the lower stratosphere of different latitudes of Northern Hemisphere are estimated by comparison of last equation solutions with the published results of aircraft measurements of RaD and RaF concentration in 1961 and 1962. These estimated values agree with the known qualitative ideas on latitudinal and seasonal changes of the stratospheric mixing intensity.

1. Introduction

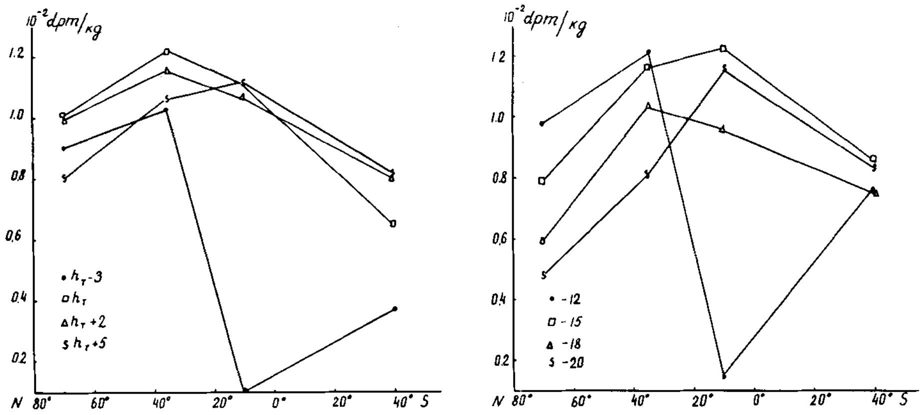
The use of different tracers (especially radioactive ones) in meteorology has increased considerably in recent years. Important qualitative ideas about the circulation and exchange mechanisms in the stratosphere and mesosphere were obtained by means of fission products W^{185}, Rh^{102}, Cd^{109} and also ozone, water vapor, Be^7 and other natural tracers (SHEPPARD, 1963; KAROL & MALAKHOV, 1962, 1965; KALKSTEIN, 1962; JUNGE, 1963). However, the development of modern meteorology and weather forecasting, based on the hydro- and thermodynamics, requires more precise knowledge of factors and parameters governing the general circulation of the atmosphere. Such data are especially important for the stratosphere and mesosphere, as common aerological means do not allow us to perform there the meteorological measurements with the required frequence and precision, and some meteorological elements cannot be determined at all. For instance, the poor knowledge of air mass radiation characteristics allows the computation of only so called "adiabatical" vertical velocities in the stratosphere, but their correspondence to the real vertical motions is unclear (SHEPPARD, 1963; OORT, 1965).

In this paper a method of quantitative determination of exchange parameters in the lower stratosphere from tracer concentration measurements is proposed. Parameters of vertical exchange are calculated by use of published results of the long lived radon daughters: lead-210(RaD) and polonium-210(RaF) measurements in the stratosphere.

2. Transport of tracer substances in the meridional plane of the lower stratosphere

The analysis of the observed concentration distributions for some radioactive isotopes in the lower stratosphere leads to the conclusion that the longitudinal concentration changes are small in comparison to the vertical and the latitudinal ones. So for the first approximation it can be assumed that these distributions are zonally homogeneous and isotope transport takes place almost identically in the lower stratosphere in all meridional planes. This is true both for ozone and radioactive isotopes of natural origin and for products of nuclear explosions some time (several months) after the explosion (STEBBINS, 1961; MACHTA *et al.*, 1964).

The analysis of the observed vertical profiles for several isotope concentrations (Sr^{90}, W^{185}, Rh^{102} and others) reveals that the height of their maxima in the lower stratosphere changes with latitude similarly to the tropopause height. Its maximum is at the equator and minima



a *b*

FIG. 1. Meridional distribution of RaD concentration obtained from aircraft measurements in May 1960–61: (*a*) at the tropopause level h_T; at the level 3 km below the tropopause ($h_T - 3$); at the levels 2 and 5 km above the tropopause ($h_T + 2$; $h_T + 5$); (*b*) at the levels 12, 15, 18, 20 km.

are in polar regions (NEWELL, 1963; TELEGADAS & LIST, 1964; KAROL & MALAKHOV, 1965). The isolines of measured RaD concentration in the lower stratosphere of both hemispheres, shown in Fig. 1, have also a similar latitudinal distribution.

These facts together with the investigations of ozone transport from the equatorial stratosphere to the polar regions and some other observations lead to the conclusion that in the lower stratosphere the meridional macroturbulent exchange occurs mainly along directions inclined towards the poles and crossing the potential temperature isolines (isentropes) under some small angle, as has been revealed by Starr and his colleagues (NEWELL, 1964*a*, *b*; OORT, 1965).

These directions are tangent to the curves in the meridional plane having the following equation in spherical coordinates system

$$r = r_0 + \alpha R(\theta); \quad R(0) = 0,$$

where r is the vertical coordinate; θ the colatitude, α the height difference of the tropopause between the equator and the poles. The exact position of these curves is yet unknown, but to a first approximation they can be assumed to be almost equidistant from the tropopause outside the zone of its subtropical break. In the meridional plane in the lower stratosphere a new system of orthogonal coordinates (P, ϑ) is considered with the coordinate lines

$$P(r,\theta) = r - r_0 - \alpha R(\theta) = \text{const}; \quad \vartheta = \vartheta(r,\theta) = \text{const}.$$

The direction ϑ-constant is only slightly different from the vertical one. It was estimated by NEWELL (1964*b*) that the curve P-const makes an angle of about 7′ with the isentrope direction in the zone 45–50° N.

From the equation of mass conservation, written in the vector form:

$$\frac{\partial(\varrho c)}{\partial t} + \operatorname{div}(\varrho c\mathbf{v} + \varrho c w \mathbf{e}_r) = Q \tag{1}$$

after averaging over the time and over the latitudinal circles and separating the average values and fluctuations

$$\mathbf{v} = \bar{\mathbf{v}} + \mathbf{v}'; \quad c = \bar{c} + c'; \quad \varrho = \bar{\varrho}; \quad w = \bar{w},$$

we obtain the following equation:

$$\frac{\partial c}{\partial t} + (\bar{\mathbf{v}} + \bar{w}\mathbf{e}_r)\operatorname{grad}\bar{c} + \frac{1}{\varrho}\operatorname{div}(\bar{\varrho}\overline{c'\mathbf{v}'}) - \frac{\bar{c}}{\bar{\varrho}}\operatorname{div}(\bar{\varrho}\bar{w}\mathbf{e}_r) = \frac{\bar{Q}}{\bar{\varrho}}. \tag{2}$$

The following notations are used here: ϱ = air density; $c(P, \vartheta, w)$ = concentration per air mass unit of substance fraction, which is connected with aerosols, having the gravitational velocity w; $\mathbf{v}$ = velocity vector; $\mathbf{e}_r$ = vertical unit vector; Q = sum of substance sources and sinks intensities in the volume unit. Assuming that the macroturbulent diffusion in the meridional plane is going according to the semiempirical law:

$$\overline{c'\mathbf{v}'} = -K \operatorname{grad} \bar{c}, \tag{3}$$

where in every point has the tensor K the main axes tangent to the coordinate lines (P, ϑ). After some calculations we obtain

$$\frac{\partial \bar{c}}{\partial t} + \frac{\partial \bar{c}}{\partial P}\left(\bar{v}_p - \bar{w} - K_p \frac{\partial \ln \bar{\varrho}}{\partial P}\right) + \frac{1}{r}\frac{\partial \bar{c}}{\partial \vartheta}\left[\bar{v}_\theta - \frac{K\vartheta}{r}\left(\frac{\partial \ln \bar{\varrho}}{\partial \theta} + \operatorname{ctg}\theta\right)\right] = \frac{\partial}{\partial P}\left(K_p \frac{\partial \bar{c}}{\partial P}\right) + \frac{1}{r^2}\frac{\partial}{\partial \vartheta}\left(K_\vartheta \frac{\partial \bar{c}}{\partial \vartheta}\right) + \frac{\bar{c}}{\bar{\varrho}}\frac{\partial(\bar{\varrho}\bar{w})}{\partial r} + \frac{\bar{Q}}{\bar{\varrho}}, \tag{4}$$

—the macroturbulent diffusion equation in the meridional plane of the lower stratosphere. This equation differs only slightly from the common diffusion equation in spherical coordinates. The main difference is that in the "quasivertical" mean velocity component $\bar{v}_p$:

$$\bar{v}_p = \bar{v}_r - \bar{v}_\theta \alpha R'(\theta)/r \tag{5}$$

enters the meridional velocity component $\bar{v}_\theta$.

Now the problem is to determine the coefficients of this equation, K_p, K_ϑ, V_p, using its solution and measurements of concentration $\bar{c}$. This is the "inverse problem" from the mathematical point of view.

The estimates of K_p and K_ϑ values were made by using the observations of W^{185} and others nuclear explosion products spreading during the first months after the nuclear weapon test series (STEBBINS, 1961; SHEPPARD, 1963; TELEGADAS & LIST, 1964), but there are few such data.

The different times and latitudes of the injections of these products in the stratosphere is the cause of the temporal and spatial inhomogeneity of the global stratospheric reservoir of artificial isotopes. This inhomogeneity does not allow us to simplify the diffusion equation and to find the unknown parameters in an easy way.

The measurements of cosmogenic isotopes such as Be^7, S^{35} and some others revealed that their halflives (50–60 days) are too short for tracing the slow stratospheric motions, as their concentrations are close to their equilibrium values (STEBBINS, 1961). More suitable for this purpose is Na^{22} with half life 2.6 years. Though measurements are scanty, some estimates of K_p values in the lower stratosphere were obtained in this way by BHANDARY & RAMA (1963).

The use of C^{14} and H^3 is complicated by the fact that these isotopes (as also Be^7 and Na^{22}) are of both cosmogenic and artificial origin.

The long-lived radon daughters RaD and RaF are the convenient tracers in the stratosphere, as their half lives (22 years and 138 days) are comparable with the time scales of stratospheric planetary motions. The source of these isotopes in the stratosphere is the radon, which, exhalating from the land area of the Earth surface and entering the stratosphere, leads to the creation of the almost stationary reservoirs of RaD and RaF (KAROL, 1965).

Their concentration inhomogeneties along the surfaces P-const. are small in comparison to the vertical ones, as it is evident from Fig. 1 and as follows from the observation analysis, which will be published elsewhere. Therefore in the diffusion equation, the terms describing the quasihorisontal transport can be omitted in a first approximation and only the stationary (quasivertical) transport of RaD and RaF in the Northern Hemisphere stratosphere need be considered. This is a rather rough approximation as it does not take into account the horizontal advection of RaD from the temperate latitudes of Northern Hemisphere—the place of main radon entrance in the stratosphere (KAROL, 1965). It is also possible that some quantity of RaD atoms should enter the stratosphere as product of nuclear explosions. But there are no definite evidences of that, and therefore it is assumed here that all the considered RaD atoms are of natural origin.

3. The model of RaD and RaF vertical distribution in the lower stratosphere of the Northern Hemisphere

Let us consider the one-dimensional stationary problem of determination of the $c_j^{(i)}(z, w)$ concentration distribution of i-isotope in radioactive decay chain in the j-layer of the lower stratosphere. The system of equations of this problem is the following:

$$K_j \frac{d^2 c_j^{(0)}}{dz^2} - 2K_j b_j^{(0)} \frac{dc_j^{(0)}}{dz} - \lambda^{(0)} c_j^{(0)} = 0; \quad i = 0 \text{ for Rn}; \tag{6}$$

$$K_j \frac{d^2 c_j^{(i)}}{dz^2} - 2K_j b_j \frac{dc_j^{(i)}}{dz} - (\lambda^{(i)} - \mu_j) c_j^{(i)} = -\lambda^{(i-1)} c_j^{(i-1)}$$

$$i = 1 \text{ for RaD},^1 \quad i = 2 \text{ for RaF}, \tag{7}$$

where

$$2b_j = -\frac{d \ln \varrho}{dz} + \frac{v_{pj} - w_j}{K_j} = \frac{B_j}{T_j} + \frac{v_{pj} - w_j}{K_j};$$

$$B_j = \frac{g}{R} - \gamma_j;$$

$$b_j^{(0)} = b_j|_{w=0}; \quad \mu_j = \frac{1}{\varrho} \frac{d(\varrho w)}{dz} = \frac{\gamma_j w_j}{2T_j};$$

K_j is the coefficient of quasivertical turbulent diffusion; T = temperature; $\gamma = -dT/dz$ is the temperature gradient; $R = 287$ m²/sec² °C is the gas constant; g = acceleration of gravity; z is the coordinate in the "quasivertical direction", ϑ-const, which is practically coincident with the real vertical direction and z is considered here as a vertical coordinate. All parameters and coefficients are assumed to be constant in the considered j-layer.

The solution of equation (6), which attains the given values at the borders of the layer:

$$c_j^{(i)}|_{z=h_j} = c_j^{(i)}(h_j); \quad c_j^{(i)}|_{z=H_j} = c_j^{(i)}(H_j); \quad h_j < H_j, \tag{8}$$

has the form:

$$c_j^{(0)}(z) = A_j^{(0)}[h_j, H_j; \quad c_j^{(0)}(h_j); \quad c_j^{(0)}(H_j)]$$

$$= c_j^{(0)}(h_j) \exp[b_j^{(0)}(z - h_j)] \frac{sh a_j^{(0)}(H_j - z)}{sh a_j^{(0)}(H_j - h_j)}$$

$$+ c_j^{(0)}(H) \exp[b_j^{(0)}(z - H_j)] \frac{sh a_j^{(0)}(z - h_j)}{sh a_j^{(0)}(H_j - h_j)};$$

$$a_j^{(0)} = \sqrt{b_j^{(0)2} + \frac{\lambda^{(0)}}{K_j}}. \tag{9}$$

Analogous, but more complicated formulae, containing the operators $A_j^{(l)}$ for $l = 1, 2 \ldots i$ can be obtained for the solution of system of equation for RaD and RaF which satisfies the boundary condition (8). This solution is considerably simplified if the j-layer is assumed to have an infinite thickness, and if the condition

$$H_j \to +\infty; \quad c_j^{(j)}(H_j) \to 0 \tag{10}$$

is valid. Then we obtain

$$c_j^{(0)}(z) = c_j^{(0)}(h_j) \exp \beta_j^{(0)}(z - h_j); \quad \beta_j^{(i)} = b_j^{(i)} - a_j^{(i)}. \tag{11}$$

This means that the operator $A_j^{(i)}$ is transformed into the simple exponential function. For clearness we shall investigate this case only and omit the index j, though the singularities of the general case of finite thickness layer will be pointed out.

The RaD and RaF atoms in the atmosphere are attached to the natural aerosols. It can be assumed that in the lower stratosphere the distribution of the number $n(r)$ of aerosol particles as a function of their effective radii r has the form

$$dn/d \ln r = A_K r^{-\varkappa_K},$$

where A_K = const, $\varkappa_1 = 0$ for condensation nuclei with radii $r_0 = 0.01\ \mu < r < r_1$ and $\varkappa_2 = 3$ for large sulfate particles having $r_1 < r < r_2 = 2 \div 3\ \mu$ with $r_1 = 0.1 \div 0.5\ \mu$ (JUNGE *et al.*, 1961; STEBBINS, 1961; KAROL, 1965). This model of distribution density is presented in Fig. 2. Assuming then that the average number of isotope atoms per particle is proportional to Br^m (B = const; $1 < m < 3$), we introduce the distribution of this number as the function of gravitational settling velocity w by the formulae:

$$\frac{r}{r_1} = \frac{w}{w_1}; \quad q_m(w) = B\left(\frac{r_1}{w_1}\right)^{m+1} \int_{w_0}^{w} w^m \frac{dn}{dr} dw;$$

$$q_m(w_2) = 1.$$

The graphs of this distribution density are presented in Fig. 2 for some m-values. Now the total i-isotope concentration over all the fractions of the stratospheric aerosols is given by the formula:

$$C^{(i)}(z) = \int_{w_0}^{w_2} c^{(i)}(z, w)\, dq_m(w).$$

As the density of $q_m(w)$ for $m = 2 \div 2.5$ has a sharp maximum, corresponding to "critical

[1] In the stratosphere the short-lived radon daughters are in radioactive equilibrium with radon, and RaE with RaD.

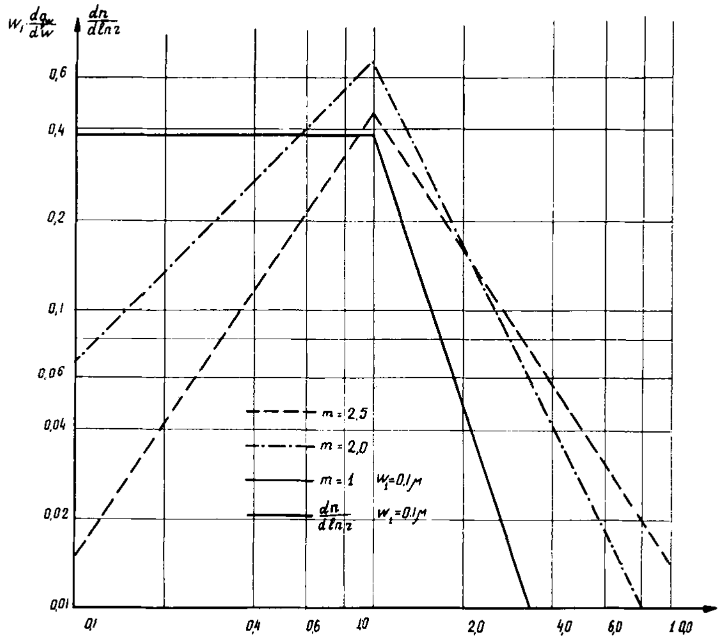


FIG. 2. Models of frequency distribution of (a) number $n(r)$ of aerosol particles as a function of their effective radii r—full line; (b) average number of isotope atoms per particle, having the radius r and gravitational velocity w—full line for $m = 1$, dashed line for $m = 2.5$, dotted-dashed line for $m = 2$.

values" $r = r_1(w = w_1)$ like the known delta-function, the filtering property of the latter function can be used for the approximate calculation of this integral.

Thus we obtain with sufficent accuracy for the summary RaD concentration ($i = 1$):

$$C^{(1)}(z) \simeq \exp\,[\beta_*^{(1)}(z-h)]\,[C^{(1)}(h) + c^{(0)}(h)\,I^{(1)}] - c^{(0)}(z)\,I^{(1)}, \quad (13)$$

where $\beta_*^{(i)} = \beta^{(i)}\big|_{w=w_1}$;

$$I^{(i)} = \int_{w_0}^{w_1} \frac{dq_m(w)}{1 - \lambda^{(i)}/\lambda^{(0)} - (w\beta^{(0)} + \mu)/\lambda^{(0)}}.$$

The $I^{(i)}$ values for RaD and RaF are very close to 1. The results of MACHTA & LUCAS' (1962) measurements and those for RaD (HASL, 1965) reveal that in the lower stratosphere the ratio radon to RaD concentrations $c^{(0)}(h)/C^{(1)}(h)$ does not exceed 1 per cent. Omitting the small terms in (13) yields:

$$C^{(1)}(z) = C^{(1)}(h)\,\exp\,[\beta_*^{(1)}(z-h)] \quad (14)$$

Following the same procedure for dealing with the RaF concentrations we obtain the following expression:

$$C^{(2)}(z) + \lambda^{(1)}C^{(1)}(z)/(\lambda^{(1)} - \lambda^{(2)}) = [C^{(2)}(h) + \lambda^{(1)}C^{(1)}(h)/(\lambda^{(1)} - \lambda^{(2)})]\exp[\beta_*^{(2)}(z-h)]. \quad (15)$$

Similar formulae can be deduced for the layer of finite thickness with the substitution of the operators $A^{(i)}$ for the exponential functions.

All these formulae are approximate for polydisperse aerosols, but they are exact for monodisperce aerosols, having the "critical" values r_1 and w_1.

4. Estimation of vertical turbulent diffusion coefficient and vertical velocity

The $\beta_*^{(1)}$ value is found by means of formula (14) from the relation of measured specific RaD activities at two-levels z and h. The relation of measured specific RaF to RaD activities at one level allows to calculate the quantity

$$\chi(z) = \frac{\lambda^{(2)}C^{(2)}(z)}{\lambda^{(1)}C^{(1)}(z)} + \frac{\lambda^{(2)}}{\lambda^{(1)} - \lambda^{(2)}},$$

for which we have from (14) and (15)

$$\chi(z) = \chi(h)\,\exp\,[(\beta_*^{(2)} - \beta_*^{(1)})\,(z-h)] \quad (16)$$

TABLE 1. *K and V_p estimates from aircraft measurements of RaD and RaF concentrations in lower stratosphere.*

Month and year	Latitude (degrees north)	Height of layer boundaries (km)	Number of averaged measurements	K(m²/sec) $r_1 = 0.1\ \mu$	K(m²/sec) $r_1 = 0.5\ \mu$	$V_p(10^{-2}$ cm/sec) $r_1 = 0.1\ \mu$	$V_p(10^{-2}$ cm/sec) $r_1 = 0.5\ \mu$
VI.1961	10–20	15–18	4–5	0.15	—	0.7	3.0
X.1961	15–20	19–21	1	0.29	0.28	−1.1	1.3
IV.1962	5–15	14–17	5	2.1	2.1	−7.3	−5.6
IV.1962	5–15	17–21	4	0.82	0.70	−1.3	1.2
VI–VII.1961	25–30	15–20	6	0.35	0.32	−0.4	1.7
IX.1961	25–35	15–18	3	0.76	0.75	−2.7	−1.2
IX.1961	25–35	18–21	7	0.47	0.47	−2.5	−0.6
XI.1961	25–30	18–20	5	0.27	0.26	−0.3	2.2
XII.1961	25–30	17–20	3	0.18	0.17	−0.2	1.8
VII.1962	30–35	14–20	2	1.1	1.1	−2.9	−0.9
IX.1961	35–45	15–21	2	2.0	2.0	−5.8	−3.9
X.1961	35–45	15–20	2	0.24	0.24	−0.4	1.6
XII.1961	35–45	15–20	2	0.23	0.23	0.0	1.9
VI.1962	35–45	12–21	2	3.4	3.2	−5.7	−3.6
X.1961	45–55	15–20	2	3.1	3.1	−8.9	−7.0
XII.1961	45–55	15–20	2	4.0	4.0	−6.9	−5.2
III.1962	45–55	14–20	3	1.4	1.4	−6.4	−4.5
VI.1962	45–50	12–21	1	1.6	2.0	−3.2	−1.6
V.1962	60–65	11–15	1	13	13	−39	−37
V.1962	60–65	15–20	1	9.4	9.4	−23	−21
VII.1962	65–70	9–15	3	6.1	6.1	−13	−11

and so we find $\beta_*^{(2)}$. The exchange parameter values can now be calculated by the formulae:

$$K = \frac{\lambda^{(2)}\beta_*^{(1)} - \lambda^{(1)}\beta_*^{(2)}}{\beta_*^{(1)}\beta_*^{(2)}(\beta_*^{(2)} - \beta_*^{(1)})} + \frac{\mu}{\beta_*^{(1)}\beta_*^{(2)}};$$

$$v_p - w_1 = K\left(\beta_*^{(i)} - \frac{P}{T}\right) - \frac{\lambda^{(i)} - \mu}{\beta_*^{(i)}}. \qquad (17)$$

As the temperature gradient γ is small μ is small in the lower stratosphere of temperate and high latitudes, the additive terms in the last formulae are small also. In this case the K value is practically independent of w_1 and the same is true for the $v_p - w_1$ value. If the $v_p - w_1$ value exceeds 0.1 cm/sec some uncertainity introduced by w_1 does not influence the obtained v_p values, but in general the inadequate knowledge of stratospheric aerosol size distribution is the cause of the marked inaccuracy in K and especially v_p estimates.

Similar formulae for calculation of K and $v_p - w_1$ can be obtained for the layer of finite thickness using the measured RaD specific activity at three different levels (at the boundaries and in the middle of the layer). The use of these formulae requires statistically good concentration measurements at three levels in homogeneous atmospheric layer. Such data are almost absent in the published measurements of RaD and RaF in the stratosphere (KAROL, 1965; HASL, 1965); therefore we shall not use here the formulae for the layer of finite thickness.

It must be mentioned, however, that the condition of vanishing concentration at infinity at the upper level of infinite thickness layer implies that the RaD concentration must diminish with height in this layer. In fact this occurs mainly for zero or descending vertical flows. For significant positive v_p values the RaD concentration can even increase with height and to this case the above formulae for K and $v_p - w_1$ cannot be applied. So the method described can estimate preferably negative v_p values as will be seen below.

Aircraft measurements of RaD and RaF concentrations in the stratosphere below the 20–21 km level were performed in the second half of 1961 and in 1962 over Alaska, over the western part of continental U.S.A. over Central America, over central and southwestern parts of the Pacific and over the southern Australia. The results of radiochemical analysis of filters with aerosol samples have recently been published by HASL of U.S.AEC (1965). The estimates of K and v_p values, calculated by means of above formulae (14)–(17) are listed in Table 1. Here the mean monthly values of RaD specific activity and of the ratio of RaF to RaD specific activities for some regions in the Northern Hemisphere and for some months with more numerous observations are used together with some climatic temperature data for the lower stratosphere (KHANEVSKAYA, 1964; MOSKALEVA, 1965). All the estimates were obtained for two "critical" values of particle radius $r_1 = 0.1$ μ and 0.5 μ, shown to be the most probable ones by the recent investigations (JUNGE *et al.*, 1961; ROSENBERG & NIKOLAEVA-TERESHKOVA, 1965).

It is evident from the table that the calculated K and $v_p - w_1$ values are almost independent of the r_1 values even in the tropical stratosphere with large temperature gradient. The K value increases from equator to pole and from winter to spring and summer in accordance with the known qualitative ideas on the seasonal and latitudinal changes of vertical turbulent diffusion coefficient in the lower stratosphere (STEBBINS, 1961; KAROL & MALAKHOV, 1965). The numerical values of K are close to the values obtained in author's previous publication (KAROL, 1965) for tropical and subtropical latitudes, but in high latitudes the new K values are by an order of magnitude higher than the previous estimates, where the vertical velocity was not taken into account.

The estimated order of v_p magnitude $\sim 10^{-2}$ cm/sec is equal to the velocity in the lower stratosphere of different latitudes for the I.G.Y. period, calculated by OORT (1965) obtained by using the adiabatic assumption. A detailed comparison is so far too early to make, either because of small number of RaD and RaF measurements, or because of the significant uncertainty, introduced in v_p value by the meridional velocity component and the rather arbitrary choice of w_1 aerosol gravitational velocity "critical value". It can be noted however, that the v_p estimates have in general the bigger negative values at high latitudes than in tropics and former values are close to the estimate 1.5 km/month $= 6 \cdot 10^{-2}$ cm/sec, obtained by TELEGADAS & LIST (1964) from observations of Rh^{102} downward propagation. This corresponds also to some ideas on meridional circulation of the stratosphere (MURGATROYD & SINGLETON, 1961; KAROL & MALAKHOV, 1965).

5. Conclusion

The method proposed here, besides its direct use for quantitative investigations of stratospheric circulation, can be applied to the study of some important meteorological phenomena in the stratosphere.

For the solution of problems concerning the vertical or advective character of "explosive warmings" of the polar stratosphere in spring (GODSON, 1963), the RaD and RaF measurements at several stratospheric levels at the time and in the region of this phenomenon should be performed with the subsequent calculation of the real vertical velocities.

In recent papers of TUCKER (1964) and REED (1964) semiempirical models of distribution and 26-month oscillation of zonal wind and temperature in the equatorial stratosphere are presented and velocity estimates of probable meridional and vertical motions—associated with these models are obtained. For the verification of these models, such estimates should be compared with the vertical velocity values, obtained by the method under consideration. The available amount of RaD and RaF measurements does not allow us to make such comparison now.

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КОЛИЧЕСТВЕННОЕ ИЗУЧЕНИЕ ПРОЦЕССОВ ОБМЕНА В СТРАТОСФЕРЕ С ПОМОЩЬЮ ДОЛГОЖИВУЩИХ ПРОДУКТОВ РАСПАДА РАДОНА

Кратко описана модель процесса переноса примеси в меридиональной плоскости нижней стратосферы и выведено соответствующее дифференциальное уравнение для концентрации примеси. Для долгоживущих продуктов распада радона RaD(Pb^{210}) и RaF (Po^{210}) в нижней стратосфере северного полушария такие уравнения могут быть приведены к одномерному стационарному виду, описывающему квазивертикальный перенос этих изотопов. Получены оценки средних квазивертикальных значений скорости и коэффициента турбулентной диффузии в нижней стратосфере различных широт северного полушария путем сравнения решений последних уравнений с опубликованными результатами самолетных измерений концентрации RaD и RaF в 1961–62 гг. Эти оценки согласуются с известными качественными представлениями о сезонных и широтных изменениях интенсивности перемешивания в стратосфере.