STUDY OF INTERNAL GRAVITY WAVES IN THE METEOR ZONE

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An important component of the dynamical regime of the atmosphere at heights near 100 km are internal gravity waves (IGW) with periods from about 5 min to about 17.5 hrs. which propagate from the lower atmospheric layers and are generated in the uppermost region of the atmosphere. As IGW propagate upwards, their amplitudes increase and they have a considerable effect on upper atmospheric processes: 1) they provide heat flux divergences comparable with solar heating (SHVED, 1977; GAVRILOV and SHVED, 1975, 1982; CHUNCHUZOV, 1978, 1982); 2) they influence the gaseous composition (GAVRILOV and POGORELTSEV, 1980) and produce wave variations of the concentrations of gaseous components and emissions of the upper atmosphere (KRASSOVSKI, 1972; WEINSTOCK, 1978; GAVRILOV and SHVED, 1982; GAVRILOV and YUDIN, 1982); and 3) they cause considerable acceleration of the mean stream (CHUNCHUZOV, 1971; GAVRILOV, 1976; GAVRILOV and SHVED, 1982), etc.

Considerable progress in IGW studies has been achieved in recent years in connection with the MAP and GLOBMET research programs. Interest in IGW studies has increased notably. This can be traced to the mathematical experiments on numerical modelling of the middle atmospheric circulation at 10-150 km altitude started in the 1980's. These experiments have revealed that the drag created by molecular and turbulent viscosity is insufficient for the formation of the meridional circulation observed, and therefore an additional source of acceleration must exist. It is proposed that IGW serve as such a source (HOLTON, 1982, 1983; MIYAHARA, 1984).

The considerable effect of IGW on middle atmospheric circulation has stimulated interest in IGW studies. At present IGW in the upper atmosphere are investigated by all the methods available, both new (MST radars; BALSLEY, 1983; BALSLEY et al., 1983; the use of surface industrial explosions; GOKHBERG, 1983) and traditional (rocket soundings, HIROTA, 1984; observations of variations in nocturnal airglows characteristics, KRASSOVSKI et al., 1978; MOLINA, 1983; SHEFOV et al., 1983; methods of incoherent scattering and partial reflections, VINCENT, 1984; radar location of meteor traces, etc.). These investigations have yielded ample information about IGW near the mesopause. Taking into consideration the specific subject of this Conference, we shall dwell on the results obtained making use of the radiometeor method.

IGW studies by the radiometeor method were started in France (SPIZZICHINO, 1969, 1972) and later continued in other countries. In the USSR this kind of research was started jointly by the scientists of Leningrad State University and groups of researchers in Kharkov and Frunze (GAVRILOV and DELOV, 1976; GAVRILOV et al., 1976). These groups continued their research (KARIMOV and LUKYANOV, 1979, KALCHENKO et al., 1985). Later similar investigations were carried out in Obninsk and Kazan (KAZANIKOV and PORTNYAGIN, 1981a, b; SIDOROV et al., 1983). The investigations differ in the instrumentation and methods of statistical identification of IGW. Table 1 presents the IGW parameters measured by different scientists and the IGW characteristics calculated from the parameters measured. In early works (SPIZZICHINO, 1969; GAVRILOV and DELOV, 1976; GAVRILOV et al., 1976; KARIMOV and LUKYANOV, 1979) individual values of wind velocities for single meteors were averaged over the entire horizontal area of the meteor zone observed by the radar. Therefore, only the amplitude V, period τ_w and vertical wavelengths λ_z were measured in cases when the the determination of echo height made it possible to divide the meteor zone into a number of layers according to height. KAZANIKOV and PORTNYAGIN (1981a, b) measured the wind simultaneously in four mutually perpendicular directions. The use of coherent analysis for spatially remote regions made it possible to determine the horizontal wavelengths, velocities and directions of propagation of IGW with large horizontal wavelengths $\lambda_h \gtrsim 700$ km.

In recent papers (KALCHENKO et al., 1985; GAVRILOV and KALOV, 1985) an algorithm for singling out IGW from radiometeor data has been used which was developed earlier by GAVRILOV (1981, 1984a). The algorithm is based on dividing the area observed by the radar into a number of subareas in the direction of the beam of vision of the antenna. Statistical analysis of the velocity spectra obtained for these areas makes it possible to discriminate IGW from the background noise and to determine the horizontal wavelengths λ_{11} of IGW along the axis of the antenna direction diagram. If the radar can measure echo height (KALCHENKO et al., 1985) it is possible to measure λ_z.

The radiometeor method is limited in the measurement of IGW parameters because of the necessity of average data over a large spatial area and the finite number of meteor echoes per time unit (GAVRILOV and DELOV, 1976; GAVRILOV, 1984a). As a rule, IGW with $\tau_w \gtrsim 0.5$-1 h, $\lambda_{11} \gtrsim 100$-200 km and $\lambda_z \gtrsim 5$-30 km are accessible to measurement. In this connection, it is interesting to note an original method proposed by SIDOROV et al., 1983, for measuring IGW during meteor showers when a great number of meteor radioechoes located practically on the same plane makes it possible to measure IGW with periods of several minutes.

IGW in the meteor zone have small but markedly variable λ_z. Therefore, when measured at stations which do not measure echo height, they yield only rough, considerably smoothed estimates. The advent of automated height measuring radars with a large number of useable radioechoes and appropriate vertical resolution (KALCHENKO et al., 1985) has made it possible to measure the IGW vertical wavelengths and to improve the accuracy of the determination of wave characteristics. The results presented in the paper by KALCHENKO et al., (1985) appear to be, for the present, the most comprehensive and accurate estimates of the IGW parameters from radiometeor data; they are given in Table 2 and are in good agreement with the results of previous radiometeor measurements and data obtained using other methods.

It can be concluded that the periods, wavelengths, amplitudes and velocities of IGW propagation in the meteor zone are now measured quite reliably. However, for estimating the influence of IGW on the thermal regime and the circulation of the upper atmosphere these parameters are not as important as the values of wave fluxes of energy, heat, moment and mass.

Table 1

IGW parameters measured and calculated from radiometeor data.

Authors	Measured IGW Parameters	Calculated IGW Parameters
Spizzichino, 1969, 1972	V, λ_z, $\tau \gtrsim$ 1h	F_z
Gavrilov and Delov, 1976	V, λ_z, $\tau \gtrsim$ 1h	F_z
Karimov and Lukyanov, 1979	V, $\tau \gtrsim$ 1h	
Kazanikov and Portnyagin, 1981a,b	$\lambda_h \gtrsim$ 1000 km V, A, $\tau \gtrsim$ 1h	F_z, F_{xz}, F_{yz}
Sidorov et al., 1983	V, 8 min $\lesssim \tau \lesssim$ 1h	
Kalchenko et al., 1985	V, λ_z, $\lambda_{11} \lesssim$ 1000 km $\tau \gtrsim$ 0.5 h	
Gavrilov and Kalov, 1985	V, $\lambda_{11} \lesssim$ 1000 km, $\tau \gtrsim$ 0.5 h	F_z,F_x,F_y,F_{xz},F_{xy}, F_{Tx}, F_{Ty}, F_{Tz}, F_{mx}, F_{my}, F_{mz}

Note: V is the amplitude, τ is the period, λ_h and λ_z are the horizontal and vertical wavelengths, λ_{11} is the length along the antenna horizontal axis, A is the azimuth of IGW propagation, F_i, F_{ij}, F_{Ti} and F_{mi} are the components of the wave fluxes of energy, moment, heat and mass, respectively.

Table 2

Values of IGW characteristics from radiometeor measurements (KALCHENKO et al., 1985) and estimations (GAVRILOV and KALOV, 1985).

Parameter	Range of values	Parameter	Range of values
τ	0.5-6 hrs	F_{xz}	1-4 $m^2 s^{-2}$
V	6-20 $m\ s^{-1}$	F_{yz}	1-10 $m^2 s^{-2}$
λ_{11}	100-800 km	F_{xy}	2-20 $m^2 s^{-2}$
λ_z	10-30 km	F_{Tx}	1-20 erg $cm^{-2} s^{-1}$
$\underset{\sim}{C}$	20-160 $m\ s^{-1}$	F_{Ty}	1-200 erg $cm^{-2} s^{-1}$
$\underset{\sim}{F}_z$	1-6 erg $cm^{-2} s^{-1}$	F_{Tz}	1-6 erg $cm^{-2} s^{-1}$
$\underset{\sim}{F}_x$	1-20 erg $cm^{-2} s^{-1}$	F_{mx}	2-10 10^{-8} kg $m^{-2} s^{-1}$
$\underset{\sim}{F}_y$	1-10 erg $cm^{-2} s^{-1}$	F_{my}	1-4 10^{-6} kg $m^{-2} s^{-1}$
		F_{mz}	1-3 10^{-8} kg $m^{-2} s^{-1}$

The difficulty here is that to calculate the above mentioned wave fluxes, which require knowledge of the wave variations of pressure, density, temperature and vertical velocity, while the radiometeor method makes it possible to measure only the horizontal wind velocity. In a paper by GAVRILOV (1984b) a method is proposed for calculating the wave fluxes of energy, heat, moment and mass from radiometeor data using IGW theory. The method has been applied in GAVRILOV and KALOV (1985) to statistically analyze the Obninsk radiometeor measurements. Although no echo height information is available from Obninsk, a great amount of observational material (about 9,000 IGW harmonics) and application of statistical modelling has made it possible to obtain sufficiently reliable estimates for the monthly mean values of wave fluxes which are presented in Table 2 and which provide the evidence of the important contribution of IGW to the energetics and dynamics of the upper atmosphere.

Study of the IGW vertical structure has revealed that from 60 to 80% of the IGW harmonics in the meteor zone propagate upwards (KALCHENKO et al., 1985; VINCENT, 1984). The IGW amplitudes increase with height (Fig. 1a), the increase being influenced by the superposition of oscillations due to partial reflection of the IGW energy (SPIZZICHINO, 1969, 1972; GAVRILOV and DELOV, 1976; KALCHENKO et al., 1985). An interesting feature is a zone of increase of IGW decay at heights about 92 km which has been discovered by KALCHENKO et al., (1985). A decrease in the IGW amplitudes and a disturbance of the linear phase variation with height can be systematically observed in this zone (Fig. 1).

When the seasonal variation of IGW parameters in the meteor zone is discussed, the existence of an annual cycle with the maximum amplitude in winter and minimum in summer is mostly generally accepted. However, recent research has revealed a more complicated situation. Fig. 2a presents the seasonal and latitudinal distribution of the IGW intensities at 20-70 km obtained by HIROTA (1984) from the analysis of meteorological rocket soundings. It can be seen that from 40°N northwards an annual harmonic with a maximum in winter is predominant in the seasonal variation of the IGW intensity. However, at low latitudes a semi-annual variation with maxima in spring and autumn prevails. This kind of pattern is confirmed by observations at heights about 100 km from radiometeor data (GAVRILOV and KALOV, 1985; KARIMOV and LUKYANOV, 1979) and from observations of wave variations of the night airglow emission of atomic oxygen 0 λ 5577A (GAVRILOV and SHVED, 1982), as is shown in Fig. 3.

In the annual variation of the IGW amplitudes V in Obninsk at 56°N (Fig. 2b) the annual harmonic prevails, whereas at stations located in more southward areas (FRUNZE, 43°N and ASHKHABAD, 38°N) the semi-annual variation of V appears with maxima in spring and autumn. The reasons for seasonal variations in the IGW intensity are supposed to be the changes in the activity of wave sources in the lower atmospheric layers and also changes in the filtering properties of the middle atmosphere due to seasonal adjustments of the circulation in the strato-mesosphere.

Interesting results have been obtained from radiometeor investigations of the intra- and inter-diurnal variations of the IGW energy. Fig. 3c shows the results of the analysis of the IGW energy variations E_w with τ_w 1-6 hrs from the data of field expedition measurements of the Kharkov group

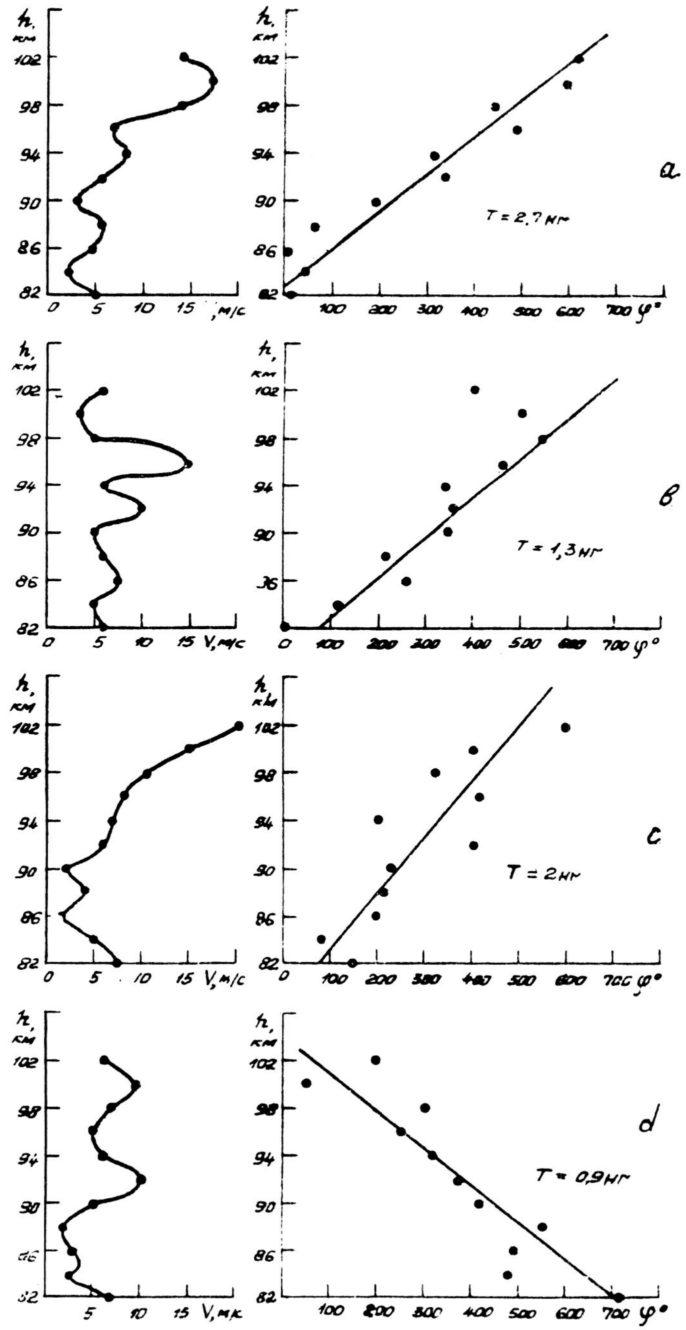

Fig. 1 Altitude variations of amplitudes (left) and phases (right) of period T IGW harmonics.

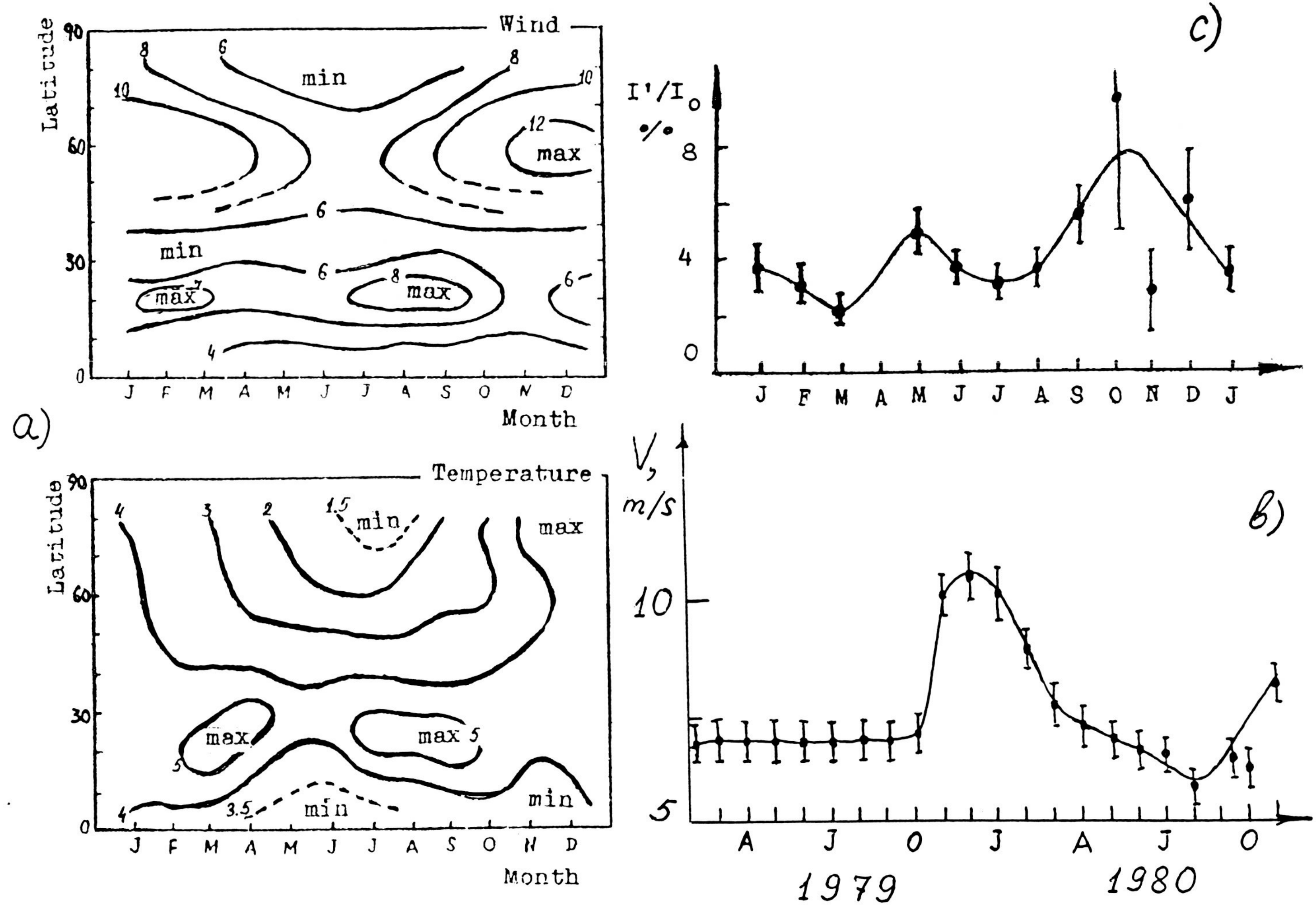

Fig. 2 Latitude and seasonal distributions of IGW amplitudes in the stratosphere (a) and seasonal variations of IGW amplitudes at 90-100 km over Obninsk (b) and Ashkhabad (c).

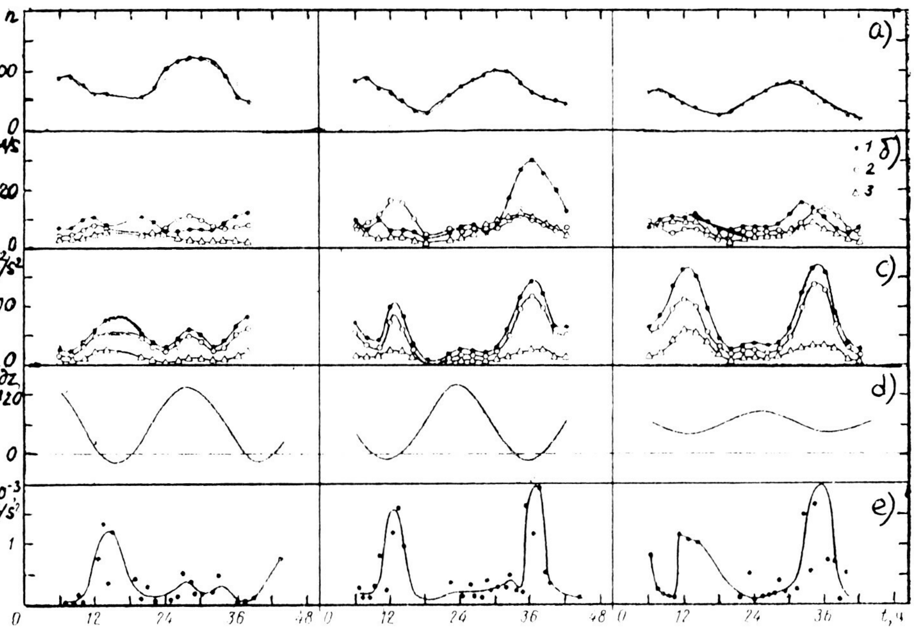

Fig. 3 Variations of meteor rates (a), IGW amplitudes (b), IGM intensity (c), potential temperature vertical gradient (d) and intensity of turbulence (e) at 85-90 km (left), 90-95 km (center), and 95-100 km (right) in the tropical meteor zone.

in the tropical zone on May 7-8, 1970 (GAVRILOV et al., 1981). The diurnal variations of E_w are quite obvious. Fig. 3d presents the calculated variations of the vertical gradient of potential temperature in the diurnal tide. It can be seen that the maxima of E_w coincide with the passage of the zones of convective instability of the tide. Similar diurnal and semi-diurnal variations of E_w have also been found at the middle latitudes (KALCHENKO et al., 1985; GAVRILOV and KALOV, 1985), the diurnal harmonic being dominant. In the middle latitudes the tidal amplitudes are smaller than in the tropics, and no convective instability appears. The cause of the diurnal and semidiurnal variations of E_w here is apparently the modulation of propagating IGW caused by the tidal fields of wind and temperature.

In recent years great attention has been paid to the study of tropospheric sources of acoustic gravity waves in the atmosphere. These are earthquakes, industrial explosions, mountains, displacement instabilities, convective motions, thunderstorm clouds, etc. However, all these are local in both time and space. In particular, mountains are not located everywhere and instabilities, earthquakes or thunderstorms in the troposphere do not always appear, whereas wave motions in the upper atmosphere are observed at all times and places. Hence, the problem arises of finding permanently active sources of wave motions in the atmosphere. Such a source is known: it is the non-linearity of the atmospheric hydrodynamic equations due to which any dynamical formation in the atmosphere (a turbulent or cyclonic vortex, jet, front, etc.) generates wave motions.

For the sake of simplicity, let us examine the case of a two-dimensional barotropic model. In consideration of fast waves, the natural time scale in $T_w = L/c$, where L is the horizontal scale of the stream and c is the sound velocity. When transformed to dimensionless coordinates, the hydrodynamic equations have two-dimensionless parameters; $f = lL/c$ (l is the Coriolis parameter), and the Mach number M introduced as a perturbation parameter before the nonlinear terms. The conventional method of expansion of hydrodynamic variables into asymptotic series over the perturbation parameter M converges only if $\tilde{t}M < 1$, where $\tilde{t} = t/T_w$ is dimensionless time. In order to overcome this difficulty, we shall use the recently developed "method of multiple timescale successive approximations" (LEIBOVICH and SIBASS, 1977; SITENKO, 1977; JEFFREY and KAWAHARA, 1982) where time variables in the hydrodynamics equations are presented as asymptotic series $\partial/\partial\tilde{t} = \partial/\partial\tau + M\,\partial/\partial T + M^2\,\partial/\partial\theta + \ldots$, where τ, T, θ...is a hierarchy of formally introduced temporal variables which are considered to be independent and which are chosen to ensure the convergence of the asymptotic series in M, with $\tau = O(\tilde{t})$, $T = O(M\tilde{t})$, $\theta = O(M^2\tilde{t})$, etc. In other words, the variable θ characterizes changes in the hydrodynamic fields with the scale of fast waves T_w; the variable T characterizes the slow advective evolution with scale $M^{-1}T_w$, and the variable θ characterizes still slower changes with scale $M^{-2}T_w$, etc.

The use of the method of multiple timescale successive approximations yields, for the zero-order terms in M, a system of linear equations, the solution of which after the initial period of adaptation (when waves originated by the maladjustment of the initial conditions attenuate) tends to background fields independent of and satisfying the equation:

$$\beta U_o = -\partial\pi_o/\partial y \;;\; \beta V_o = \partial\pi_o/\partial x \;;\; D_o = 0 \;;\; \beta\Omega_o = \Delta\pi_o, \tag{1}$$

where U_o, V_o are dimensionless velocity components along the horizontal axes x and y; D_o and Ω_o are the divergence and vorticity of the horizontal velocity; π_o is the normalized relative variation of the near-Earth pressure. The absence of the dependence of π_o, U_o, V_o, D_o and Ω_o on τ reflects the obvious fact that during the time of the run of the fast wave the background fields do not have time to change. A system of equations of the first order of the asymptotic expansion in M leads to the equation of change of the potential vorticity:

$$\partial(\Omega_1 - \beta\pi_1)/\partial\tau = -\,\partial(\Omega_o - \beta\pi_o)/\partial T - \vec{V}_o \cdot \vec{\nabla}\Omega_o \tag{2}$$

where $\vec{V}_o$ is the velocity vector of the zero approximation; the subscript "1" marks the first-order values in M. The right-hand side of equation (2) is independent of τ. Therefore, to eliminate the divergence of the asymptotic series with the increase of τ, according to the method of multiple timescale successive approximations, we assume the zero-order values to be dependent on T and equate to zero the left-and right-hand sides of (2) separately, thus obtaining

$$\partial(\Omega_o - \beta\pi_o)/\partial T = -\,\vec{V}_o \cdot \vec{\nabla}\Omega_o \;;\; \Omega_1 - \beta\pi_1 = 0 \tag{3}$$

The first, conventional equation for the evolution of the background potential vorticity, when added to (1), closes the system of equations for the background values in the quasigeostrophic approximation. It can be seen from (3) that, due to advection, the evolution of the background fields is a slower process than wave propagation and is characterized by the variable T with a scale of variation $M^{-1}T_w$ (see above). The second equation in (3) together with the other equations of the first approximation, yields a system of equations for D_1, Ω_1 and π_1 of the type:

$$\frac{\partial^2\pi_1}{\partial\tau^2} - \Delta\pi_1 + \beta^2\pi_1 = G_o \;;\; \frac{\partial^2 D_1}{\partial\tau^2} - \Delta D_1 + \beta^2 D_1 = -\,\beta\vec{V}_o \cdot \vec{\nabla}\Omega \;;\; \Omega_1 - \beta\pi_1 = 0, \tag{4}$$

where

$$G_o = (\partial U_o/\partial x)^2 + 2(\partial U_o/\partial y)(\partial V_o/\partial x) + (\partial V_o/\partial y)^2 \tag{5}$$

The two first equations in (4) are linear wave equations with inhomogeneous linear "forcing" terms on the right-hand sides which describe the generation of waves by the quasigeostrophic motions of (1) and (3). The equations in (4) present a quantitative interpretation of the qualitative theory of Obukhov-Monin (OBUKHOV, 1949; MONIN, 1969) of the generation of wave motions in the process of permanent maladjustment of the wind and pressure fields due to nonlinear effects and the tendency of the atmosphere towards reconstruction of the quasigeostrophic agreement between these fields.

The system of equations (1), (3) and (4) is simplified for small-scale motions where the Coriolis force may be neglected. In this case, the zero-order background motions are solenoidal and the wave fields are potential, the generation of waves being defined by a single value G_o, i.e. in the limit, transition to the wellknown Lighthill's theory of emission of sound by small-scale turbulence takes place.

When small-scale turbulence in the atmosphere is investigated, the assumption is usually made of the solenoidal nature of the velocity field. According to (4), at $\beta \to 0$ the solenoidal component of turbulence constantly excites the potential constituent, the main contribution to which is made by the running waves. Therefore, one can speak about "wave" (or "potential") turbulence in the atmosphere which is caused by solenoidal turbulence. The basic property of "wave" turbulence is fast energy distribution with the group velocity. Therefore, wave packets brought about by solenoidal turbulence in the tropo-stratosphere perform an efficient transfer of energy to the upper atmosphere where a layer of intensive wave turbulence is formed which is observed as irregular wave noise.

All the above also applies to larger scale motions, for which the Coriolis force is essential and $\beta \sim 1$. The statistical regime of such motions can be treated as macroturbulence. The only difference is that the geostrophic constituent with a non-zero potential vorticity rather than the solenoidal is the background component here. For the wave component, the potential vorticity equals zero.

The periods and wavelengths of the waves generated depend on the scales of the synoptic and turbulent motions. The small-scale turbulence excites acoustic waves. The meso-scale turbulence is a source of IGW, while large-scale synoptic motions generate low-frequency inertial gravity and planetary waves.

Analysis of the qualitative regular features of the behavior of the wave source G_o in (4) for mathematical models of vorticity with elliptical current lines and a rectilinear jet has shown that dynamical formations with lifetime $\tau_o \sim 1$ h produce the maximum of IGW radiation with period $\tau_w \sim 2.2$ hrs and horizontal wavelengths $\lambda_h \sim 160$-2000 km. Estimates of the vertical energy flux at heights of the order of 100 km from a rotation vortex with $\tau_o \sim 1$ h lie within $F_z \sim 1.6 \cdot 10^{-2}$ to 20 erg cm^{-2} s^{-1} for the values of $M_{max}^{o} \sim 0.01$-0.06, where M_{max} is the maximum value of the Mach number M in a dynamical formulation. Thus, the intensities of the meso-scale vortices existing in the atmosphere can provide the observed values of the energy wave flux to the upper atmosphere.

The mechanism examined of the continuous generation of waves by meteorological and turbulent motions is apparently the basic mechanism in the formation of the background level of the wave disturbance in the upper atmosphere. Other wave sources in the atmosphere (see above) can produce local and short-term enhancement of the wave noise.

Observations using state-of-the-art meteor radars can be used for a detailed study of both the interaction between IGW and the mean flux and the sources of wave motions. IGW observations by a global network of

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radiometeor stations are a pressing necessity for they can help to work out a season-latitude-longitude model of the IGW parameters and energy characteristics.

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