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COMMUNIST CHINA'S RESEARCHES IN DYNAMIC METEOROLOGY IN THE PAST DECADE

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CHINA'S RESEARCHES IN DYNAMIC METEOROLOGY IN THE PAST DECADE

[This is a full translation of an article written by Hsu Erh-hao, Department of Meteorology, University of Nanking, appearing in Ch'i-hsiang Hsueh-pao (Journal of Meteorology), Vol XXX, No 3, pages 243-250.]

Our superior Socialist system has blazed a vast, boundless path for the development of science. Before liberation, our researches in dynamic meteorology were very inadequately founded. Rapid development, however, has since come to pass.

During the past decade, our work in dynamic meteorology has been mainly devoted to atmospheric circulation. Recently, much has been accomplished in the field of numerical weather forecasting. These two phases of meteorological study will be accorded special treatment by the editorial board for meteorological reports.

To avoid repetition, this study will be a partial and not an over-all summation of dynamic meteorology. The study will be divided into in three parts: Part I dealing with the problem of aerodynamics; Part II relating to the problem of critical layer aerodynamics (including surface layer aerodynamics); Part III concerning the experience of Soviet Union's advanced meteorological sciences, etc.

I.

Theoretical research in principles and laws governing the movement of tropical cyclones constitutes an important problem for examination in China. In respect to angular research in dynamics by Yeh Tu-cheng (1950)[1], the cyclonic movement of the tropical belt was examined in the light of a straight atmospheric current; it was assumed that the tropical cyclone was of the Rankine vortex type and that the equation for its orbit was derived from the velocity of the east wind, that is, the fiducial atmospheric current.

It was shown that apart from the net westward velocity the value of V and the magnitude and orbit of the cyclone were determined by its oscillating speed and its vibrating range and cycle.

Two case histories were cited in his article to indicate clearly the nature of oscillation. In turning toward the north, the course of a tropical cyclone, when influenced by the south wind, was to be decided by the manner in which it came into contact with the stream from the south.

The author computed the course of the cyclone, as it turned, in respect to four distinct conditions. While its orbit would vary with prevailing conditions, one phenomenon was held in common--that the velocity of the cyclone tended to diminish before turning and that its movement was accelerated after turning. This phenomenon was verified by actual observation.

In discussing the movement of a vortex under baroclinic atmospheric pressure, Hsieh I-ping and Chen Ch'iu-shih (1956)[2] obtained a general equation for vortex motion which embodied various factors of vortex movement. Its significance in physics was analyzed and discussed by the authors, and its theory was concretely applied by them to the movement of typhoons in weather forecasting.

This proved that a typhoon would move in the combined direction of the isobaric surface and the average isothermal line. Thus the course of typhonic motion was interpreted in the light of proven theory, and a revision accordingly was made. By means of a simple hypothesis on isobaric distribution, the aurthors worked out an equation on a typhoon's orbit which covered cyclical oscillation and other phenomena hitherto undiscovered.

The problem of atmospheric kinetic energy plays an important role in dynamic meteorology. Huang Shih-sung (1952)[3] pointed out that in the entire atmosphere only the anticyclonic system could generate net kinetic energy, and that the system of atmospheric circulation was essentially one of net dissipation of kinetic energy. The author also pointed out that the role played by the anticyclone of the subtropical belt in the generation of atmospheric kinetic energy was extremely important.

In a detailed analysis of the problem of atmospheric kinetic energy production, Yeh Tu-cheng (1954)[4] pointed out that only through the action of a part of the fluid, compressible potential energy could kinetic energy to generated. The greater the compressibility, the greater will be the potentiality for conversion into kinetic energy. The earth's rotation reduces the possibility for the release of potential energy and its conversion into kinetic energy, for only through the earth's drift-rotation can kinetic energy be generated.

Seen from the standpoint of equation for an atmospheric kinetic energy equilibrium, and judged by its average performance, three-dimensional diffusion is accompanied by higher atmospheric pressure whereas three-dimensional convergence is followed by lower atmospheric pressure. Energy dissipated through friction is thus restored.

The author took exception to the view that the barometric maximum at the subtropical belt accounted mainly for the generation of atmospheric kinetic energy, for the anticyclonic system should not necessarily be regarded as the depository of kinetic energy.

Yeh Tu-cheng (1954) and Huang Shih-sung (1952) differed in their computations. The latter obtained his conclusion through the adoption of a static equation, while the former chose to retain dw/dt in using a straight-motion equation. Yeh Tu-cheng (1954) also made computations on the atmospheric "effective potential energy" of the Northern Hemisphere.

Regarding the problem of angular momentum equilibrium propose to conduct research on the upward pressure mechanism of angular momentum, which moves in the stratosphere from the lower to the middle-high latitudes. This momentum begins its activity at the surface layer of the easterlies belt between 30° N and 30° latitude.

Just how this momentum is lifted from a low latitude to the stratosphere is a problem to be examined. In discussing the atmospheric angular momentum equilibrium north of easterlies and westerlies line, Yeh Tu-cheng and Yang Ta-sheng (1955)[5] derived their equation for angular momentum equilibrium by the theorem of stationary conditions as follows:

$$\Omega a \int \cos^3\varphi \rho c_n d\sigma + \int \frac{\partial p}{\partial x} \cos\varphi dv + \int \tau_x \cos\varphi ds = 0,$$

In this equation are shown Q for angular velocity of the earth's rotation, τ_x for east-west directional friction stress between the earth and its atmosphere, a for the earth's radius, $d\sigma$ for the elementary capacity of the east and west winds' critical region, c_n for wind velocity vertical to $d\sigma$, and dv for ultimate volumetric capacity, and ds for the surface area's ultimate capacity.

The equation seems to indicate that in connection with the atmospheric angular momentum equilibrium north of the critical region, of the easterlies and westerlies, large-scale emission of turbulence--generally considered to be important--is rendered conspicuous by its absence.

Only the discharge of angular momentum from the earth's rotation, caused by the longitudinal circle stream (first part), is given to balance surface friction (third part), and dissipation of angular momentum by the mountain range (second part of the equation).

Near the equator, the air stream rises to the upper atmospheric westerlies belt, then flows toward the high-latitude area before falling at latitude 30°. This falling air stream reverts to the easterlies belt as it approaches to the surface.

Thus, at the meridian, an enclosed circle known as Hadley's longitudinal circle stream is completed. Because of its relation to the earth's rotation, the absolute angular momentum of the atmosphere increases on approaching the equator. With the aid of Hadley's longitudinal circle stream, the air-mass of large angular momentum is lifted up while that reverting to the equator is of small angular momentum, so that a surplus angular momentum may be discharged into the stratosphere.

The formation of atmospheric circulation with reference to its average behavior is a problem of great importance, for only through research on this important issue can we really understand the mechanism of atmospheric circulation and the factors influencing the fluctuation of atmospheric circulation. This will be of much help in weather forecasting, especially for long-range weather predictions.

In this connection, scholars abroad have conducted researches on the formation of average troughs and ridges, either through oscillation caused by topographical momentum or through turbulence caused by variation in heat field distribution.

We endeavor to point out the joint effect of topography and the heat field on this matter. The two aspects of baroclinic pressure may be represented as an equation (Chu Pao-chen, 1957)[6] for computing 500mb average oscillating height.

$$z(x, y) = \frac{f^2}{gH} \frac{3u(\bar{u} - u_T)}{3\bar{u}^2 - u_T^2} \int_0^{2\pi} \int_0^{2\pi} \eta(\alpha, \nu) \phi_0(x - \alpha, y - \nu) d\alpha d\nu +$$

$$+ MC_1 \frac{u_T}{3\bar{u}^2 - u_T^2} \int_0^{2\pi} \int_0^{2\pi} Q_m(\alpha, \nu) \phi_h(x - \alpha, y - \nu) d\alpha d\nu.$$

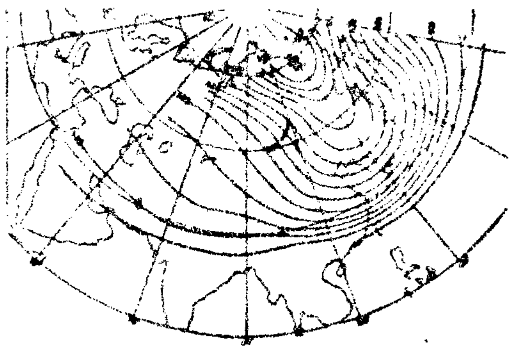


Figure 1. Heat source, heat convergence, and routine topographical turbulence in Asia at 500mb height in January.

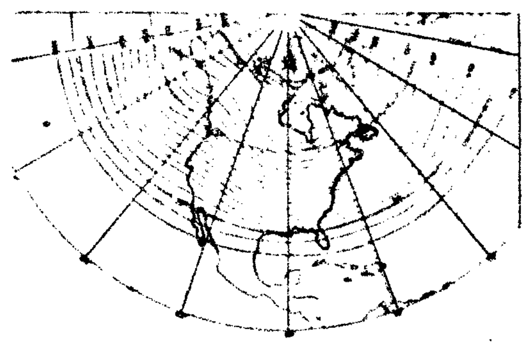


Figure 2. Heat source, heat convergence and routine Rocky and Greenland Mountains' oscillation at 500mb height in January.

With the exception of constants, the equation given above shows $\eta(a,v)$ and $Q_m(a,v)$ as functions for topography and the heat field, respectively, and ϕ_o and ϕ_h as influencing functions for topography and heat source.

Figures 1 and 2 show an average 500mb circle for January in East Asia and North America as computed from the above equation.

Figure 3. (a) 500mb height (average condition) in January in East Asia. (b) 500mb height (average condition) in January in North America.

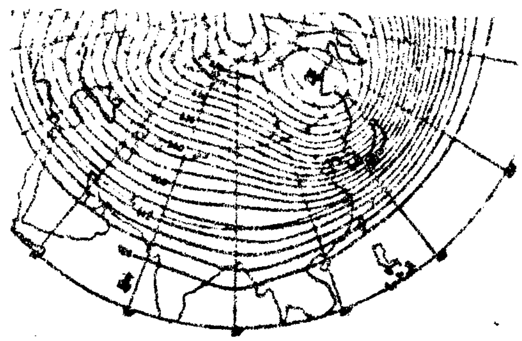


Figure (a)

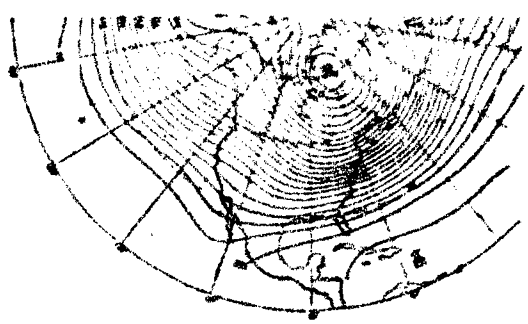


Figure (b)

The heat field distribution chart employed for theoretical computation was calculated from actual source material. Computations for 500mb average heat field distribution in the Northern Hemisphere are being worked out. The above task is founded on the theorem for mimor oscillation. In connection with three-dimensional baroclinic atmospheric pressure, topographical vibration of limited range seen in the westerlies belt may be written into the equation[7] as follows:

$$\phi'(\theta, \lambda, p) = fp_0 g\sqrt{\sigma} \sum_{n=1}^{\infty} \sum_{m=1}^{\infty} \frac{e^{-\frac{1}{2}(\xi-\xi_0)} [\psi_1(\xi_0)\psi_2(\xi) - \psi_2(\xi_0)\psi_1(\xi)]}{2k\Delta} \times$$
$$\times [\eta_n^m \cos m\lambda + \eta_n'^m \sin m\lambda] P_n^m(\cos\theta).$$

Here ϕ stands for topographical altitude;

$$\xi = \frac{2k\sqrt{\sigma}}{f\bar{\lambda}_p}\left[\bar{\lambda}_0 + \bar{\lambda}_p \frac{f}{\sqrt{\sigma}}(P_0 - P)\right], \quad k^2 = \frac{n(n+1)}{R^2}, \quad \eta_n^m, \quad \eta_n'^m$$

constitute the coefficient of topography following an expansion of the spherical function. Δ is a function of ψ_1, and ψ_2, while ψ_1 and ψ_2 represent a two-phase interpretation of an equation in advanced geometry.

$$\frac{d^2\psi}{d\xi^2} - \frac{d\psi}{d\xi} + \frac{\gamma}{\xi}\psi = 0$$

Figure 4 shows how an average 700mb circle in winter may be computed from this theorem. The distribution of principal troughs and ridges shown in Figure 4 coincides with the results of observation.

Another aspect of topographical momentum is seen in its influence on the development of troughs and ridges. In the case of straight westerlies circulation, the windward region of the Mongolian, Sinkiang, and Tibetan plateaus, respectively, finds air circulation moving up the slope, and the action is reversed in the leeward area.

It is due to this momentum that air pressure tends to increase with an upward movement and to decrease with a downward circulation.

In Figure 5 is shown the change in elevation distribution in 24 hours brought about by the effect of topographical momentum on a 500mb plane when the velocity of westerlies is held constant at five meters per second. It is apparent that there is a ppsitive change in elevation in the windward area, and that the change in the leeward region is of the negative type.

Because of this momentum factor, troughs and low pressure centers tend to weaken as they pass over the Mongolian and Tibetan plateaus while ridges and high pressure areas gain in intensity.

However, the situation is reversed on the leeward side. These phenomena have been verified by actual observation. In Figure 5 the stationary point east of the plateaus falls within an area marked by a negative change in elevation. This is due to the fact that over this area atmospheric turbulence remains feeble, and that there is little distinction in weather patterns. The discrepancy between theory and fact springs from the fact that the velocity of the west wind is theoretically held constant at five meters per second, for the sake of computation although blow over this region is unusually weak.

The influence of the Tibetan plateau on the velocity of atmospheric circulation is seen in accelerated circulation on the northern slope and retarded movement on the southern. According to a simple assumption, an equation on wave frequency may be constructed from the incline of the terrain.

$$c = U - \frac{L^2}{4\pi^2}\left(\beta + \frac{fg}{RT_h}\frac{\partial h}{\partial y}\right).$$

The equation reveals that wave velocity is $\partial h/\partial y < 0$ on the northern and $\partial h/\partial y > 0$ on the southern slope. Therefore, under similar meteorological conditions, the velocity of air circulation over high plateaus is likely to be higher on the northern than on the southern slope.

Figure 4: A 700-milli-isobaric scheme computed in consideration of comparable disturbances observed at the Tibetan Plateau and the Rocky Mountains.

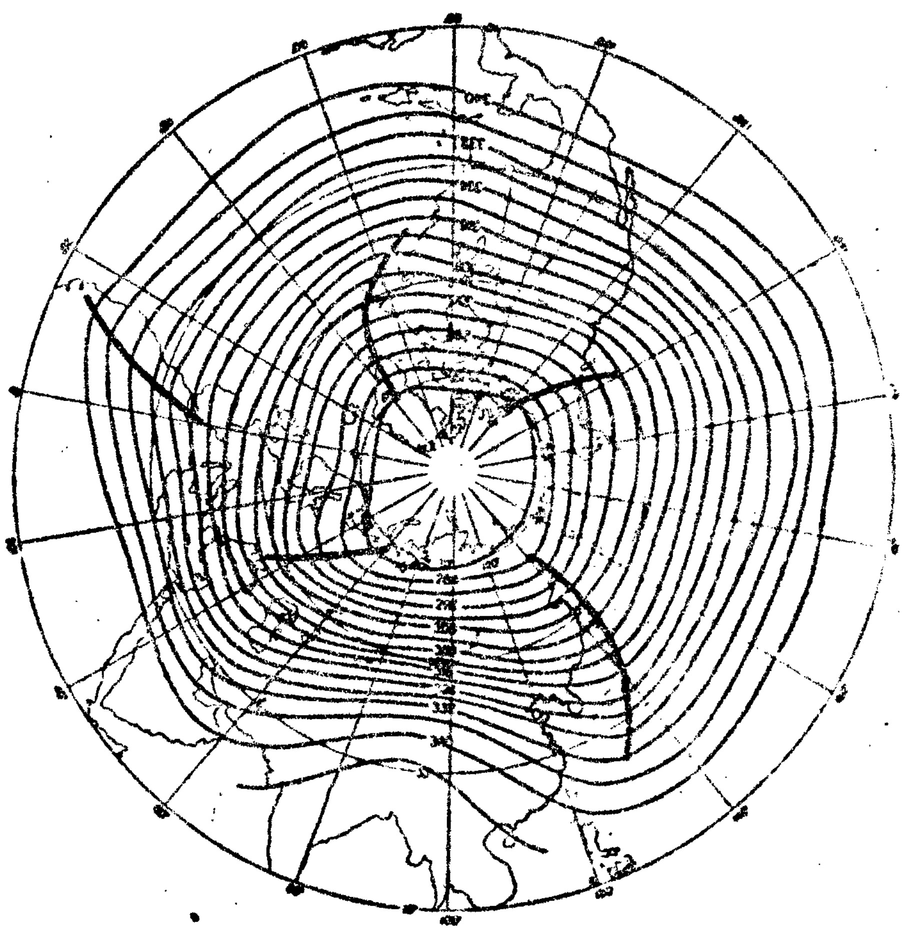
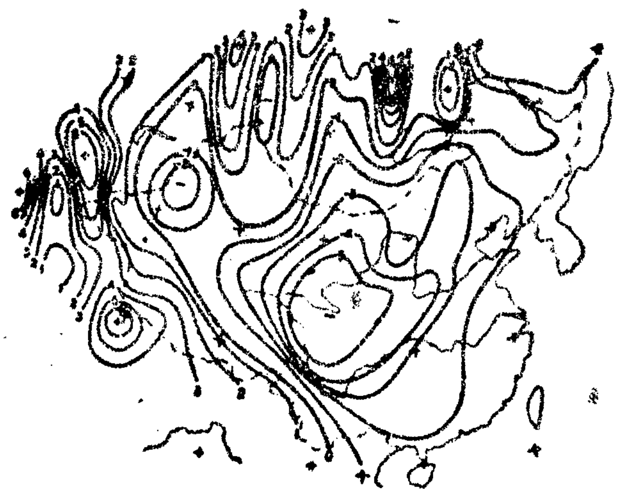


Figure 5: 500mb fluctuation in elevation caused by diffusion and convergence of an air stream as it passes over the plateau.



(Unit: 10 m per day), $w = v_o \ (\partial h / \partial y)$, $V_o = -5$m per second.

Basically, broad atmospheric movement has reference to the earth's rotation. In explanation of this phenomenon, Yeh Tu-cheng (1957)[10] stated that the basic reason was two-sided: (1) speaking of broad motion, air mass is but a thin shell; (2) production of radiation due to the difference in temperature between high and low latitudes exerts little effect on Torricelli's ratio.

The author also pointed out that an upset, for any reason of the earth's rotation equilibrium in the wind pressure field, in the case of minor air motion, should be construed as an accommodation in the air pressure field to suit the newly formed wind field, so as to achieve an equilibrium in the earth's rotation. In the case of broad air circulation, the wind field is adjusted to the newly formed air pressure field.

At times of broad disturbance, the problem of instability looms large in dynamic meteorology. In the past, researches were limited to the westerlies belts where vertical and horizontal shears had taken place. By conducting researches simultaneously on the two types of shear encountered in the westerlies belt, Chen Lung-hsun (1959)[11] provided the necessary principle on the development of unstable oscillation.

At the same time he established that in time of disturbance, due to the influence of baroclinic pressure, kinetic energy emerging from such turbulence would turn toward the kinetic energy of the fiducial air current, and that kinetic energy generated by such air disturbance would obtain its effective potential energy from the baroclinically pressured air-mass.

In damping periods, due to $\frac{dz}{dt} > 0$ (z being the absolute vortex rate), kinetic energy arising from atmospheric disturbance would still turn toward the kinetic energy of the fiducial air current.

Recently, significant results were obtained by Ch'ao Chi-ping (1959)[12] through the utilization of Lyapunov's theory for conducting researches on the problem of air mass instability.

Through an equation for atmospheric pressure tendency, fluctuations in surface atmospheric pressure may be determined by the distribution of horizontal diffusion in the upper atmosphere. For the further understanding of surface atmospheric pressure, it is necessary to analyze the cause of the appearance of horizontal diffusion.

Hsu Erh-hao (1951 a,b)[13] therefore, worked out an equation, basically similar to that of Taborovskiy, for obtaining the most common horizontal diffusion rate. This equation embodies 13 items accounting for the presence of diffusion.

Of the 13 items, nine are important and should be examined in the order of their magnitude. These items account for the production of diffusion in both horizontal and vertical movement. In these studies are included the works of J. Bjerkues, Durst-Sutcliffe, and others. The author also examined diffusion fields in the light of three simple patterns of atmospheric motion.

By using Taborovskiy's equation, Wang Shao-wu (1956)[14] computed fluctuations in surface atmospheric pressure, thereby establishing wind velocity diffusion and orbital change as two determining factors for the development of a turbulent air-mass. The manner in which the dense stratosphere in the "spearhead" belt had contributed to fluctuation in surface air pressure was computed by Chuang Ying-mo (1957)[15],and it was shown that changes in surface atmospheric pressure were in most cases due to dense stratosphere, and that air diffusion had been relegated to a place of secondary importance.

This difference in opinion proves that the mechanism of surface air pressure fluctation has not been well understood. Whether the magnitude should be computed in an equation for air pressure tendency or not is a problem yet to be solved.

In a three-dimensional analysis on an air-mass vortex field in the course of a cold wave in East Asia, Hsu Erh-hao (1958)[16] discovered that the upper atmosphere of the surface high pressure center was in no way a pure vortex area. The diffusion part of the vortex equation apparently played an important role in the development of the vortex equation, and neither a 5,500m nor a 3,000m area should be considered to approximate a non-convergence or non-divergence area.

II.

Students of meteorology in China have conducted probes and researches on the regularized activity of the lower layer jet stream of the air-mass. The works of Liu Cheng-hsing (1958)[17] and Su Tsung-hsien (1958, 1959)[18] were founded on the theorems of Obukhov-Monin, and the work of Ting Shih-ch'eng (1958)[19] was based on Holzman's theory.

These authors had compared the results of their computation with actual observations; they had found their qualitative analyses to be correct, although much remained to be desired in their quantitative analyses, largely because of the fact that they had explored only a portion of the source materials at their disposal and that little comparison had been done.

There is a difference in the jet stream exchange procedure between a neutral and a non-neutral air-mass, and that difference is to be determined by Ri (Richardson's coefficient). The jet stream exchange coefficient K for a non-neutral air-mass may be written as $K = K_o \psi (Ri)$, K_o being the exchange coefficient for a neutral air-mass.

Liu Cheng-hsing (1958) adopted the following equation in dealing with the stabilized layer of the air-mass:

$$\psi(\mathrm{Ri}) = (1 - \alpha \mathrm{Ri})^{1/2},$$

$$\varphi(\zeta) = [\sqrt{\zeta^2 + 1} - \zeta]^{-1},$$

$$\zeta = z/L, \quad \varphi(\zeta) = \frac{\kappa z}{v_*}\frac{dv}{dz} = \frac{z}{T_*}\frac{dT}{dz}.$$

Thus, the essence of meteorology is given expression to. The authors felt, however, that an unstable air-mass (troposphere) should be dealt with separately by a simplified equation for a jet energy equilibrium as follows:

$$\frac{d\bar{T}}{dz} = -\alpha' w'^2 \left(\frac{g}{T_0}\right)^{-1} L_H^{-1},$$

By using L_H as the mixture length for the troposphere, the following equation is derived:

$$\varphi_M(\zeta) = \varphi_Q(\zeta) = -A\kappa^{4/3}(\zeta)^{-1/3},$$
$$\varphi_T(\zeta) = -\alpha' A\kappa^{4/3}(\zeta)^{-1/3},$$

In the above equation, $\varphi_M(\zeta), \varphi_T(\zeta), \varphi_Q(\zeta)$ refer to reduced functions of momentum, quantity of heat, and vapor exchange coefficients. By comparing computed results on atmospheric contour activity with Obukhov-Monin's research materials, the findings, though somewhat lower, are basically uniform.

Su Tsung-hsien (1958), in working out a simplified equation for jet stream energy, estimated that in the case of a stabilized layer of air-mass the following equation could be applied:

$$\psi(\mathrm{Ri}) = (1 + \beta\mathrm{Ri})^{-1},$$
$$\varphi(\zeta) = \frac{1 + \sqrt{1 + 4\zeta}}{2}.$$

On approaching a neutral layer, $\zeta \ll 1$, $\varphi(\zeta) \simeq 1$ approximates the logarithm formula; when $\zeta \gg 1$ (stabilized period) prevails $\varphi(\zeta) \simeq \sqrt{\zeta}$,

and element, follows the square root process according to height. Therefore,by using a separate but far more complete equation, Su Tsung-hsien (1959) obtained his formula for a jet stream energy equilibrium as follows:

$$\psi(\mathrm{Ri}) = (1 - \alpha\mathrm{Ri})^{1/4};$$
$$\alpha|\zeta| = \pm\left(\varphi - \frac{1}{\varphi^3}\right).$$

In the equation, the + sign represents the stabilized layer and the - sign the unstabilized. The author stressed that the inner physical properties of the lower layer jet activity were reflected in the two equations given above.

When ζ is equal to 0 and when φ is equal to 1(at a neutral layer), K is equal to K_o. Here the meteorological situation shows a logarithmic scatter. In time of extreme instability, a great expansion in $|\zeta|$ and a great contraction in φ are indicated when $\varphi(\zeta) \simeq (a\zeta)^{-1/3}$ and $f(\zeta)$, function for wind velocity distribution, are equal to $\zeta^{-1/3} + c$. The results of computation coincide completely with comparable theories.

In the case of extreme stability, ζ expands greatly, $\zeta \simeq a\zeta$, and the value K approaches constancy, that is, the heat conduction value of air molecules. By using a = 1 for a stabilized layer and a $\simeq$ 1.5 for an unstable layer in computing the wind velocity contour, the author felt that his conclusion was in line with Obukhov-Monin's observations and research, with the exception of an upward bias in time of extreme instability and a slight downward bias in time of stability.

By adopting of Holzman's formula and by assuming that $l_s = l\sqrt{1-\epsilon Ri(z)}$, and that the distribution of potential temperature follows a logarithmic pattern, Ting Shih-ch'eng (1958) obtained a wind velocity contour formula for the lowerlayer of the atmosphere; he emphasized that on approach to a neutral layer, or under gale velocity, or at a very low height, logarithms could be used to illustrate the wind velocity contour.

Stratified atmospheric temperature is a major determining factor for the wind velocity contour at the lower layer and bias in logarithmic computation.

Wind velocity scatter is computed by the square root method when Ri is unusually large. Compared with the observational records of Best, Deacon, and others, the results of the author's computations are, qualitatively speaking, in line with their standard.

Assuming that the jet stream exchange coefficient for the air-mass critical region distribution is taken to analyze and calculate Peiping's wind velocity distribution curve, Yang Ta-sheng (1957)[20] reached the conclusion that the results were identical when actual records were compared.

$$K(z)=(\alpha-bz)^2$$

Under a similar hypothesis, Yang Ta-sheng and Li Ling-ts'un (1957)[21] analyzed and computed Hankow's wind velocity distribution; in addition, Dairen's wind velocity scatter was analyzed and calculated by Yang Ta-sheng and Cheng Kuang (1958)[22]. In both cases, the results were found to agree with the facts.

The problem of measuring the jet stream coefficient at the air-mass critical region, under nonstationary conditions was discussed by Chao Pai-ling (1956_a)[23], when wind velocity was divided into two parts:

$$u=U+u', \quad v=V+v',$$

Assuming that U and V values in the equation are satisfactory, then the equations will be:

$$-lV = -\frac{1}{\rho}\frac{\partial p}{\partial x} + k\frac{\partial^2 U}{\partial z^2},$$

$$lU = -\frac{1}{\rho}\frac{\partial p}{\partial y} + k\frac{\partial^2 V}{\partial z^2};$$

But

$$u' = -\frac{1}{l}\frac{dU}{dt}, \quad v' = \frac{1}{l}\frac{dV}{dt}.$$

After obtaining values for U and V, u' and v' may be solved by using t differential. By adding the results obtained, wind velocity scatter at the air-mass critical region under nonstationary conditions may be computed. From this a formula for obtaining K, the jet stream coefficient, may be written.

Ekman's and Laykhtman's formula is used to illustrate specifically the method of obtaining the K value. In comparing his K value with that of Laykhtman, the author noted that much discrepancy--the difference was more than two times--had existed in the case of individual analysis, but the results were more or less similar when averages were taken.

By adopting K formulas for computing the jet stream coefficient as adopted by Obukhov and Budyko, and by computing Peiping records, Chao Pai-ling (1956_b)[24] came to the conclusion that the scope of applying Obukhov's formula was narow and that of using Budyko's formula was much broader.

III.

The Kibel' theory on forecasting methods is regarded as one of the earliest and most effective methods in recent developments of numerical weather forecasting; it plays an important role in the history of meteorology.

Since liberation, his theory has elicited broad, enthusiastic interest among meteorological workers in China. In this connection, an over-all interpretation had been provided by Yeh Tu-cheng (1954)[26].

While emphasizing his many important accomplishments, it was simultaneously pointed out that his theory should be applied to conditions in China such as the problem of topography, the less probable fluctuations in meteorological factors in China than in the Soviet Union, the effect of latitudinal changes on the earth's rotation f, etc.

The effect of latitudinal changes on the earth's rotation f, with reference to the Kibel' theory, was discussed in great detail by Chu Pao-chen.

In analyzing (1) the approximate value of temperature advectional fluctuation and of atmospheric pressure advectional fluctuation, and in demonstrating (2) fluctuations in temperature dynamics, he asserted that there was no effect whatever of altitudinal changes on f value in the advection of temperature and dynamic fluctuation, or fluctuations in the advection of atmospheric pressure; that is to say, the effect of f on y (northward) was inconsequential.

But in indicating the dynamic fluctuation of atmospheric pressure when the effect of f value on latitudinal changes was considered, he pointed out that a few items, including $\partial f/\partial y$, should be added to the formula, as had been originally propounded by Kibel'.

The formula involving $\partial f/\partial y$ should be compared with one including a divergence item ($\partial^2\theta/\partial, \partial\theta$), in order to assess their relative magnitudes.

Generally speaking, the difference would be regarded as immaterial if the former should prove to be 20 percent of the latter. It was, however, established that the greater the gradient of surface atmospheric pressure, the greater would be the effect noted in the formula involving $\partial f/\partial y$. Therefore, during the development of low pressure and the expansion of the gradient of atmospheric pressure, the importance of a formula involving $\partial f/\partial y$ should be noted.

By using a formula in which heat was given to account for the formation of wind, Hsu Erh-hao (1954)[27] re-established some of the basic equations of Kibel' which were given physical interpretation.

He pointed out that changes in advective pressure above various heights and fluctuations in surface advective pressure were of the same type below the principal (tao) layer; but that their classifications altered as they moved above the principal layer. The absolute value of advective pressure directly increased with its distance from the principal layer. Irrespective of height, the dynamic change in pressure was grouped together with surface pressure, though its absolute total value decreased as height increased.

The author also pointed out that the movement of the temperature pressure system at all heights above a convective layer was controlled by a single principal layer. In other words, uniformity was noted in the rate of motion for the temperature pressure system at all heights. Ku Cheng-ch'ao (1955) [28] attempted to broaden the Kibel' f equations into three types.

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