2k. Meteorological Information[1]

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2k-1. List of Symbols

c_p	specific heat of dry air at constant pressure
c_v	specific heat of dry air at constant volume
d	coefficient of molecular diffusion
D	coefficient of eddy diffusion
f	Coriolis parameter
g	acceleration of gravity
n	distance; spacing
p	pressure
Q	source strength; i.e., total amount of material released from a source
r	radius of curvature
R	gas constant for dry air
T	temperature
T_v	virtual temperature
T_{mv}	mean virtual temperature
U	west wind speed
v	speed
V	gradient wind speed
V_g	geostrophic wind speed
w	water-vapor mixing ratio
Z, z	height
β	northward variation of Coriolis parameter
γ	lapse rate of temperature
ζ	mean molecular speed; mixing velocity
λ	free path; mixing length; wavelength
ρ	density of air
σ	standard deviation
ϕ	latitude
Φ	geopotential
χ	concentration
ω	angular velocity of the earth

[1] All material not otherwise credited is abstracted either from List [23] or from Letestu [22]. These publications should be consulted for more complete explanations and additional references. For an encyclopedic summary of the status of knowledge (1950) in the principal fields of meteorology and atmospheric physics, including extensive references, see Malone [24]. More recent general references include [20], [1], [4], and [3]. For special annotated bibliographies see Rigby [30].

2k-2. Physical Constants

Pressure at mean sea level, 1 atm

$= 1{,}013.250$ millibars (mb) $= 1.013250 \times 10^6$ dynes cm^{-2}

$= 76$ cm Hg (at standard gravity of 980.665 cm sec^{-2} and temperature of 0°C)

Mass of the atmosphere $= 5.14 \times 10^{21}$ g

Apparent molecular weight of dry air $M = 28.9644$[1]

Gas constant for 1 kg of dry air, $R = 287.05\ J\ \mathrm{kg^{-1}\ K^{-1}} = 6.8607 \times 10^{-2}$ cal* $\mathrm{g^{-1}\ K^{-1}}$

Specific heat of dry air at constant pressure $c_p = \dfrac{7R}{2} = 0.2401$ cal $\mathrm{g^{-1}\ K^{-1}}$

Specific heat of dry air at constant volume $c_v = \dfrac{5R}{2} = 0.1715$ cal $\mathrm{g^{-1}\ K^{-1}}$

2k-3. Composition of the Atmosphere

TABLE 2k-1. NORMAL COMPOSITION OF CLEAN, DRY ATMOSPHERE AIR NEAR SEA LEVEL*

Constituent gas and formula	Content, % by volume	Content variable relative to its normal	Molecular weight
Nitrogen (N_2)	78.084	–	28.0134
Oxygen (O_2)	20.9476	–	31.9988
Argon (Ar)	0.934	–	39.948
Carbon dioxide (CO_2)	0.0314	†	44.00995
Neon (Ne)	0.001818	–	20.183
Helium (He)	0.000524	–	4.0026
Krypton (Kr)	0.000114	–	83.80
Xenon (Xe)	0.0000087	–	131.30
Hydrogen (H_2)	0.00005	?	2.01594
Methane (CH_4)	0.0002	†	16.04303
Nitrous oxide (N_2O)	0.00005	–	44.0128
Ozone (O_3)	Summer: 0 to 0.000007	†	47.9982
	Winter: 0 to 0.000002	†	47.9982
Sulfur dioxide (SO_2)	0 to 0.0001	†	64.0628
Nitrogen dioxide (NO_2)	0 to 0.000002	†	46.0055
Ammonia (NH_3)	0 to trace	†	17.03061
Carbon monoxide (CO)	0 to trace	†	28.01055
Iodine (I_2)	0 to 0.000001	†	253.8088

* From "U.S. Standard Atmosphere, 1962" [34]; see also Glueckauf [16] and Keeling [19].

† The content of the gases marked with a dagger may undergo significant variations from time to time or from place to place *relative* to the normal indicated for those gases; for example, O_3 has a maximum concentration in the stratosphere.

2k-4. Geopotential. The geopotential Φ of a point at a height z above mean sea level is the work which must be done against gravity in raising a unit mass from sea level to height z.

$$\Phi = \int_0^z g\, dz \tag{2k-1}$$

where g is the local acceleration of gravity at height z. For most meteorological work geopotential is measured in terms of the *geopotential meter* (gpm). By definition, 1 gpm $= 9.8 \times 10^4$ cm^{2}sec^{-2}. In the lower atmosphere, 1 gpm $\simeq$ 1 geometric meter.

[1] Carbon-12 isotope scale for which $C^{12} = 12$.

* Thermochemical calorie.

Table 2k-2 shows the relationship between geopotential and geometric height as a function of latitude.

2k-5. Hypsometry. The differential form of the hydrostatic equation, the equation expressing the relationship of pressure p, density ρ, and height z in the atmosphere, is

$$dp = -\rho g\, dz \tag{2k-2}$$

Introducing the definition of geopotential, the hydrostatic equation becomes

$$dp = -\rho\, d\Phi \tag{2k-3}$$

Substituting the equation of state for dry air, introducing the concept of virtual temperature,[1] and integrating, Eq. (2k-3) becomes

$$\Delta\Phi = RT_{mv} \log_e \frac{p_1}{p_2} \tag{2k-4}$$

where $\Delta\Phi$ is the geopotential difference between levels having pressures p_1 and p_2, T_{mv} is the mean virtual temperature of the layer of air between p_1 and p_2, and R is

TABLE 2k-2. RELATION OF GEOPOTENTIAL TO GEOMETRIC HEIGHT

Latitude	Geopotential meters (gpm)										
	10,000	20,000	30,000	40,000	50,000	100,000	200,000	300,000	400,000	500,000	600,000
	m	m	m	m	m	m	m	m	m	m	m
0°	10,036	20,104	30,204	40,336	50,500	101,811	206,948	315,577	427,874	544,029	664,243
30°	10,023	20,077	30,163	40,282	50,432	101,672	206,656	315,115	427,225	543,174	663,161
45°	10,009	20,050	30,123	40,228	50,365	101,534	206,363	314,653	426,576	542,318	662,080
60°	9,996	20,024	30,083	40,174	50,297	101,395	206,071	314,191	425,927	541,465	661,000
90°	9,983	19,997	30,043	40,120	50,229	101,256	205,779	313,730	425,280	540,613	659,923

the gas constant for dry air. For temperatures in K and geopotential in gpm (i.e., geometric meters for most practical purposes) Eq. (2k-4) becomes

$$\Delta\Phi \doteq 67.445 T_{mv} \log_{10} \frac{p_1}{p_2} \tag{2k-5}$$

2k-6. Lapse Rates. The lapse rate γ in the atmosphere is defined as the rate of decrease of temperature with increasing height (or geopotential), $\gamma = -dT/dz$. γ is ordinarily expressed in °C per 100 m (or 100 gpm).

Dry-adiabatic Lapse Rate. Dry air, or moist air in which the water vapor enters into no change of state, which ascends (or descends) adiabatically in the atmosphere will decrease (or increase) in temperature at the rate of g/c_p. The dry adiabatic lapse rate is therefore 0.98°C/100 m.

[1] The virtual temperature T_v is defined to be the temperature that dry air would have at given pressure in order to have the same density as moist air at the same pressure but at temperature T and with specified moisture content. Approximately, $T_v = T(1 + 0.61w)$, where w is the water-vapor mixing ratio, or the ratio of water vapor to the dry air. The logarithmic mean virtual temperature is required in Eq. (2k-4).

Pseudoadiabatic Lapse Rate. Saturated air ascending adiabatically in the atmosphere, so that all condensation of water vapor is into liquid water which falls out immediately and all latent heat of condensation is realized in warming the air, decreases in temperature at the pseudoadiabatic (or moist-adiabatic) lapse rate for the water stage. Table 2k-3 gives the pseudoadiabatic lapse rate for the water stage as a function of temperature and pressure. The pseudoadiabat c lapse rate for the ice (snow) stage is different from that for the water stage.

2k-7. U.S. Standard Atmosphere. A revised standard atmosphere for levels up to 700 km (Table 2k-4) was adopted by the United States Committee on Extension to the Standard Atmosphere on March 15, 1962 [34]. The lower 32 km of this representation of the atmosphere was approved by the International Civil Aviation Organization on November 12, 1963, for international standardization. The values represent idealized, middle-latitude, year-round conditions for the range of solar activity between sunspot minimum and sunspot maximum. For representative values at the extremes of the solar cycle and at various latitudes and seasons, see [33] and [34].

TABLE 2k-3. PSEUDOADIABATIC LAPSE RATE FOR THE WATER STAGE, °C/100 GPM

Temp., °C	Pressure, mb									
	1000	900	800	700	600	500	400	300	200	100
−50	0.966	0.965	0.963	0.961	0.959	0.955	0.951	0.943	0.928	0.886
−40	0.950	0.947	0.944	0.939	0.934	0.925	0.913	0.896	0.863	0.775
−30	0.917	0.910	0.903	0.893	0.882	0.866	0.842	0.807	0.746	0.615
−20	0.855	0.844	0.830	0.814	0.794	0.767	0.730	0.677	0.596	0.454
−10	0.763	0.746	0.725	0.701	0.672	0.637	0.592	0.532	0.452	0.335
0	0.645	0.624	0.601	0.573	0.542	0.505	0.462	0.409	0.345	0.262
10	0.527	0.506	0.483	0.457	0.429	0.398	0.362	0.323	0.276	
20	0.426	0.408	0.389	0.368	0.346	0.322	0.296			
30	0.352	0.338	0.323	0.307	0.291	0.273				
40	0.301	0.290	0.279	0.267						
50	0.267	0.259								

The supplemental atmospheres in [33] conform closely to the Cospar International Reference Atmosphere [6].

Basic Assumptions. It is assumed that the air is dry, obeys the perfect-gas law, and is in hydrostatic equilibrium. (The other assumptions and necessary physical constants used are given in [34].)

Latitudinal Temperature Distribution. The average temperature in January and July in the Northern Hemisphere in the lowest 80 km of the atmosphere as a function of latitude and height is shown in Fig. 2k-1 taken from [33]. The heavy lines show the *mean* height of the tropopause. There may be more than one tropopause over a given point during certain meteorological conditions; during others it may be indistinct or missing altogether. The bold vertical lines indicate isothermal conditions.

2k-8. Other Properties and Phenomena of the Atmosphere. Space precludes the presentation of numerous meteorological data on the upper atmosphere recently obtained by satellite exploration [36,15,28,18]. However, schematic representations of the structure of the upper atmosphere, indicating the typical heights at which various phenomena have been observed, are shown in Fig. 2k-2. Figure 2k-3, after Hanson in [18], shows the dependence of electron concentration on night and day and on the 11-year sunspot cycle.

TABLE 2k-4. U.S. STANDARD ATMOSPHERE*

Altitude, km	Temperature, K	Pressure, dynes/cm²	Density, g/cm³	Viscosity, poises	Speed of sound, m/sec	Molecular weight
−5	320.68	1.7776 + 6	1.9311 − 3	1.9422 − 4	358.986	28.964
−4	314.17	1.5960	1.7697	1.9123	355.324	28.964
−3	307.66	1.4297	1.6189	1.8820	351.625	28.964
−2	301.15	1.2778	1.4782	1.8515	347.888	28.964
−1	294.65	1.1393	1.3470	1.8206	344.111	28.964
0	288.15	1.0132	1.2250	1.7894	340.294	28.964
1	281.65	8.9876 + 5	1.1117	1.7579	336.435	28.964
2	275.15	7.9501	1.0066	1.7260	332.532	28.964
3	268.66	7.0121	9.0925 − 4	1.6938	328.583	28.964
4	262.17	6.1660	8.1935	1.6612	324.589	28.964
5	255.68	5.4048	7.3643	1.6282	320.545	28.964
6	249.19	4.7218	6.6011	1.5949	316.452	28.964
7	242.70	4.1105	5.9002	1.5612	312.306	28.964
8	236.22	3.5652	5.2579	1.5271	308.105	28.964
9	229.73	3.0801	4.6706	1.4926	303.848	28.964
10	223.25	2.6500	4.1351	1.4577	299.532	28.964
11	216.77	2.2700	3.6480	1.4223	295.154	28.964
12	216.65	1.9399	3.1194	1.4216	295.069	28.964
13	216.65	1.6580	2.6660	1.4216	295.069	28.964
14	216.65	1.4170	2.2786	1.4216	295.069	28.964
15	216.65	1.2112	1.9475	1.4216	295.069	28.964
16	216.65	1.0353	1.6647	1.4216	295.069	28.964
17	216.65	8.8497 + 4	1.4230	1.4216	295.069	28.964
18	216.65	7.5652	1.2165	1.4216	295.069	28.964
19	216.65	6.4675	1.0400	1.4216	295.069	28.964
20	216.65	5.5293	8.8910 − 5	1.4216	295.069	28.964
25	221.55	2.5492	4.0084	1.4484	298.389	28.964
30	226.51	1.1970	1.8410	1.4753	301.709	28.964
40	250.35	2.8714 + 3	3.9957 − 6	1.6009	317.189	28.964
50	270.65	7.9779 + 2	1.0269	1.7037	329.799	28.964
60	255.77	2.2461	3.0592 − 7	1.6287	320.606	28.964
70	219.70	5.5205 + 1	8.7535 − 8	1.4383	297.139	28.964
80	180.65	1.0366	1.9990	1.216	269.44	28.964
100	210.02	3.0075 − 1	4.974 − 10	†	†	28.88
150	892.79	5.0617 − 3	1.836 − 12			26.92
200	1235.95	1.3339	3.318 − 13			25.56
250	1357.28	4.6706 − 4	9.978 − 14			24.11
300	1432.11	1.8838	3.585			22.66
400	1487.38	4.0304 − 5	6.498 − 15			19.94
500	1499.22	1.0957	1.577			17.94
600	1506.13	3.4502 − 6	4.640 − 16			16.84
700	1507.61	1.1918	1.537			16.17

* From "U.S. Standard Atmosphere, 1962," NASA, USAF, USWB, Washington, D.C., December, 1962.

† Equations used to compute viscosity and speed of sound not applicable above 90 km.

Note. A one- or two-digit number (preceded by a plus or minus sign) following the initial entry indicates the power of 10 by which that entry and each succeeding entry of that column should be multiplied.

2k-9. Dynamical Relationships. *Coriolis Parameter.* The *apparent* force per unit mass acting upon a particle whose motion is described in a coordinate system fixed to the surface of the earth, due to the rotation of the earth in space, is proportional to the Coriolis parameter $f = 2\omega \sin \phi$, where ω is the angular velocity of the earth ($\omega = 7.292 \times 10^{-5}$ radian sec^{-1}) and ϕ is latitude. If v is the speed of a particle of unit mass, the apparent force is equal to fv. This force is directed to the right of the direction of motion in the Northern Hemisphere and to the left in the Southern Hemisphere.

Geostrophic Wind. Steady, straight, frictionless air motion in an unchanging pressure field, with gravity as the only external force acting, so that the horizontal pressure-gradient force is balanced by the apparent force due to the earth's rotation (the

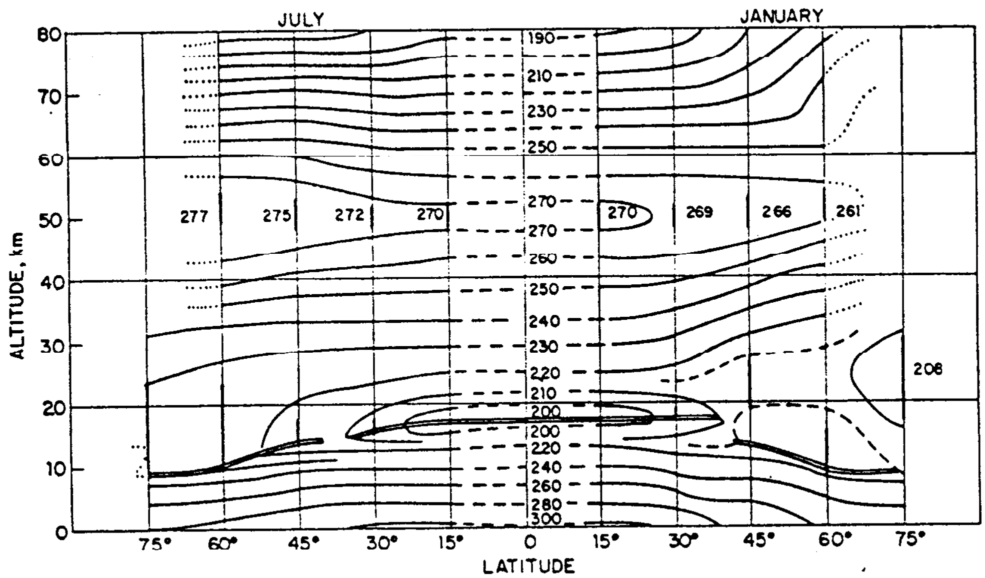


FIG. 2k-1. Temperature-altitude cross section for January and July. ("*U.S. Standard Atmosphere Supplements, 1966*," ref. 33.)

Coriolis force), is called the *geostrophic* wind. The geostrophic wind blows perpendicularly to the direction of the pressure gradient with low pressure to the left in the Northern Hemisphere, to the right in the Southern Hemisphere.

TABLE 2k-5. VALUE OF THE CORIOLIS PARAMETER f

Latitude	f, sec^{-1}	Latitude	f, sec^{-1}
0°	0	50°	1.1172×10^{-4}
10°	0.2533×10^{-4}	60°	1.2630×10^{-4}
20°	0.4988×10^{-4}	70°	1.3705×10^{-4}
30°	0.7292×10^{-4}	80°	1.4363×10^{-4}
40°	0.9375×10^{-4}	90°	1.4584×10^{-4}
45°	1.0313×10^{-4}		

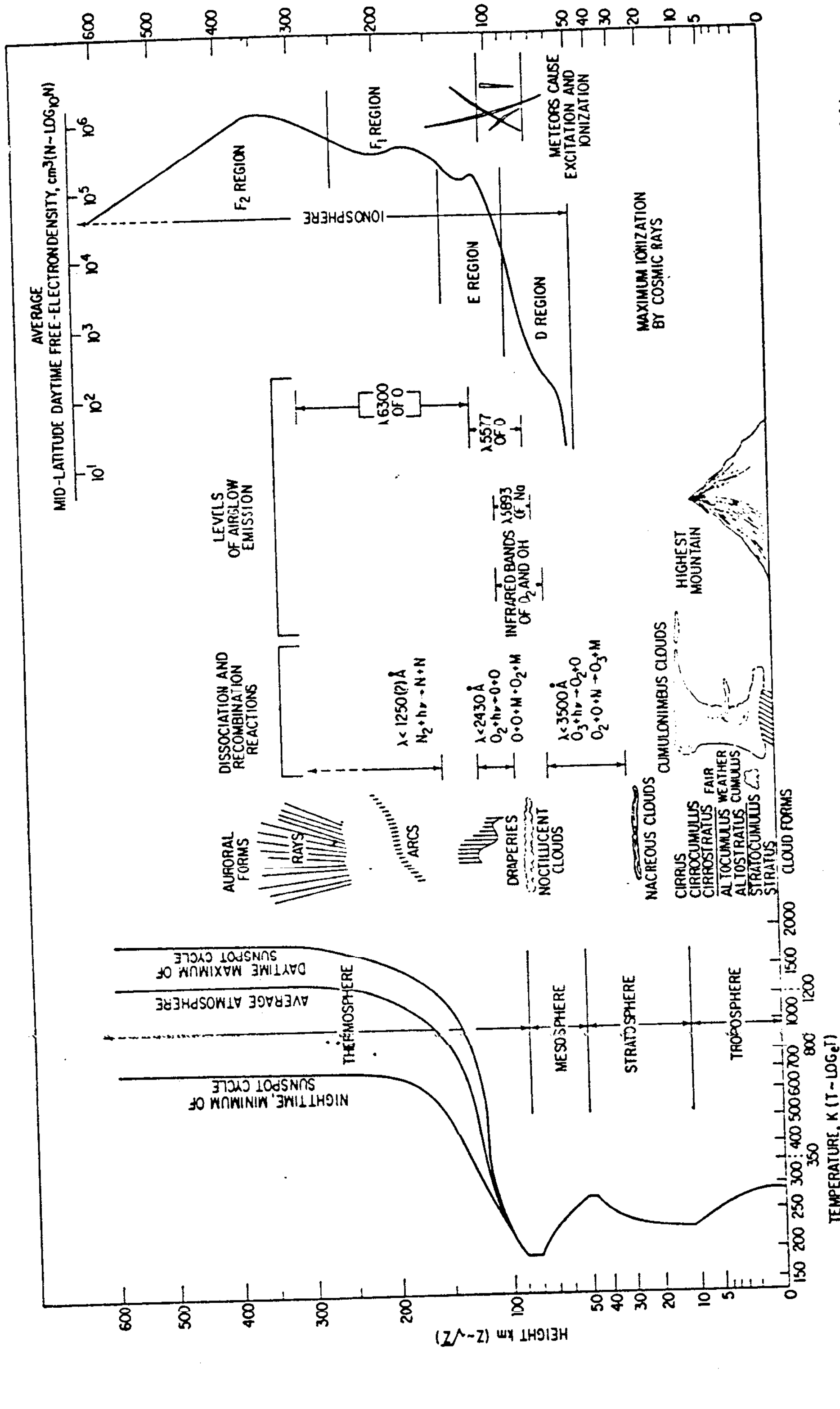


FIG. 2k-2. Structure of the upper atmosphere. (*Prepared in collaboration with W. W. Kellogg and A. Kochanski.*)

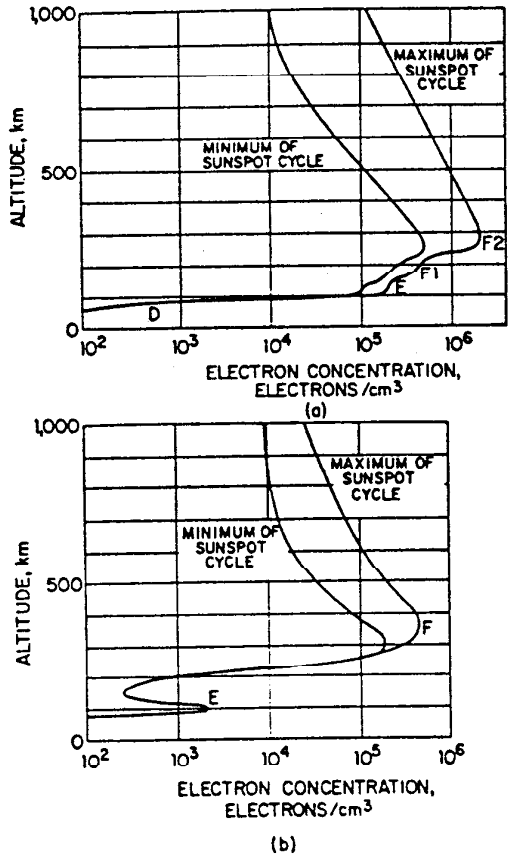


FIG. 2k-3. Diurnal and solar-cycle variations in the structure of the ionosphere. (*a*) Daytime. (*b*) Nighttime. (*After Henson*, ref. 18.)

On a surface of constant pressure, the equation for the speed of the geostrophic wind V_g is given by

$$V_g = \frac{1}{f}\frac{\partial \Phi}{\partial n} \tag{2k-6}$$

where $-\partial\Phi/\partial n$ is the gradient of geopotential on the constant-pressure surface normal to the direction of the geostrophic wind. On a constant-level surface,

$$V_g = \frac{1}{f\rho}\frac{\partial p}{\partial n} \tag{2k-7}$$

where ρ is the density of the air and $-\partial p/\partial n$ is the horizontal pressure gradient normal to the geostrophic wind component.

Gradient Wind. To improve the approximation of the geostrophic wind to the true wind in the free atmosphere, other terms may be included in the equation of motion. The most common additional term is that which expresses the acceleration arising from the curvature of the path of the moving air parcel. The addition of this term to the expression for the geostrophic wind speed gives the *gradient* wind speed V.

$$V = \frac{rf}{2}\left[-1 + \left(1 + \frac{4V_g}{rf}\right)^{1/2}\right] \tag{2k-8}$$

where r is the radius of curvature of the trajectory of the air parcel and the following sign convention is used: for cyclonic curvature $rf > 0$, for anticyclonic curvature $rf < 0$.

Zonal Motion. The average motion of the atmosphere is predominantly geostrophic and zonal. The zonal motion between sea level and 50 mb, for summer and winter, Northern and Southern Hemispheres, is shown in Fig. 2k-4, from Mintz [25].

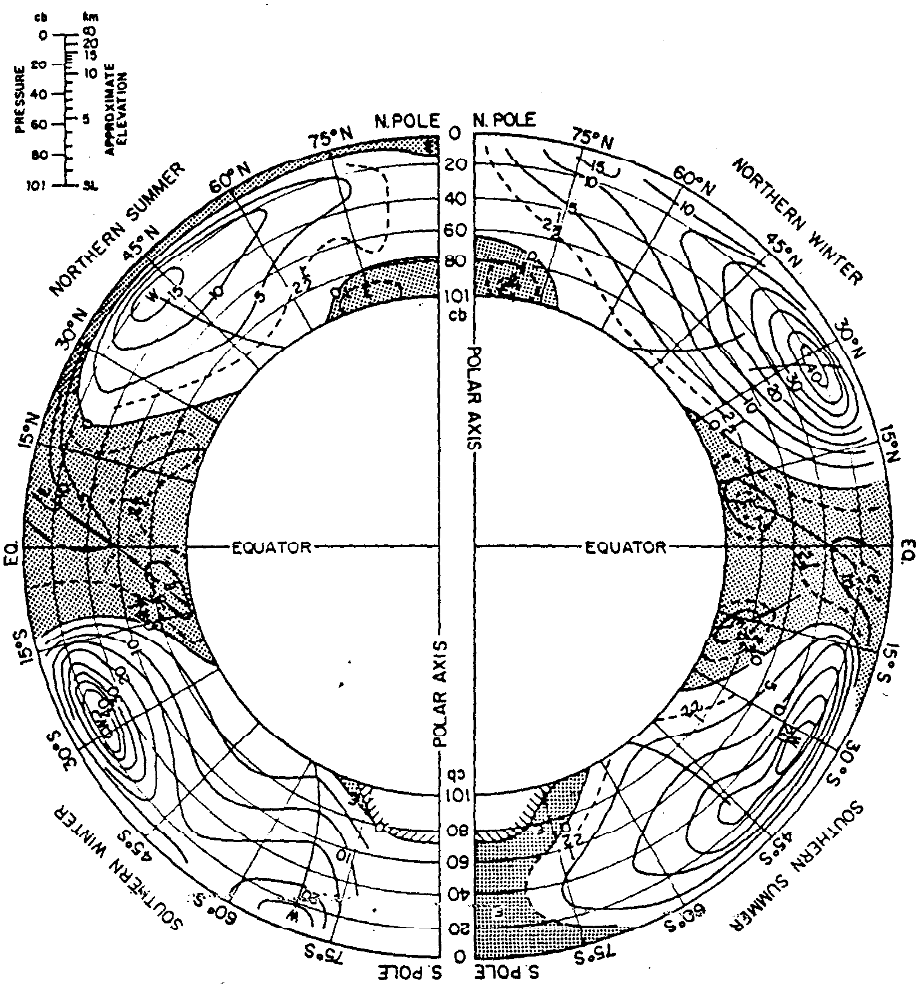


FIG. 2k-4. Zonal circulation of the atmosphere, m sec^{-1}, averaged over all longitudes. W represents motion from the west, E is motion from the east. (*From Mintz*, ref. 25.)

For levels above 35 km, see Murgatroyd in ref. [2]. A pronounced 26-month oscillation of the zonal wind in the equatorial stratosphere has been observed (see, e.g., Reed [29]).

Eddy Motion. Superimposed on the average zonal motion of the atmosphere are eddy circulations covering a wide spectrum, including cyclones and anticyclones in the lower troposphere and planetary or Rossby waves in the middle troposphere. Under *barotropic* conditions frequently observed in the middle troposphere, the speed c of planetary waves is given by

$$c = U - \frac{\beta\lambda^2}{4\pi^2} \tag{2k-9}$$

where U is the west wind speed, β the northward change of the Coriolis parameter, and λ the wavelength. For an introduction to current numerical techniques of modeling and predicting atmospheric processes, especially atmospheric motions, see Thompson [31].

Energy Conversions. Figure 2k-5, from Oort [26], shows an estimate of the generation G, dissipation D, and conversion C rates for energy processes in the atmosphere. In the average, the energy cycle proceeds from mean available potential energy P_M via eddy available potential energy P_E and eddy kinetic energy K_E to the mean kinetic energy (K_M).

2k-10. Radiation. *Solar Constant.* The solar constant, the mean value of the total solar radiation, at normal incidence, outside the atmosphere at the mean solar distance = 0.140 w cm^{-2} (p.e. = 2%) [18].

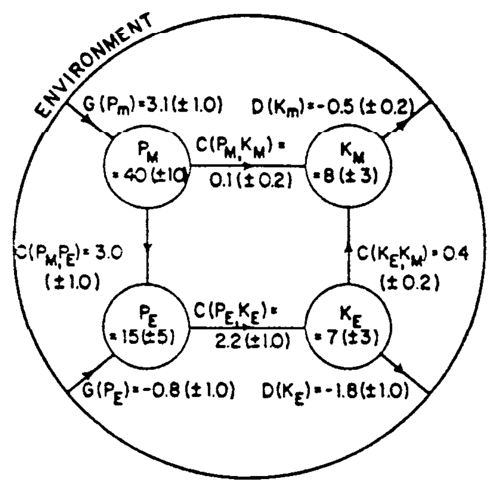


FIG. 2k-5. Tentative flow diagram of the atmospheric energy in the space domain. Values are averages over a year for the Northern Hemisphere. Energy units are in 10^5 joules m^{-2} (= 10^8 ergs cm^{-2}); energy transformation units are in watts m^{-2} (= 10^3 ergs cm^{-2} sec^{-1}). (*From Oort*, ref. 26.)

Insolation. Figure 2k-6 shows the average daily solar radiation received on a square centimeter of horizontal surface at the ground during January and July on cloudless days [11] (solid lines) and on days with average cloudiness [13] (dotted lines). The units are gram-calories per square centimeter per day.

Albedo. Table 2k-6 gives a range of albedo measurements[1] observed for various type of surface.

Heat Balance of the Atmosphere. Taking the incident solar radiation as 100 units, Byers [5] has computed the heat budget of the atmosphere as shown in Table 2k-8.

2k-11. Clouds.[2] The drop-size spectra of typical cloud types are given in Fig. 2k-7.

2k-12. Climatology. Space limitations preclude the presentation of climatological data. In addition to standard climatological texts, see [17], [32], [7], [8], [9], and [35]; the reports of World Data Center A, especially the subcenters on Meteorology, Upper Atmosphere Geophysics, and Rockets and Satellites; and various numbers in the key to Meteorological Records Documentation series, especially No. 4.11 [10].

[1] For a more complete list, including sources, see List, *op. cit.*, pp. 442–444.

[2] Data furnished by Dr. H. J. aufm Kampe, Signal Corps Engineering Laboratories, Ft. Monmouth, N.J.

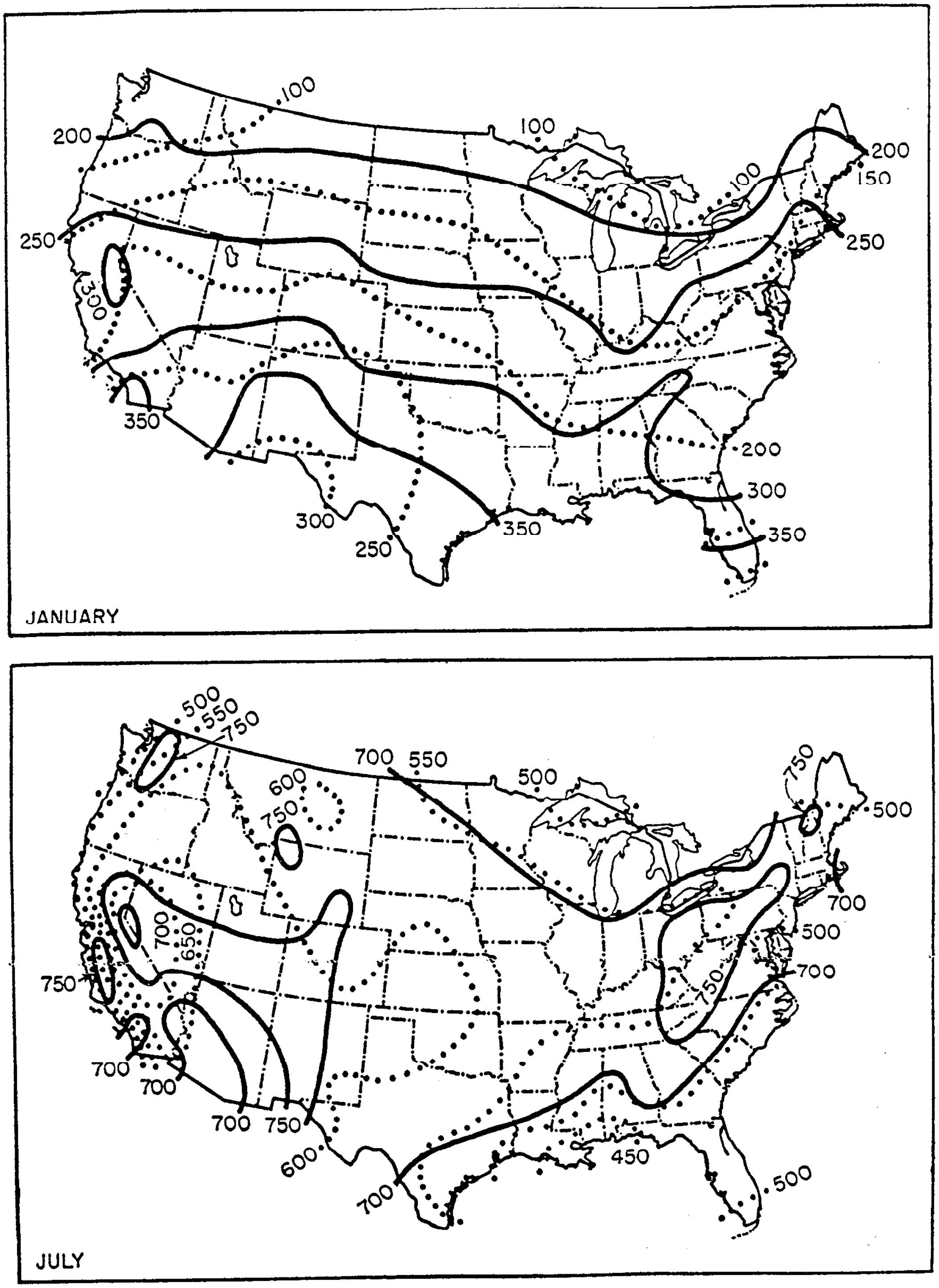


FIG. 2k-6. Average daily solar insolation (g-cal cm^{-2} day^{-1}) at the ground on cloudless days (solid lines) and on days of average cloudiness (dotted lines). (*After Fritz and MacDonald* [11, 13].)

TABLE 2k-6. ALBEDO MEASUREMENTS

	%
Forest	3–10
Fields, grass, etc	3–37
Bare ground	3–30
Snow, fresh	80–90
Snow, old	45–70
Whole earth, visible spectrum	39
Whole earth, total spectrum	35
Clouds*	5–85
Water (reflectivity values are given in the following table)†	

Elevation of sun	90°	70°	50°	40°	30°	20°	10°	5°	0°
Reflectivity, %	2.0	2.1	2.5	3.4	6.0	13.4	34.8	58.4	100.0

* For clouds in the absence of absorption the albedo is a function of the drop-size distribution, liquid water content, and cloud thickness. See S. Fritz [12].

† The reflectivity of a water surface for solar radiation is a function of the sun's elevation angle. The values given have been computed for a plane surface; however, the observed reflection from disturbed surfaces shows only small deviation from these values.

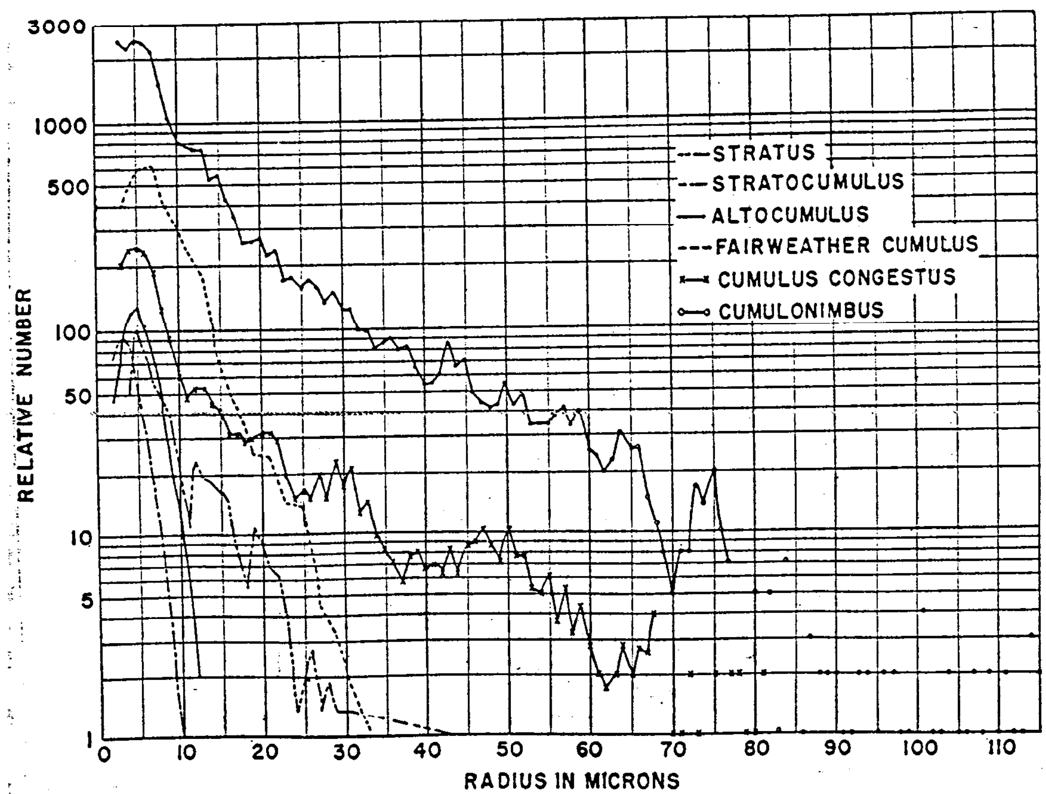
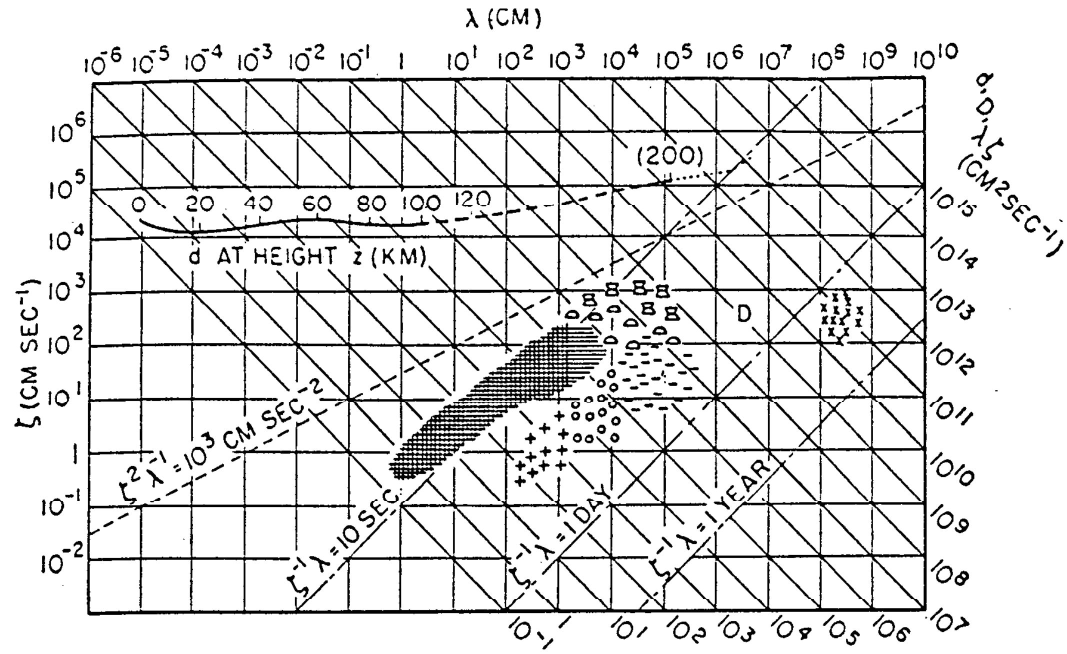


FIG. 2k-7. Cloud drop-size spectra. (*Prepared by H. J. Aufm Kampe.*)

2k-13. Atmospheric Diffusion.[1] In most meteorological problems, the effects of molecular diffusion are far outweighed by the turbulent eddies present in the atmosphere. One approach to this problem is to treat the phenomenon in a manner analogous to that of molecular diffusion. The coefficient of diffusion in such applications is a function of the size of the turbulent eddies and is therefore dependent on the

[1] For the definition of diffusion coefficient cf. p. 2-221.



D AT HEIGHT z IS DENOTED BY CHARACTERISTIC AREAS ON THE DIFFUSION DIAGRAM:

- 1-10 KM FOR ORDINARY TURBULENCE — 0-1 KM
- 1-10 KM FOR CUMULUS CONVECTION — 25-35 KM
- 1-10 KM FOR CUMULONIMBUS CONVECTION — 45-80 KM
- 10-25, 35-45 AND 80-100 KM
- HORIZONTAL GROSS-AUSTAUSCH OF THE GENERAL CIRCULATION

FIG. 2k-8. Diffusion diagram. (*From Lettau.*)

TABLE 2k-7. AVERAGE WATER CONTENT OF TYPICAL CLOUDS

Cloud type	Cirrus*	Cumulus congestus	Fair-weather cumulus	Stratus	Strato-cumulus
Water content, g m^{-3}	0.01–0.05	4	1	0.3	0.2

* Cirrus clouds consist mainly of column-shaped ice crystals. In cirrostratus, single, more or less completely built columns (twin crystals) of about 100 microns in length and 25 microns in diameter predominate. In dense cirrus and cirrocumulus clouds the columns are incompletely built and occur in clusters. The length of the individual crystals in such clusters is approximately 100 to 300 microns and the diameter 30 to 100 microns.

TABLE 2k-8. HEAT BUDGET OF THE ATMOSPHERE*

Absorbed	Units	Lost	Units
Solar radiation	13	Infrared radiation to earth's surface	106
Latent heat	23	Infrared radiation to space	47
Infrared radiation from earth's surface	120	Eddy transfer to ground	3

* After Byers [5].

time and space scale. Figure 2k-8 (see Lettau [21]) gives the magnitude of the coefficient of eddy diffusion D as a function of the characteristics of the eddies, as well as the variation of the coefficient of molecular diffusion d with height. In Fig. 2k-8, each point of the λ, ζ plane determines a diffusion coefficient (cm^2 sec^{-1}). In molecular diffusion, $\lambda \approx$ free path and $\zeta \approx$ mean molecular speed; $d = \lambda\zeta$ is fixed by the density and temperature of the atmosphere; consequently, the height variation of d is marked by a curve. In eddy diffusion, $\lambda \approx$ mixing length and $\zeta \approx$ mixing velocity; owing to the variability of these elements, $D = \lambda\zeta$ and its variation with height are denoted by characteristic areas when the possible variability of D is narrowed by the consideration of limiting values of eddy accelerations (ζ^2/λ) and time terms (λ/ζ).

Another approach to the problem of turbulent diffusion [Pasquill (27)] has been used to deal with the small-scale dispersion of contaminants in the lower atmosphere. The normal (gaussian) distribution function provides a solution of the form

$$\chi(x,y,z,t) = Q(2\pi\sigma_y)^{-\frac{3}{2}} \exp \frac{-r^2}{2\sigma_y^2}$$

to the Fickian diffusion equation. In this generalized formula (Gifford [14]), χ is the concentration in the *cloud* or *plume* of material (which may be invisible), Q is the source strength at an instantaneous point source, and σ_y^2 is the variance of the dis-

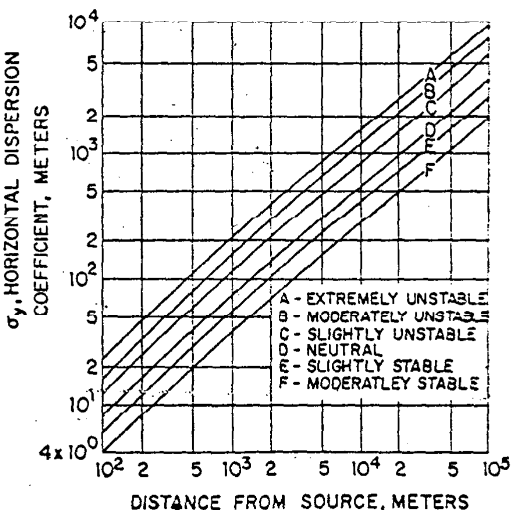


FIG. 2k-9. Graph of σ_y. (*From Gifford.*)

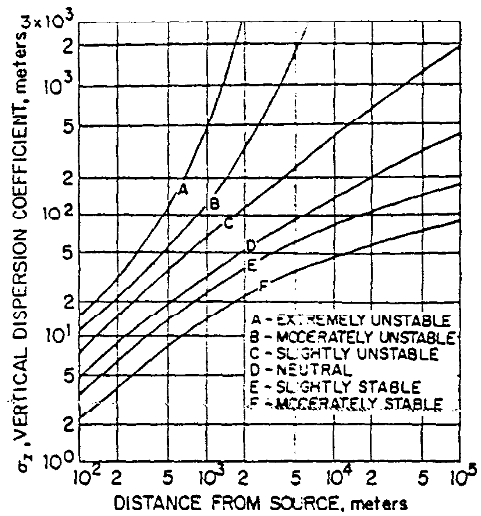


FIG. 2k-10. Graph of σ_z. (*From Gifford.*)

tribution of material in the plume. Since x is assumed equal to $\bar{u}t$, where $\bar{v}_x$ is the unfluctuating wind velocity component and t is the travel time of the cloud, $r^2 = [(x - \bar{v}_x t)^2 + y^2 + z^2]$. (Initially it is assumed that the diffusion is isotropic, i.e., $\sigma_x = \sigma_y = \sigma_z$.) In practice, the assumption is made that the diffusion takes place independently in the three coordinate directions, so that with the graphs of σ_y and σ_z shown in Figs. 2k-9 and 2k-10 it is possible [14] to compute the concentration under different conditions of atmospheric stability represented by the curves A, B, etc.

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