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UNPUBLISHED PRELIMINARY DATA

DYNAMICAL HEATING OF THE UPPER ATMOSPHERE

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Abstract. The energy that is deposited in the ionosphere, by internal atmospheric gravity waves propagating upward from below, is assessed on the basis of recent observational data. The implied heating rates are found to range from 10° K/day (near the 95-km level) to 100° K/day (near 140 km), and they therefore compete with solar radiation as the primary source of heating in the ionospheric E region. The tidal input may be comparable, but its height of deposition is more difficult to assess. The residual wave energy that reaches the F region carries a flux that may exceed 10^{-4} watt/m^2, so it may play a significant role in determining the heat budget of these higher levels. The waves should be accompanied by reversible temperature fluctuations of $\pm 10^{\circ}$ K and more, low in the E region, and they may therefore account for irregular temperature structure that has been reported.

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Introduction. The bulk of the atmosphere above the tropopause appears to receive its heat energy primarily through the absorption of optical radiation [Murgatroyd and Goody, 1958]. At very high altitudes, however, additional sources of heating are available and may become significant on occasion. One such source is provided by internal atmospheric gravity waves, which give evidence of propagating dynamical energy upward through the mesosphere into the ionospheric regions [Hines, 1960, 1963a, b].

The heating that may be caused by these waves is assessed in the present paper. It is found to be sufficient to modify temperatures in the E region, and may in fact account for the very rapid rise of temperature (with increasing height) that occurs there. Residual wave energy, propagating into the F region, contributes a small but sensible amount to the heat budget of the higher levels. Reversible temperature changes are also generated by the waves, of an amplitude that should be detectable and may already have been observed.

Earlier estimates. The present discussion follows upon another [Hines, 1963a] that was limited, insofar as heating was concerned, to heights near 95 km. At those levels, it was estimated, wave energy was being dissipated by molecular viscosity at a rate of about 5×10^{-2} watt/kg, and a corresponding heating rate of a few degrees per day was inferred. (One watt/kg will produce heating at a rate of 80° K/day, in an atmosphere of mean molecular mass 29 and a specific heat per molecule of

5×10^{-23} joules/^{o}K. This estimate depended sensitively on the spectrum that was assumed for the distribution of wave energy, as between modes of different wavelengths, and that spectrum could be inferred only crudely.

At the same time, the then current estimate of turbulent dissipation near the 95-km level, of 7×10^{-3} watt/kg [Blamont and de Jager, 1961; Greenhow, 1959], was noted as producing a heating of smaller but similar magnitude. An eddy viscosity about equal to ten times the molecular viscosity was also inferred, and would have implied a dissipation of wave energy at a rate of 5×10^{-1} watt/kg. Though this energy would have to be dissipated by the turbulence in turn, and though it exceeded the estimate of turbulence dissipation by almost two orders of magnitude, the discrepancy was not considered to be serious in view of the gross uncertainties inherent in all the estimates and, indeed, in the very applicability of standard turbulence concepts.

Instead, the view was adopted that a reasonably consistent picture had been evolved, in which turbulence was maintained in being by the wave system, and in which a heating of a few degrees per day could be produced by the waves. This heating might result primarily by direct molecular dissipation, or it might proceed mainly via the intermediate process of turbulence. In either event, it was comparable to the average heating produced at the same levels by the absorption of solar radiation [Murgatroyd and Goody, 1958], and must, of course, exceed the radiative input at night and in the winter polar regions (if a similar rate obtained). A secondary

source of heat in the latter regions appears to be required, to offset the rapid radiative cooling that would otherwise result, and so to maintain the rather high temperatures that are indicated there [Murgatroyd and Goody, 1958]. Such a source could be provided by subsidence and related photochemical release [Kellogg, 1961; Maeda, 1963; Young and Epstein, 1962] or possibly by acoustic heating [Maeda, 1964], but the role of internal gravity waves and associated turbulence appeared also to merit further examination.

New E-region estimates. It is now possible to estimate the gravity-wave dissipation at middle latitudes by a totally independent means, and to extend the calculation upward in height. Most valuable for this purpose is the analysis by Kochanski [1964] of the vertical variation of wave amplitude as revealed by numerous rocket-released vapor trails. This analysis is not wholly free from uncertainty, for the separation of wave-induced winds from the background was and will remain open to some ambiguity, but the results nevertheless constitute a tremendous advance and provide the best information of their type currently available.

Theory shows that, in an isothermal nondissipative atmosphere, free from background wind shear, the amplitude of the wave-induced horizontal wind speed (U_x) should increase with altitude (z) as the gas density (ρ) decreases, the product ρU_x^2 remaining constant [Hines, 1960]. It was

found by Kochanski [1964], however, that

$$\rho U_x^2 \propto \exp -(z/z_o) \qquad (1)$$

with

$$z_o = 7.6 \text{ km} \qquad (2)$$

over the height range 70-140 km. (U_x is represented by Kochanski as W, and the cited value of z_o is derived from the exponential decrease of $\rho^{1/2} \times W$ described on p. 3659 of his paper.)

This observed decrease of wave energy with increase of height might be due in part to partial reflection, associated with the height variation of temperature or of background wind that occurs in practice; some such reflection could provide a reasonable explanation, in terms of a standing-wave component, for ionization stratifications that are reported to occur at certain preferred heights at lower levels (as reviewed, for example, by Ellyett and Watts [1959]).. On the other hand, viscous dissipation is known to be strong for the wavelengths observed [Hines, 1960], and much of the energy must be going into heat whether the remainder is reflected or not. This view is strengthened by the observed increase of minimum half-wavelength (with increasing height), as revealed by the vapor-trail analyses of Kochanski [1964, Fig. 8] and Zimmerman [1964], both of which give excellent agreement with the theoretical viscous cut-off [Hines, 1964]. The following calculation of energy dissipation neglects partial reflection, but its conclusions are unlikely to be in error by more than a factor of two on this account.

Low in the E region, the bulk of the gravity-wave energy resides in modes whose vertical wavelength (λ_z) is of the order 10 km [Kochanski, 1964]. This is consistent with the wavelength of 12 km inferred by Hines [1960] from various reports of meteor data, including that of Greenhow and Neufeld [1959], and the wave period (τ) inferred from the latter report was about 1.2×10^4 sec. These waves would have a vertical component of group velocity (V_z) about equal in magnitude to the vertical phase speed: $V_z \simeq \lambda_z/\tau \simeq 1$ m/s. They oscillate nearly horizontally, and their mean energy density is given very nearly by $\rho U_x^2/2$. Their vertical energy flux is given by

$$F_z = \rho U_x^2 V_z/2, \qquad (3)$$

and the vertical convergence of this flux, $-dF_z/dz$, yields the rate at which they are losing their energy. The vertical variation of V_z that results from the vertical variation of temperature is negligible in comparison to the variation of ρU_x^2 given by (1) and (2), so the wave system appears to be losing energy at the rate $\rho U_x^2 V_z/2z_o$. When calculated for a unit mass of the atmosphere, the rate of energy loss becomes

$$\varepsilon = U_x^2 V_z/2 z_o. \qquad (4)$$

Kochanski finds $U_x = 40$ m/s as a mean amplitude at 95 km, which combines with the values of V_z and z_o already given to yield an estimate of $\varepsilon \simeq 10^{-1}$ watt/kg.

This estimate may be slightly high (because of partial reflections) or slightly low (because it makes no explicit provision for the energy borne by modes of smaller scale and shorter period, for most of which V_z would be greater). It is, however, totally independent of the mechanism -- molecular or turbulent -- that is assumed to be removing the wave energy, totally independent of coefficients that pertain to those mechanisms, and substantially independent of the detailed spectral distribution. Its agreement with the earlier estimates of wave dissipation, 5×10^{-2} or 5×10^{-1} watt/kg, is the more remarkable on this account. Even the discrepancy that persists, between it and the earlier estimate of turbulent dissipation, is somewhat relieved by the higher values that are now being associated with turbulence (e.g., $1.5 - 3.5 \times 10^{-2}$ watt/kg [Roper and Elford, 1963; Noel, 1963], and even 7×10^{-2} watt/kg [J. E. Blamont, private communication]).

Regardless of its route, the dissipated wave energy should appear ultimately as heat, and it implies a heating rate of the order 10° K/day. As previously argued, this heat input is comparable to the average radiative input, and its dominance over the latter at night and in the winter polar regions must be anticipated.

An even more striking conclusion can be reached for somewhat higher levels. Around 105-110 km, Kochanski reports $U_x \simeq 55$ m/s, which combines with earlier values to yield $\varepsilon \simeq 4 \times 10^{-1}$ watt/kg or a heating of about 30° K/day.

U_x decreases to 40 m/s once again near the 140-km level, but the waves that still remain observable as waves are confined to wavelengths $\lambda_z \gtrsim 30$ km. This is to be expected from viscous quenching criteria

[Hines, 1960, 1964]. (Incidentally, the dissipation here can no longer proceed via turbulence, for the turbulence terminates typically at heights of 100-110 km [e.g., Blamont and de Jager, 1961]. The increase of molecular kinematic viscosity more than makes up for the loss of eddy viscosity as a dissipative mechanism, however.) These quenching criteria further indicate that the period to be associated with the limiting λ_z (i.e., 30 km) is no greater than 10^3 sec, and the vertical phase speed is then close to 30 m/s. V_z will be somewhat less than this, but the energy density will exceed $\rho U_x^2/2$ by a compensating factor. The implied dissipation rate exceeds 1 watt/kg, and leads to a heating of the order 100° K/day. This conclusion must be modified by the fact that Kochanski reports the waves to be present at 140 km for only 50 percent of the time, but that conclusion in turn must be qualified because his method of analysis is such as to minimize the role that is attributed to the larger-scale waves that are relevant here.

Doubtful though the exact value of the dynamical heating rate may be, there is no denying the implication that a very substantial heat input is available from internal gravity waves in the E region. This heating, of course, exceeds the radiative input at night, and may very well exceed that input even when both are integrated over 24 hours. It may, in fact, be the primary cause for the very rapid rise of temperature with height that is found in the E region -- the relative rate of increase being there the maximum achieved in the whole of the atmosphere -- and, by limiting the flow of heat downward from the F region, it may exert a strong influence on the temperature of the whole thermosphere.

F-region heating. The direct input of wave energy to the F region is worth evaluating, though the uncertainties increase once again. A density of about 3×10^{-9} kg/m^3 obtains at 140 km [CIRA, 1961], and combines with other values already given to yield an energy flux of about 7×10^{-5} watt/m^2 as the input to the overlying F region, 50 percent of the time. This estimate could be low by as much as an order of magnitude, contributed by the larger-scale waves that blend into the 'residual wind' in Kochanski's analysis. Such waves, if responsible for ionospheric traveling disturbances as has been argued [Hines, 1960], are in fact present most of the time [Heisler, 1963]. Their energy density could equal, and their V_z's tend to exceed, the values employed above. A mean energy flux exceeding 10^{-4} watt/m^2 is therefore a distinct possibility.

For purposes of comparison, it is usual to cite 10^{-3} watt/m^2 (= 1 erg/cm^2/sec) as the rate at which the F region gives up its heat energy, via thermal conduction into the E region [Bates, 1951; Harris and Priester, 1962; Hunt and Van Zandt, 1961; Johnson, 1958]. Much of this energy is thought to be supplied by the extreme ultraviolet component of solar radiation, which carries a flux about three times as great [Hall et al, 1963; Hinteregger, 1961]. However, this flux is vertically incident only at the subsolar point, it is present at all only by day, and its conversion to heat is thought to be only 15-30 percent efficient [Chamberlain, 1961; Hanson and Johnson, 1961]. Its ability to meet the heating requirements of the F region is doubtful, and no other solar radiations have been

advocated to make up the deficiency. Indeed, one analysis [Harris and Priester, 1962] suggests that a second heat source of quite different diurnal variation may be required to account for the observed density variations.

The calculation presented above gives some basis for looking to the dynamical energy of internal gravity waves as an auxiliary heat source for the F region. It is certain, for one thing, that the entire energy flux that enters that region will degenerate to heat there. Moreover, because of variations in the (as yet unknown) sources of the waves, and because of tidal changes that will affect their passage through the E region, the flux will undoubtedly contain some diurnal modulation. Any speculation as to the phase or amplitude of such a modulation would be idle at the present time, however, and so will not be pursued here.

Magnetic storms. At times of magnetic storms and active auroral displays, an additional component of the gravity-wave spectrum may be anticipated. This is because of the intensive heating that occurs in the auroral zones, whose spatial and temporal scales lie within the range that could efficiently couple energy into the waves. It has been suggested by T. Gold [private communication] that energy might be transported from the auroral zones to lower latitudes by means of the gravity waves, and thereby provide for the excessive heating that is revealed by satellites at such times [Jacchia, 1964].

This suggestion meets no particular difficulty with respect to travel times, for the energy could propagate to lower latitudes at speeds of almost 1000 m/s in a ducted mode [cf. Press and Harkrider, 1962, Fig. 6] that is revealed, for example, by the gravity-wave component of the high-altitude nuclear blast waves reported by Obayashi [1962, note added in proof] and by a small but important class of ionospheric traveling disturbances found in the data of Tweten [1961] and Valverde [1958]. The magnitude of the effect is subject to uncertainty, however, and remains for further examination when more suitable data become available.

Reversible heating in the E region. The heating that has been treated to this point has been irreversible, the energy being lost from the wave system. However, until dissipation becomes severe in any given mode, that mode produces reversible, adiabatic heating, as one facet of its oscillation. The fractional temperature variations ($\delta T/T$) are related to U_x by

$$\delta T/T = (P - R)X^{-1} U_x \qquad (5)$$

where P, R and X for a given mode are as defined by equations (15) - (17) or (23)- (25) of Hines [1960]. The full expressions are complicated, but reduce to the simpler form

$$\delta T/T \simeq \pm i (\gamma - 1)^{\frac{1}{2}} C^{-1} U_x \qquad (6)$$

when, as is the case for the dominant E-region waves, λ_z does not exceed appreciably the local scale height of the atmosphere, and τ is substantially greater than $\tau_g \equiv 2\pi C g^{-1} (\gamma - 1)^{-\frac{1}{2}}$ (which is the Brunt-Vaissala period

for an isothermal atmosphere). Here γ is the ratio of specific heats, C the speed of sound, and g the acceleration due to gravity; the "i" indicates a phase quadrature between the time or place of maxima in δT and U_x, while the "$\pm$" uncertainty can be eliminated if the horizontal direction of phase propagation is known in relation to the direction of U_x. The temperature deviations that arise if τ tends toward τ_g exceed those indicated by (6), provided the restriction on λ_z is retained. The theory is based on an assumed isothermal atmosphere, but this idealization should not seriously restrict the validity of (5) and (6) as first approximations.

On the insertion of representative values into (6), it will be found that temperature fluctuations of $\pm 10^{\circ}$ K can be expected low in the E region, and as much as $\pm 30^{\circ}$ K at 110 km, quite typically. Irregular temperature profiles just below the 100-km level have been inferred by sound-ranging methods, with deviations often as large as $\pm 10^{\circ}$ K, and these deviations apparently exceed the probable error of measurement [Smith et al, 1964]. It is natural to suggest that they represent the reversible heating imposed dynamically by gravity waves, rather than quasi-static anomalies or purely spurious results. This suggestion is amenable to a detailed check (with respect to phase as well as amplitude) by refined programs of simultaneous temperature and wind sensing, and these programs could be extended to include density variations when measurement techniques improve slightly. Such analyses would go far towards consolidating or clarifying the picture of irreversible heating previously presented, in addition to their more direct benefits.

Tidal and prevailing-wind input. The atmospheric tides carry significant energy into the E region and distribute it through higher levels. They should not be ignored in any discussion of dynamical input, though their role is difficult to assess at present.

The semidiurnal tide at the base of the E region exhibits a vertical variation of phase amounting to 3-7°/km [Greenhow and Neufeld, 1961], which corresponds to a λ_z of 50-120 km. Higher in the E region, vapor trails reveal tidal λ_z's more of the order 30 km [Kochanski, 1964], and two sequences of trails [Rosenberg and Edwards, 1964] suggests that this λ_z should again be associated with the semidiurnal component. It would be convenient if these λ_z's were to correspond to the so-called '2, 2' mode of tidal oscillation that predominates at ground level and in 'resonance' theory, and whose associated 'effective height of the atmosphere' is h = 7.9 km [e.g. Wilkes, 1949]; but they don't, nor is there any good reason why they should [Hines, 1963a].

They imply, instead, modes with h $\simeq$ 2.6 km, when interpreted on the basis of conventional tidal theory, but their identification can be made no more precise than that. This is unfortunate, for the relation between observed winds and corresponding energy fluxes varies considerably from one mode to another, as an examination of the formulae (33), (34) and (39) of Wilkes [1949] will reveal. (The concept of a vertical group speed is no longer valid, because of the discrete nature of the spectrum of tidal modes.) Representative calculations may be carried out, nevertheless, and they suggest that the semidiurnal tidal flux upward into the E region is

about 10^{-3} watt/m^2, while the energy density at 90 km is about 5×10^{-3} joule/m^3 and so the effective vertical speed of energy flow is about 2 m/s. The figures are based on a speed of 20 m/s for the semidiurnal tidal wind at 90 km [Greenhow and Neufeld, 1961], on an $h \simeq 4$ km, and on uncertain trigonometric factors which arise in the mode analysis and which render the estimates of flux and speed suspect by a factor of four or so.

These values of flux, density and speed, are quite comparable to those of the gravity-wave system low in the E region. The tidal energy, if deposited there, should produce a heating ~~of the same order~~ as that already inferred for the gravity waves; if deposited at higher levels, the heating could be even greater.

This raises a further point of uncertainty with respect to the tides: the height of energy deposition. It seems quite clear that the energy is largely removed from the tidal wave itself below heights of 115 km [Hines, 1960]. Even when the wave penetrates to greater heights, as it does on occasion [e.g., Kochanski, 1964], it does not maintain anything like a constancy of ρU_x^2. The energy might be deposited in heat through the direct action of viscosity or through a cascading of energy in the gravity-wave spectrum, but hydromagnetic dissipation -- or, equivalently, ~~the~~ 'ohmic loss' or 'joule heating' -- seems also to be a strong contender [Hines, 1963a]. In the course of such dissipation, however, electric fields are established and act to extend the whole tidal phenomenon upward to levels that the tidal wave itself is unable to reach (as reviewed by Fejer [1964] or Hines [1963a], for example).

The extent to which the semidiurnal tide is in fact carried upward electrodynamically is by no means clear, conventional theory to the contrary. This is because conventional theory neglects any vertical variation of tidal wind in the E region, whereas all recent information suggests that variation to be strong. Its effect would be to discharge, low in the E region, much of the electric field that would otherwise build up, and to produce there most of the ohmic losses of the system.

Limited data exist to suggest that the diurnal tide exhibits little variation of phase with height [Greenhow and Hall, 1960], and, if confirmed, they may imply a local source for this tide rather than upward propagation through the mesosphere. Similarly, the prevailing winds of the E region would be generated primarily by local input, and should not be expected to reverse themselves within the E region. In any event, the diurnal tide [e.g. Kato, 1956] and the prevailing winds [van Sabben, 1962] are now receiving increasing attention as the likely sources of the quiet-day ionospheric current system, and it is their energy rather than that of the semidiurnal tide that would then control the electrodynamic system at higher altitudes.

Regardless of the driving agency, some estimate may be made of the joule heating that is associated with the observed magnetic variations. One calculation is provided by Cole [1962], and it implies a dissipation of 5×10^{-5} watt/m^2, in a region centered on the 140 km level. This is certainly compatible with the semidiurnal tidal flux cited above, while

the appropriate figures for the diurnal and prevailing components are unknown. It is, moreover, as Cole points out, of possible significance to the heat budget of the higher levels. While it is less by an order of magnitude than the requisite 140-km flux previously quoted, it is also uncertain by an order of magnitude. (The local dissipation rate to which this corresponds, at the 140-km level, can be calculated only if an explicit assumption is made as to the magnitude of the associated electric fields. Representative calculations suggest the rate to be about 3×10^{-1} watt/kg, which would produce a heating rate of about 25° K/day, but again with a substantial uncertainty.)

In summary of this final section, then, neither the propagation of tidal energy into the ionosphere, nor the electrodynamic transfer of tidal and prevailing wind energy within the ionosphere, should be neglected in any serious attempt at establishing the normal thermal budgets of the various upper atmospheric regions. But the specific regions in which these processes will be most significant, and the degree of their significance, cannot yet be established with useful precision.

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