PHYSICS OF METEOR FLIGHT IN THE ATMOSPHERE

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   PHYSICS OF METEOR FLIGHT IN THE ATMOSPHERE

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Introductory Note

The booklet has originated from a series of lectures on "The Physics of Meteors and their Cosmic Relationships," delivered in the spring term of 1957 at the Department of Physics of the University of Maryland, during the author's stay there as Visiting Professor.

The present scope is somewhat narrower and covers chiefly the phenomena occurring during the flight of the meteor through the terrestrial atmosphere. The intention is to provide a basis for further research.

The basic units of the cgs system are used throughout unless stated otherwise. Thus, normally lengths will be in centimeters, velocities in cm/sec, and so on.

A number of symbols, listed as Principal Notations at the end of the book, will have always the same meaning and, to save space, will be defined only when they occur for the first time. Other notations will be currently defined in the text.

Contents

The Setting

1.1 Problems

The physical theory of meteors considers processes which are taking place during the flight of the meteor through the atmosphere and, for larger bodies, at impact on the earth's surface. The main purpose is to predict the variation of mass, velocity, luminosity, and ionization along the meteor trajectory.

In meteor physics, calculations are sometimes only approximate; however, results to a "close order of magnitude" ($\pm$ 50 %) often suffice.

The depth of penetration of meteors into the atmosphere depends mainly upon their size and velocity. The quantitative difference in atmospheric density may be so great that a qualitative difference in the treatment becomes necessary. As a consequence, the theory of the processes occurring in collisions of meteors with our planet can be subdivided into three major "problems," with transitions between them.

Problem 1: This concerns the case when the "free path" of the air molecules is greater than the linear dimension (radius) of the meteor. The impact momentum and energy are transmitted to the nucleus of the meteor by direct hits of the air molecules. No hydrodynamic cushion or "air cap" is formed. Practically all meteors of the "visual" range belong to this case; they are so small and disintegrate in atmospheric strata of so low a density that the condition of free path is always amply fulfilled (Fig. 1).

One consequence of the lack of shielding is the relatively high coefficient of heat transfer; with this, only a small fraction of the kinetic energy of the meteor is needed to achieve complete vaporization of its substance; the meteor disintegrates before noticeably decelerating. Hence in many cases the problem can be further simplified by assuming the velocity to remain nearly constant.

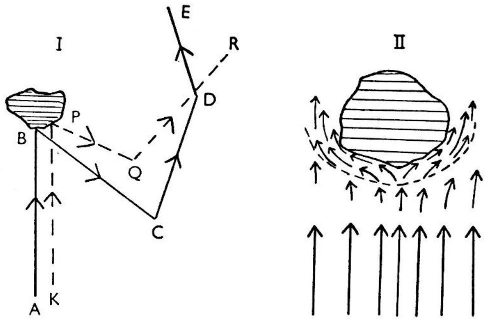

Fig. 1. Motion of air relative to meteoroid. I = Problem 1. The air molecule A hits the meteoroid at B and rebounds along the zig-zag track BCDE, being in collision with other air molecules at C and D; the stretches BC, CD are large as compared with the diameter of the meteoroid.

II = Problem 2. The air stream obviates the obstacle along streamlines shown by arrows. Between the broken line and the meteoroid an air cap, or condensation, is formed. The free path of the molecules is small as compared with the diameter of the meteoroid.

Problem 1a: This is the particular case of micrometeors which are efficiently decelerated in atmospheric strata of so low a density that the temperature of intense vaporization is not attained, and the energy is dissipated through radiation. Micrometeors may be brought to rest without much of ablation, and may in the course of time settle to the ground as particles of dust.

Problem 2: This is the case when the free path of the air molecules is smaller than the linear dimension of the meteor. A hydrodynamic cushion, or air cap, is formed in front of the meteoroid (nucleus) (Fig. 1). The heat transfer is impeded, the more so, the deeper the meteor penetrates into the atmosphere. The aerodynamic resistance, or drag is also smaller than for equal velocity and atmospheric density in Problem 1, because the air cap helps in streamlining the flow of air. Once the meteor has penetrated deep enough into the atmosphere for an air cap to be formed, the protective quality of the latter will favor further penetration. The heat transfer coefficient may decrease by two orders of magnitude, as compared with Problem 1, whereas the atmospheric drag can only decrease by a factor of about 2. Ablation is greatly suppressed, yet deceleration is affected very little. For a sufficiently large initial size, the meteor may be stopped before it has completely disintegrated.

So-called "fireballs" are of a brilliance equal to that of the moon, or brighter, and the meteorite falls belong to this class.

Problem 3: This is the case of collision of a meteor with a dense body, solid or liquid. From the preceding two cases this differs by the meteor and the medium being of comparable density. The conditions are extremely complicated. On account of the high aerodynamic pressures and plastic deformation, the distinction between meteor and medium is practically erased; the propagation of shock waves and shock fronts, instead of the motion of the meteor body, is to be considered. The theory of formation of meteor craters, collisions between meteors, or between asteroids in space, and meteoric skin erosion of space vehicles belong to the scope of Problem 3.

In the present tract we will concern ourselves mainly with the first two problems.

1.2 Historical

The beginnings of a physical theory of meteor phenomena seem to have been first laid down by Öpik in 1922.[1] A relation of general validity between the variable mass and velocity of a meteor was derived. It was established that the intensity of observable radiation from a meteor is proportional to the mass of vapor set free per unit of time, or that brightness can be measured by the rate of ablation. A light curve, or variation of brightness of the meteor along its path, was calculated for an atmosphere of constant density.

Hoppe[2] derived a light curve upon the same principle for motion in an atmosphere of variable density.

An attempt of a detailed physical theory of meteors was made by Lindemann and Dobson in 1923.[3] Unfortunately, it was marred by serious misinterpretations of the laws of physics. Their most serious shortcoming amounted to a neglect of the law of conservation of energy; namely, they calculated the temperature of air in front of the meteor from the trivial adiabatic formula of compression; yet the adiabatic formula is only valid when the compression is slow as compared with the gaskinetic velocity of the molecules. This condition is not fulfilled in the compression of air in front of the moving meteor. In the words of Sparrow,[4] "The use of the adiabatic equation by Lindemann and Dobson is equivalent to the assumption that a velocity of 60 km/sec is small compared to one of 0.5 km/sec." The procedure cannot even be called an approximation, less than 1 per cent of the kinetic energy being accounted for. For this and other reasons their treatment of the meteor problem is unacceptable.

A most important step forward in the physical theory of visible meteors was made in 1926 by Sparrow.[4] He pointed out that these objects (meteors of the visual range) cannot produce an air cap and that momentum and energy transfer are achieved through direct bombardment of the meteor nucleus by air molecules. Sparrow actually laid the foundation of the modern detailed theory of the balance of energy and mass in visible meteors (Problem 1). A further contribution to the meteor theory on rational lines came from H. B. Maris.[5]

A study entitled "Basis of the Physical Theory of Meteor Phenomena" was published by Öpik in 1937.[6] This is an analysis of allegedly all relevant factors, instrumental in the process of ablation and atomization of the meteor substance. No doubt, many a factor may nevertheless have been omitted, such as, e.g., sputtering, or the newly discovered phenomenon of dustballs. However, it seems that the enumeration is reasonably complete from the standpoint of processes occurring in stonelike visible meteors. Together with an earlier paper, titled "Atomic Collisions and Radiation of Meteors,"[7] which contains a sketch of a quantitative semi-empirical theory of the efficiency of ionization and excitation in collisions of slow atoms or ions, this study has served as a basis for further research.

The factors numerically evaluated included aerodynamic drag, coefficient of accommodation, radiation losses, vaporization, fusion, viscosity and spraying of liquid at fusion, stability of liquid drops in flight, role of rotation, aerodynamically induced oscillations and damping of rotation, formation of an air cap, shielding effect of own vapors, heat transfer by conductivity, and impact radiation from vaporized substance.

From this stage, further theoretical progress can be achieved mainly by numerical integrations. The integrations need not be of high numerical accuracy, but of paramount importance is the condition of a realistic approach; this means that no relevant factor should be overlooked or omitted merely for the purpose of mathematical expediency.

Meteor theory has been progressing through the application of the method of trial and error. A provisional set of assumptions is followed by analysis which shows where the assumptions are in need of adjustment; the adjusted set is taken for a new start, and the procedure is repeated until consistent results are obtained.

It may happen that inconsistency cannot be eliminated, which would mean that some of the assumptions are basically wrong. Such was the case with the analysis of the heights of appearance and disappearance of visible meteors;[8,9] the observational results, coupled with atmospheric data obtained from rocket research, could not be reconciled with any rational assumptions as to the process of ablation of a stony meteoroid. The conclusion was reached that the meteors of the visible range, as a rule, are not stony compact bodies, but some kind of loose aggregates of dust particles, "stoneflakes" or "dustballs." Although the "classical" theory of ablation, as it could be called, proved inapplicable, without the help of this theory the correct conclusion could not have been reached.

In other cases, e.g., for micrometeors and for large meteors (fireballs) and meteorites, the assumption of a dense stony (or iron) kernel has proved consistent with the observations.

The Atmosphere

2.1 Meteors as Probes of the Upper Atmosphere

The correct interpretation of meteor phenomena presupposes a knowledge of the structure and composition of the atmosphere to a height of at least 130 km.

On the other hand, it may seem to be possible to derive the properties of the atmosphere from meteor observations when the average density and shape of the meteors are postulated. In such a manner, Whipple[10] calculated atmospheric densities from the deceleration of bright photographic meteors within the altitude range of 60 to 95 km. The results, however, are subject to large systematic errors which are altitude-dependent and, therefore, are of doubtful value.

Discrepancies for meteors of the visual range led to the recognition of dustballs[8,9] showing that our knowledge of meteors is much less certain than that of the atmosphere. The problem is to be inverted and, by taking as a basis the observed structure of the atmosphere, e.g. that of the Rocket Panel,[11] one has to infer the properties of meteors that are compatible with the observations.

2.2 Diffusion and Turbulence

A quarter of a century ago the problem of the structure of the upper atmosphere was bedevilled by the theory of conductive, or diffusive equilibrium (Humphreys, Jeans). In undisturbed gas diffusion the constituents of the atmosphere settle into states of equilibrium which are independent of each other; the partial pressures are then governed by the equation

$$dp/p = -dH/a, \tag{2-1}$$

where

$$a = kT/(\mu g) \tag{2-2}$$

is the scale height. Here p = partial pressure, μ = molecular mass (absolute), k = Boltzmann's constant, g = acceleration of gravity, H = altitude. The pressure of a light gas like hydrogen will decrease more slowly with altitude than that of the heavy constituents and, although relatively insignificant at sea level, hydrogen would predominate in the upper atmosphere (above H = 80 km). This concept proved a mathematical fiction in the true sense of the word. Sparrow,[4] and especially Chapman[12] denied any role to hydrogen. Öpik[6] pointed out that, contrary to observation, atomic lines of more than 4 ev excitation could not appear in the spectra of meteors at velocities below 30 km/sec if hydrogen were the chief constituent; the lack of differentiation of the atmospheric constituents he explained by turbulent mixing.

Some properties of turbulence which play a role in the theory of meteors are outlined below.

Turbulence replaces laminar flow when the Reynolds number,

$$Re = \rho v L/\eta, \tag{2-3}$$

exceeds a certain limit of the order of 1000. Here ρ is the density, η the coefficient of viscosity of the fluid, v an average value of the velocity of flow, and L a characteristic linear dimension (radius of pipe, thickness of layer of flow at right angles to the direction of motion, etc.). $D_g = \eta/\rho$ is the kinematic viscosity, identical with the gaskinetic coefficient of diffusion. With Re = 1000, Eq. 2–3 defines the maximum admissible thickness of a laminar layer (skin layer).

In the free upper atmosphere $L \sim a \sim 1.2 \times 10^6$, $v \sim 10^4$, $\eta = 2 \times 10^{-4}$ (at + 300°C), whence

$$Re \cong 6 \times 10^{13} \rho. \tag{2-4}$$

Hence, for $\rho > 1.7 \times 10^{-11}$ g/cm³ the Reynolds number indicates spontaneous turbulent flow; this corresponds to an altitude below 130 km.[11] Forced turbulence will extend to even greater altitudes, being caused by the impact of turbulent elements from below.

Turbulence leads to accidental fluctuations of the velocity components; for a normally horizontal flow, such as the winds in the ionosphere are (average wind velocity 95 m/sec in 100–120 km level;[13] 50 m/sec in 80–100 km level;[14] 54 m/sec at $H = 90$ km and 83 m/sec at $H = 98$ km[15]), vertical components arise which are instrumental in mixing the atmosphere through turbulent or eddy diffusion. Although favored by mountains, these vertical motions belong to the very nature of turbulence; there is no need for solid obstacles to cause them—they are produced in the collisions of disorderly moving turbulent "packets" in the free atmosphere.

Measurements with a sensitive anemometer by Öpik (1927–39, unpublished study of diffusion of gas clouds) give an average vertical deviation of wind direction of $\varepsilon = \pm 6°$ even under the smoothing presence of a level surface; being a time average, this implies a semiamplitude in the vertical velocity component equal to

$$u = 2v \sin \varepsilon = 0.2v, \tag{2–5}$$

v being the wind velocity.

In an atmosphere of stable lapse rate an air packet which has acquired a vertical velocity component u will cool from adiabatic expansion while rising; it will be slowed down by a differential acceleration of gravity equal to

$$\Delta g = F_1 g(\nu - \nu_0)x/T \quad (\text{cm/sec}^2), \tag{2–6}$$

where x is the differential height, $\nu = dT/dx$ the lapse rate, ν_0 the adiabatic lapse rate, and F_1 a dimensionless factor which allows for the lateral exchange of heat by second-order turbulence; empirically $F_1 = 0.096$.[16] The work per unit mass is

$$\int_0^x \Delta g \, dx,$$

and the vertical motion stops when this equals $\frac{1}{2}u^2$. Hence the maximum vertical length of path of the air packet becomes

$$x_{\max} = u[F_1 g(\nu - \nu_0)/T]^{-\frac{1}{2}}. \tag{2-7}$$

With $T = 360°\mathrm{K}$, $\nu = +0.6 \times 10^{-4}$ deg/cm as for altitudes of 110–140 km,[11] $\nu_0 = -0.8 \times 10^{-4}$, $g = 940$, we have $x_{\max} = 220u$. With $v = 10^4$, $u = 2000$, $x_{\max} = 4.4 \times 10^5$ cm. Allowing for aerodynamic resistance and dissipation of the air packet, the vertical exchange depth for turbulent mixing may be assumed equal to one-half of that, or to $\Delta H = 2.2 \times 10^5$ cm.

The coefficient of turbulent diffusion in the vertical direction, equal to the product of effective length of vertical path times average vertical velocity component, thus becomes

$$D_t = \tfrac{1}{2}u \cdot \Delta H, \tag{2-8}$$

or

$$D_t = 2.2 \times 10^8 \ (\mathrm{cm^2/sec}). \tag{2-9}$$

The flow of mass in the direction x is given by

$$\rho\left[\left(\frac{\partial z}{\partial x}\right) - \left(\frac{\partial z}{\partial x}\right)_0\right]$$

(g/cm² sec), where $\partial z/\partial x$ is the gradient of partial concentration of an atmospheric constituent, $(dz/\partial x)_0$ the equilibrium value of the gradient. For turbulent diffusion $(\partial z/\partial x)_0 = 0$. For gaseous diffusion $(\partial z/\partial x)_0$ is determined by Eq. 2–1, with the actual local values of p and T.

The isotropic coefficient of gaseous diffusion is given by

$$D_g = \tfrac{1}{3}\lambda_0 \bar{v}, \tag{2-10}$$

where λ_0 is the mean length of path (for 90° deflection in angle), $\bar{v}$ the gaskinetic arithmetical mean velocity.

For an initial density distribution of excess substance, cylindrically symmetrical around a straight axis (as in ionized columns of meteors) and of the form $\exp(-x^2/\Delta^2)$, the diffused distribution after the lapse of time t becomes $\exp[-x^2/(\Delta^2 + 4Dt)]$ when there is no external supply of