SATELLITES AND METEOROLOGY

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ANY reference to satellites usually brings to mind research in outer space or, at least, in the highest layers of the atmosphere. It is quite clear, however, that satellites can play an invaluable role in the study on a global scale of processes that take place in the lower layers of the atmosphere, and are responsible primarily for changes in the weather. One of the most instructive examples of this kind is the use of the possibility of tracing the distribution of clouds with the aid of television cameras mounted on a satellite. In spite of the relatively extensive network of meteorological stations, approximately $^4/_5$ of the earth's surface, under the oceans and seas, is very sparsely covered by meteorological observations (only through sporadic ship measurements). Even the distribution of the meteorological stations on land is highly irregular: large portions in sparsely populated regions are still "blank spots" to a considerable extent. Thus, the now existing network of meteorological stations does not permit an effective observation of the cloud conditions (and conditions of other meteorological elements) over the entire surface of the earth. But the simultaneous launching of a few satellites bearing television cameras would be enough to trace practically continuously the distribution of the clouds over the entire earth's sphere. Similar capabilities are uncovered by the use of satellites to investigate the energy balance of the earth as a planet, the thermal conditions of the earth's surface and of the atmosphere, and many other important problems. The practical value of these problems lies in the fact that they are of decisive importance in the analysis and forecasting of weather-producing processes.

It is the purpose of this article to review the foreign researches in which satellites are used to study the lower layers of the atmosphere (troposphere and stratosphere). We first describe briefly the program for meteorological observations with the aid of satellites and discuss the associated general problems. We then proceed to a specific analysis of the most important types of meteorological observations with the aid of satellites.

During recent years many proposals have been made concerning the use of satellites for meteorological purposes (see Singer[4]). Let us list the types of problems of greatest interest.

1. Tracking clouds by television. The purpose of this research would be to observe the distribution of clouds over the earth's sphere. Since television tracking is possible only during the day, it is proposed to use thermal direction finding for observation of clouds at night.

2. Research on the radiation balance between the earth and the atmosphere. It is proposed to measure by means of satellite-borne instruments the components of radiation balance of the earth-plus-atmosphere system, so as to obtain information on the laws governing the energy balance of the earth as a planet.

3. Study of the thermal stratification and composition of the atmosphere. Inasmuch as the thermal radiation of the "earth-atmosphere" system (called the outgoing radiation) is determined by the thermal stratification of the atmosphere and by the altitude distribution of the absorbing and radiating components of the atmosphere (primarily water vapor and carbon dioxide), the results of measurements of the outgoing radiation in different parts of the spectrum can be used as a source of information on the thermal stratification and composition of the atmosphere.

4. Investigations of the cloudiness and precipitation with the aid of radar. The highly successful results of research of cloudiness and precipitation made with earth-based radars give grounds for hoping that radio meteorology can also assume an important place in meteorological observations with the aid of satellites. Related to this are researches on atmospherics, aimed at ascertaining the distribution of centers of thunderstorm activity over the earth's sphere.

5. Distribution of energy in the ultraviolet spectrum of the sun. It is intended here not only to investigate the spectral composition of the ultraviolet solar and scattered radiation and its variation, but also to determine from these data the ozone content and to solve a few other problems connected with the optical properties of the atmosphere.

Along with the foregoing most important groups of problems, discussions have been raised concerning some other unexplained matters (investigation of the turbidity of the atmosphere, measurements of the mass of the atmosphere and the pressure on the surface of the earth, etc.).

The specific plans for the use of satellites in meteorological observations do not cover as yet many of the questions raised above. An idea of the specific researches planned in the USA can be gained from the list of American meteorological satellites, borrowed from the review of J. Holmes[16] (see Table I).

An examination of this table shows that the main trends in future research will involve television tracking of cloudiness and a study of the radiation

Table I. American meteorological satellites.

Name of satellite	Date of launching	Type of rocket	Angle of inclination of orbit	Stabilization	Scientific apparatus
Tiros I	April 1, 1960	Thor-Able	50°	Unoriented	Two television cameras
Tiros II	November 23, 1960	Thor-Delta	50°	Ditto	Television cameras, instruments for measurement of infrared radiation from the earth
Tiros III	1961	Thor-Delta	50°	Ditto	The same as on Tiros II
Nimbus I	1962	Thor-Agenda B	Polar orbit	Oriented relative to the earth	Television cameras, instruments for measuring the earth's infrared radiation and solar radiation
Nimbus II	1963 or 1964	Thor-Agenda B	Ditto	Ditto	The same as on Nimbus I, and also infrared spectrometer and radar
Aeros	1964 or 1965	Centaur or Saturn	Ditto	Ditto	The same as on Nimbus II

balance of the earth-atmosphere system. It is intended to use satellites for radio meteorological research at a later stage, followed by a study of the spectral variation of the infrared radiation of the earth-atmosphere system for purposes of thermal sounding of the atmosphere. An essential feature of the latest stage of this research is the use of satellites, oriented relative to the earth, in polar orbits. This means that the television cameras and radiation receivers will always be directed normal to the earth's surface. It is assumed that the Nimbus satellite, weighing approximately 260 kg, will be launched in an almost circular orbit approximately 1,000 km above the earth's surface on an orbit inclined 80° to the equator.

As noted earlier, the principal advantage of meteorological observations from satellites is the possibility of gaining information on the characteristics of different processes in the atmosphere over the entire (or almost entire) territory of the earth's sphere, and also a continuous tracking of processes that take place over definite large portions of the earth. This raises the question of the most expedient satellite orbits.

It must be stated first that circular orbits are the most suitable for meteorological investigations, for they obviate the need for corrections for change in height above the earth's surface. As to the choice of the orbit inclination, there is no doubt that polar and equatorial orbits are of greatest interest. In the former case the satellite-borne instruments can "see" the entire earth's surface. If a satellite is launched northward on a polar orbit at local noon, it will be observed at noon at all latitudes, up to the North Pole. After passing through the pole, the satellite enters the earth's shadow and will cross all latitudes at local midnight. Owing to the earth's rotation about the sun, the local time of passage of the satellite across each latitude will advance four minutes each day. In order to eliminate the effect of the earth's rotation around the sun and to ensure the appearance of the satellite above a point of given latitude at precisely local noon (on the illuminated side) or local midnight (on the shadow side), it is obviously necessary to incline the satellite orbit slightly in a western direction relative to the plane of the equator. Calculation shows that if the orbit altitude is 500 km, the satellite must be launched 7° to the west of the pole. Using a system of such "quasi-polar" satellites, so launched that they cross all the latitudes at fixed times, it becomes possible to track processes over the entire earth's atmosphere almost continuously. R. Heaviland (see reference 16) believes, for example, that even two satellites in polar orbits located in mutually perpendicular planes would be sufficient for the purpose, if the orbits are of 3600 and 7200 km high. Naturally, in this case geographic distributions of the investigated quantities will be mapped on a very large scale. In view of the variety in the scales of the processes in the atmosphere, it is naturally necessary to have not only information averaged over large areas, but also information on local processes of relatively small scales. This means that one cannot make do with high-altitude satellites alone.

Along with the "polar" satellite, considerable interest attaches to a satellite launched along the equator. Such a satellite, at an altitude of approximately

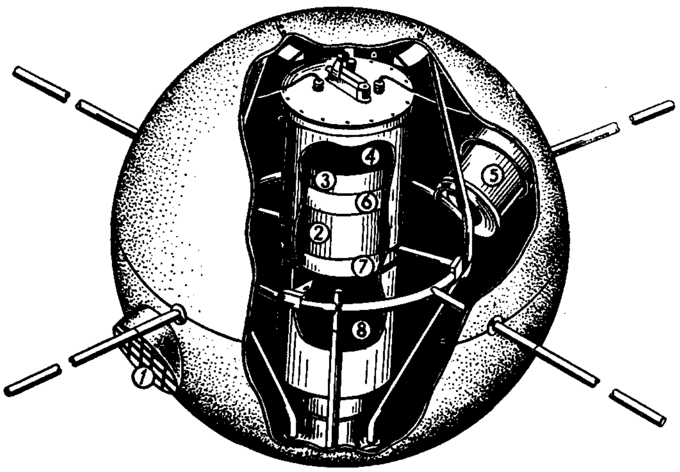

FIG. 1. Diagram of the satellite Vanguard II. 1 and 5 – photocell; 2 – recorder; 3, 4, 6, 7 – electronic units; 8 – power sources.

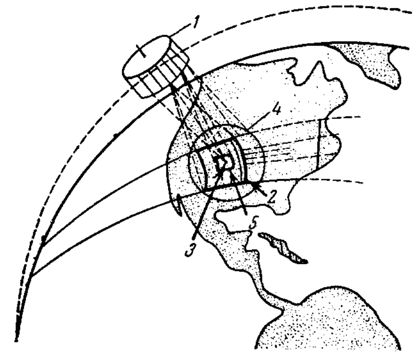

FIG. 2. Diagram illustrating television tracking of clouds from a satellite. 1 – Satellite; 2 – field of view of wide-angle camera; 3 – field of view of camera with narrow angle; 4 – line of scanning with receiver of infrared radiation.

1000 km, could yield information on processes occurring in an equatorial belt extending from 30° N to 30° S. The use of an orbit approximately 35,000 km high is also of interest. A satellite moving in such an orbit would be located at all time above the same point of the earth's surface. A system of such satellites, located over different geographical regions, can most effectively gather and transmit information from other meteorological satellites.

Having examined the general problems connected with the use of meteorological satellites, we now turn to a discussion of various specific problems that constitute the program of meteorological research with the aid of satellites.

1. TELEVISION TRACKING OF CLOUDS

As already noted above, television tracking of clouds with satellites is one of the most attractive possibilities of using satellites for meteorological observations. Even the first attempts made in this direction met fully with these expectations.

The first experimental investigation of the distribution of clouds was with the aid of the Vanguard II satellite, launched on February 17, 1959 (see references 11 and 12). The Vanguard II, a sphere 50.8 cm in diameter weighing 9.8 kg, was launched on an elliptical orbit with 628 km perigee and an 3380 km apogee. The inclination of the orbit to the plane of the equator was 33.2°. Two photocells 1 and 5 (Fig. 1) were installed on the satellite, sensitive to the spectral region from 0.6 to 0.8μ and subtending an angle of approximately 1°. The photocells were mounted in such a way, that their receiving apertures were at an angle of 45° to the satellite's axis of rotation. It was assumed that as the satellite rotated the photocells would be alternately aimed on the earth and would scan the distribution of brightness over the earth's surface, similar to television scanning of a picture. Owing to the high albedo of the clouds and the lower albedo of the land and sea surfaces (with the exception, naturally, of regions covered with snow or ice), the clouds can be identified by the brightness maxima. The actual motion of the satellite was found to be so complicated, that the measured values of the brightness could not be localized at individual points. This made the experiment with Vanguard II a failure.

On the other hand, the first attempt at television tracking of clouds, accomplished with the TIROS I satellite launched on April 1, 1960, was quite successful (the name TIROS stands for Television and Infra Red Observation with Satellites). The characteristics of TIROS I are described in detail in references 23 and 27. We point out only the most essential data (with approximate conversion of the English measures and weights into metric ones). The satellite, a cylinder 102 cm in diameter and 47 cm high, weighed approximately 110 kg. It was launched in an almost circular orbit with 750 km apogee, 700 km perigee, and an initial period of revolution of 99.15 minutes. The inclination of the orbit was 48.3°, and consequently the satellite moved in a latitude band from 50° south to 50° north. Two television cameras were installed on the TIROS I. One had a field of view 218 × 1290 km, with a resolution on the order of 1.6 km. The other camera (with a smaller angle of view), provided a resolution of about 0.5 km. The image had 600 lines and was either transmitted directly (whenever the satellite was above the reception station) or recorded on magnetic tape and transmitted during the passage of the satellite over the point of reception. Since the memory capacity was low, only 32 photographs could be recorded on the tape. The transmission time of this information was 3.5 minutes. Only the illuminated part of the northern hemisphere was photographed.

Nickel cadmium batteries were used as the power source on the satellite. The entire lateral surface of the satellite was covered with solar cells used to charge the nickel-cadmium batteries (the number of solar cells was 9200).

The TIROS I was an unoriented satellite. On going into orbit, the third stage of the rocket rotated together with the satellite, for stabilization purposes, with a speed of 125 rpm. After going into orbit, the "anti-spin" mechanism decreased the speed of rotation to 12 rpm. The direction of the axis of rotation should remain constant in space. The television cameras were aimed parallel to the axis of rotation in opposite sides and thus looked on the earth alternately only during certain parts of the orbit. The orientation of the optical axes of the cameras was calculated in such a way that the cameras were aimed at the earth under conditions of best illumination. It was found in practice that the direction of the axis of rotation of the satellite does not remain constant, owing to perturbations produced by the earth's magnetic and gravitational fields (see reference 9). However, provision was made on the satellite for a special control of the orientation with the aid of the scanning device, which fixed the outline of the earth in interplanetary space by the infrared radiation from the planet.

Figure 2 shows a scheme for television tracking of clouds from a satellite. Here regions 2 and 3 are the fields of view of the television cameras, and the circle 4 indicates the horizon fixed by the scanning device. Region 5 is subtended by the non-scanning radiation receiver, with which it is proposed to measure the infrared radiation from the planet (these measurements were not made with TIROS I). During the time when the pictures received by the earth-based apparatus were satisfactory (approximately three months), more than 20,000 photographs of different portions of the earth's surface were obtained. The main difficulty in the analysis of the resultant photographs was in the construction of a grid of geographic coordinates for the localization of the resultant pictures in space. It was necessary here not only to find on the photograph the image of some identifiable detail of the earth's surface, but also to take into account the distortion in the image, due to the optical and the electronic systems.

The initial analysis of the results obtained was made in two ways.[13,14,27,31] On the one hand, individual photographs were analyzed, particularly in those cases when the viewed territory contained large-scale whirlwinds or cyclone storms, with diameters on the order of 500 — 1000 km. On the other hand, by selecting a large number of photographs (on the order of 30), "mosaics" were made up, characterizing the distribution of the clouds over a very large territory.

An unexpected result of the analysis of the first photographs obtained with the aid of TIROS I was the detection of large scale ordered systems of clouds.

It has been known for a long time that tropical hurricanes are characterized by the presence of spiral-like systems of clouds. It was found, however, that the unique spiral cloud systems, connected with the strong

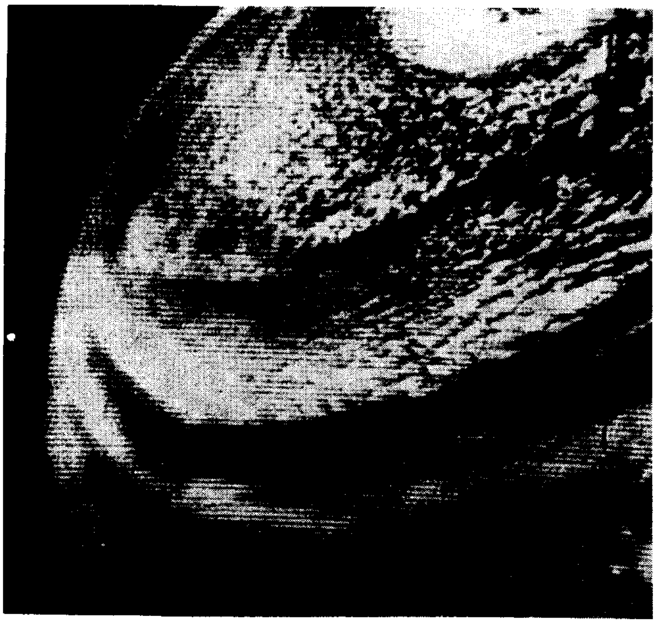

FIG. 3. Photograph of cloud distribution in the vicinity of a storm in the Pacific.

non-tropical gales and having characteristic dimensions up to 1500 km and above, are also one of the characteristic features of the distribution of clouds over the earth's surface. Figure 3 shows an example of such a spiral system of clouds, observed on April 4, 1960 some 1300 km to the west of Southern California (the bright areas represent the clouds). This photograph shows the large-scale cloud system clearly. The existence in the atmosphere of different-scale movements is also quite clearly manifest, for narrow belts, consisting of individual elements, can be discerned inside the wide large-scale cloud belts. It is probable that at higher resolution these elements could be broken up into even smaller formations. It is important to note that the photograph shown in Fig. 3 was taken over the Pacific main. The only data available for that day are meteorological observations made by several ships, and provide no reliable check on the synoptic situation. A photograph from a satellite provides a better estimate of the state of the atmosphere and can serve as a good illustration of the value of meteorological observations made with the aid of satellites.

Although spiral cloud systems are a general characteristic feature of non-tropical gales, there is still a noticeable difference in the cloud distribution in different cases. Thus, for example, gales on continents show no such fine and sharply outlined cloud belts as do oceanic gales. Continental gales are characterized by extensive cloud-cover areas, with the gaps between the cloud belts either weakly pronounced or missing completely (an example of such a gale in Nebraska, on April 1, 1960, is shown in Fig. 4). The spiral form of the cloud is more likely to be produced in this case by a wedge of dry cloudless air, which bends in the

FIG. 4. Photograph of distribution of cloudiness in the region of a storm in Nebraska.

direction of the center of the gale (upper right part of Fig. 4). A comparison of the photographs shown in Figs. 3 and 4 enables us to judge the extent to which the clouds distributions in various gales exhibit essential specific features. This means that an analysis of these features on the photographs can yield valuable information on the character of the meteorological processes.

The use of the second method of analysis of photographs, by compilation of "mosaics" of the distribution of the clouds over large territories, makes it possible to gain an idea of the synoptic situation over territories comparable with the dimensions of the earth's sphere. Using such mosaics, Fritz and Wexler[14] made an attempt to construct maps of the geographical distribution of the cloud cover over an extensive latitude belt from the west coast of the USA to the central part of the Atlantic, and also from the center of the Atlantic to the Near East. These data, however, are still merely illustrative.

Investigations of the distribution of cloudiness by television tracking from satellites have merely begun. It is therefore natural that many problems in this field are still unsolved at the present time. Thus, the question of how to differentiate between clouds, snow, and ice on a photograph has not yet been satisfactorily answered. Although visual criteria can be sufficiently reliable in this case (see Sec. 30), nonetheless the need for developing objective methods is beyond dispute. An important role may be played here by measurements of the polarization of the reflected light, since radiation scattered by clouds differs appreciably from that from snow (or ice).

The problem of selecting the strata and forms of the clouds is of undoubted importance. A particularly important fact is that only the upper layers of stratified clouds are photographed from the satellite.

It is very important to determine the altitude of the upper boundary of the photographed clouds. To solve this problem one can apparently measure the intensity of the outgoing radiation in the ultraviolet and near infrared regions of the spectrum (see reference 10). It is obvious that the almost-Rayleigh scattering in the atmospheric layer above the clouds causes the intensity of the outgoing ultraviolet radiation to depend considerably on the thickness of the outer layer, and consequently, on the position of the upper boundary of the clouds. For infrared radiation (outside the atmospheric absorption bands) the analogous dependence of the intensity on the thickness of the scattering layer will be much less pronounced, since the scattering is relatively small in this case. Were the reflection of radiation by the clouds non-selective, then the ratio of the intensities of the radiation in the ultraviolet and in the near infrared regions of the spectrum would be primarily a function of the thickness of the layer between the cloud and the radiation receiver. It should be noted that the degree of polarization of the light in the case of Rayleigh scattering also changes appreciably, depending on the thickness of the scattering layer. Consequently, polarization measurements can yield additional data on the height of the upper boundary of the clouds. The problem of the nighttime distribution of the clouds (on the shadow side of the earth) is not yet fully solved, although a suggestion has been made that measurements of infrared radiation from the atmosphere and the clouds be used for this purpose.

The foregoing unsolved problems are merely a sample of what needs to be investigated in the future. There exist, naturally, very many other serious problems.

2. MEASUREMENTS OF THE RADIATION BALANCE OF THE "EARTH'S SURFACE-ATMOSPHERE" SYSTEM

While the clouds are an indicator of the processes that take place in the atmosphere, the radiation balance of the "earth's surface plus atmosphere" system determines the energy variations of the atmosphere (we shall henceforth refer to it, for brevity, as the radiation balance of the earth). It is obvious that an investigation of the influx and outflow of heat in the atmosphere is a problem of very great importance. This is why numerous attempts have been made to calculate the earth's radiation balance (see reference 5). Many difficulties connected with similar calculations bring to the forefront the problem of an experimental investigation of the radiation balance. Since it is exceedingly important in this case to obtain information covering the entire planet, satellites are the most suitable means for such research.

Let us recall the equation of radiation balance of the "earth's surface–atmosphere" system

$$R_s = Q_0(1 - A_s) - F_\infty. \qquad (1)$$

Here Q_0 is the influx of solar radiation from outside the atmosphere, A_S is the albedo of the "earth's surface–atmosphere" system, and F_∞ is the outgoing radiation (the thermal radiation of the "earth's surface–atmosphere" system).

At the present time, the accuracy of the indirect methods used to determine on the earth's surface the solar constant amounts to several percent. The satellite measurements of the radiation flux can hardly be more accurate. Therefore the influx of solar radiation can be calculated from the well-known value of the solar constant (we recall that the latter is 2 cal/cm^2 min). Thus, what need be measured are the albedo of the planet A_S and the outgoing radiation F_∞ (we note, however, that the foregoing does not mean at all that the measurements of the solar constant from satellites are meaningless; such measurements can be very important if made with sufficient accuracy). If Q_0 is assumed constant, then to determine the albedo A_S it is enough to measure the outgoing short-wave radiation Q_∞ (i.e., the radiation reflected by the earth into space). In this case $A_S = Q_\infty/Q_0$.

The same basic apparatus can be used to measure the radiation balance and its components from satellites and from the earth's surface. Measurements on satellites, however, have certain specific features. In particular, if the satellite is not oriented relative to the earth, it is necessary in the design of the apparatus to take into account, or to eliminate, the effect of the continuous variation of the orientation of the receiving surfaces. It was precisely this consideration that guided V. Suomi[8,28] in the construction of instruments for the measurement of the earth's radiation balance from satellites.

Initially Suomi constructed a spherical radiation receiver.[8] Naturally, in this case the orientation of the receiver relative to the earth and the sun should influence its readings the least. It was proposed to install spherical radiation receivers on the ends of the satellite antennas. To be able to measure independently the individual components of the radiation balance, the surfaces of the spheres had different optical characteristics. Two receivers (Fig. 5) were in the form of thin silver spheres 3.12 cm in diameter, one blackened on the outside and the other coated with white paint (for the short-wave radiation). Two other receivers were smaller, both coated black, and provided with screens to shade them against the sun's radiation. The quantity measured directly was the temperature of the spheres. One series of temperature measure-

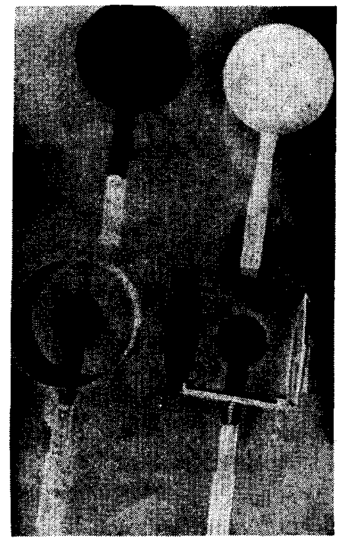

FIG. 5. Spherical radiation receivers.

ments of four spheres and of one control value should last 30 seconds, so that during the time of one revolution of the satellite around the earth approximately 180 series of measurements can be performed. Each series covers a region extending approximately 240 km in space. If it is assumed that the overall duration of the observations is 80 days, the total number of observation series should be 230,000.

Let us see what connection can be found between the measured values of the temperature of the spherical radiation receivers and the components of the radiation balance. We denote by r the radius of the sphere, and by a_S and a_L its absorptivity for short and long wave radiation, respectively. In this case we have the following approximate equation for the heat balance of the sphere when the satellite is over the illuminated side of the earth:

$$\pi r^2 (a_S S_0 + a_S D_\infty + a_L F_\infty) = 4\pi r^2 a_L \sigma T^4, \qquad (2)$$

where S_0 is the solar constant, D_∞ and F_∞ are the fluxes of the outgoing short-wave and thermal radiation, and T is the surface temperature of the sphere. It is assumed here that the radiation receiver is not shaded by the satellite itself.

In the case of a black sphere it can be assumed that $a_S = a_L = 0.95$. Denoting by T_b the temperature of the black sphere, we have in accordance with (2)

$$0.95S_0 + 0.95D_\infty + 0.95F_\infty = 4\cdot 0.95\sigma T_b. \qquad (3)$$

If the sphere is painted white and reflects the short-wave radiation, say with $a_S = 0.05$, but absorbs the thermal radiation ($a_L = 0.95$), we obtain in lieu of (2)

$$0.05S_0 + 0.05D_\infty + 0.95F_\infty = 4\cdot 0.95\sigma T_w^4, \qquad (4)$$

where T_W is the temperature of the white sphere.

If the solar constant S_0 is assumed known, the last two equations are sufficient to determine D_∞ and F_∞. Thus, for example, for the outgoing short-wave radiation we obtain

$$D_\infty = \frac{4 \cdot 0.95}{0.90} \sigma (T_b^4 - T_w^4) - S_0. \quad (5)$$

This formula is not very convenient to use, particularly for the calculation of the outgoing short-wave radiation when the zenith angle of the sun is large and the albedo of the underlying surface low, when D_∞ in (5) is a small difference of two much larger quantities [naturally, the error resulting from calculations by means of (5) will be considerable in this case]. A much more reliable procedure is to use a radiation receiver with optical properties much different from the black or white receiver (insensitive to thermal radiation). If, in addition, such a receiver is shaded from the sun, its temperature will be determined principally by the outgoing short-wave radiation. This is precisely why Suomi proposed to use spherical receivers shaded from the sun (Fig. 5), with optical properties $a_S = 0.90$ and $a_L = 0.10$. A combination of two types of screens (ring and disks) ensures shading of one of the two receivers at all satellite positions relative to the sun. The problem of determining the components of the radiation balance of the earth is solved in this case through the use of an indeterminate system of three equations for two unknown quantities. In this case it is obviously possible to determine independently the solar constant, if these three equations are regarded as equations for three independent quantities.

If the satellite is in the earth's shadow ($S_0 = D_\infty = 0$), then the temperature of any of the radiation receivers is determined only by the outgoing radiation F_∞.

The first attempt at measuring the radiation balance was made with the aid of the Explorer VII satellite, launched on October 13, 1959. This satellite weighed approximately 37 kg and was launched on an orbit with 550 km perigee and 1080 km apogee (the initial period of revolution was 101.2 minutes). The transmitter of the satellite failed the same month, and consequently the observations were relatively short-lived. The central part of the Explorer VII was a circular cylinder, bearing hemispherical (instead of the initially proposed spherical) radiation receivers on its lateral surface. The lack of published results of this investigation makes it impossible to ascertain the conclusions derived from the analysis of the observations. Suomi merely reported[28] that the measurements of the outgoing radiation F_∞ (on the dark side of the earth) were in satisfactory agreement with the mean-latitudinal distribution of the outgoing radiation calculated by G. Hooton (see reference 8).

The procedure considered above for processing the results of the measurements of the radiation balance is simplified in the extreme. We must certainly allow for parasitic "illumination" of the above-described receivers by rays reflected from the satellite hull. There is likewise no doubt that the assumption that the radiation absorbed by the sphere and that absorbed by a flat surface of size equal to the transverse cross section of the sphere, which is the basis of Eq. (2) and other formulas, is only approximately valid. Real coatings (lampblack, magnesium, etc.) always have a reflectivity that depends on the angle of incidence of the radiation (i.e., they are not orthotropic). Therefore to determine the amount of radiation absorbed by the sphere it is necessary to take into account the non-orthotropic nature of its surface and consequently the specific geometry of the radiation receiver. Finally, there is still another principal difficulty in the interpretation of the results of measurements from satellites, regardless of the type of radiation receiver employed.

A radiation receiver mounted on a satellite must remain at a very large altitude above the earth's surface (it is known that a satellite with perigee less than 200 km cannot last long). Therefore the receiving surface of the radiation receiver "sees" tremendous areas of the earth's surface. Thus, for example, with a satellite altitude of approximately 1,000 km, the field of view subtended by the instrument is approximately 7% of the entire surface of the earth. An averaging scale of this magnitude is too large. In addition, it must be borne in mind that, depending on the specific nature of the particular problem, information may be needed on the components of the earth's radiation balance averaged over areas of different sizes.

We note, incidentally, that if we are interested in the transformation of the radiation fluxes by the atmosphere it is enough merely to measure the integral fluxes D_∞ and F_∞ at altitudes on the order of 20 – 30 km. All the remaining (upper) part of the atmosphere can be regarded as optically empty (diathermic) with respect to the transformation of the integral radiation fluxes. All this raises the problem of finding the connection between the radiation balance or its components, measured from a satellite, and the corresponding values obtainable by measurements at lower altitudes. In other words, we seek a connection between the values of the radiation balance or its components obtained at different altitudes in the atmosphere.

There is, however, another approach to the solution of this problem. If we limit the angle of view of the radiation receivers so that it subtends a relatively small portion of the earth's surface, then we can average the measured quantities over small areas even when the satellite altitude is large. Let $I(\vartheta, \psi)$ be a function characterizing the angular distribution of the intensity of the radiation, reaching the satellite (ϑ is the angle between the directions of the ray and the nadir and ψ is the azimuth angle). Then an instrument covering a solid angle Ω will measure the following radiation flux:

$$F_\Omega(h) = \int_{(\Omega)} I(\vartheta, \psi) \cos\vartheta \, d\Omega. \quad (6)$$

Here h is the height of the satellite over the earth's surface. We denote by A the area of the earth's surface subtended by the solid angle Ω. In this case one can assume that radiation reaches the receiver from the elementary area A within the limits of the solid angle Ω. In practice we are interested in the hemispherical ascending radiation flux from the aforementioned elementary area, i.e., the influx of radiation from this area onto an open horizontal surface, located at a certain height h_1. Let us denote this quantity by $F(h_1)$. It is obvious that $F(h_1) \neq F_\Omega(h)$. Thus, it is necessary to find the connection between these quantities. Only then will the problem of determining the sought hemispherical radiation flux $F(h_1)$ be solved.

It is important here that the "localization" of the flux $F(h_1)$ (the determination of the altitude h_1) is not a trivial problem. It is quite clear that h_1 cannot be defined as the altitude of the point for which the contour of the elementary area A coincides with the line of the visible horizon (we denote the altitude of this point by h_1'). It is easy to understand that we should always have $h_1 > h_1'$. In the case of a flat earth (which corresponds to low values of h) an instrument with an unscreened flat receiving surface always "sees" the earth's surface up to the true horizon. Nonetheless, the dimensions of the elementary area whose radiation is being measured will increase with increasing altitude of the instrument. This is caused by the fact that the main contribution to the influx of radiation on the receiving surface is made by the portions of the earth's surface located near the nadir. It follows therefore that the quantity A should be understood to mean a certain "effective" area, which contributes the bulk of the radiation. Naturally, the effective area should always be less than the field of view of the instrument, and consequently, in the case of a spherical earth, the inequality $h_1 > h_1'$ should hold.

Let us proceed now from the foregoing qualitative considerations to a quantitative analysis, based on the use of some results obtained by London, Oyama, and Viebrock.[22] Assume that the satellite is located at the point A on a circular polar orbit (Fig. 6). The instrument has its unscreened flat receiving surface facing the earth (the shaded strip on Fig. 6 corresponds to that part of the atmosphere in which the radiation fluxes are appreciably transformed; from this point of view the upper boundary of this region can be considered to be the top of the atmosphere). In this case the receiving surface "sees" the earth's surface within the limits of a cone, the generatrix of which makes an angle Φ_m with the vertical. The radiation flux $F_S(\theta)$ incident on the receiving surface is

$$F_s(\theta) = \int_{-\Phi_m}^{\Phi_m} I_s(\theta, \Phi) \cos \Phi \, d\Phi, \tag{7}$$

where $I_S(\theta, \Phi)$ is the intensity of the radiation reaching the satellite. We note that we are considering in this case a two dimensional model of the earth.

As can be seen from Fig. 6

$$\sin \Phi_m = \frac{r}{R}. \tag{8}$$

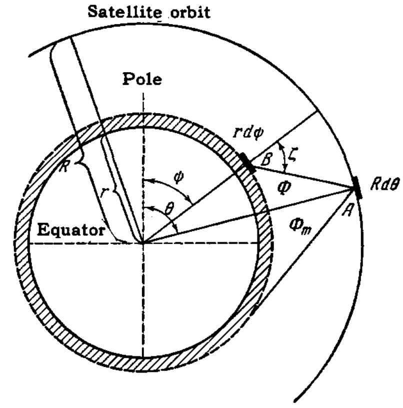

FIG. 6. Illustrating measurements of the radiation balance with satellites.

If, for example, the satellite has an altitude of 700 km, then $\Phi_m = 64°$. Let $F(\psi)$ denote, further, the radiation flux of interest to us (but not measured directly) at the level of the arbitrarily assumed upper limit of the atmosphere (at the point B). Since the thickness of the atmospheric layer that transforms the radiation fluxes is small, we have

$$F(\psi) = \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} I(\psi, \zeta) \cos \zeta \, d\zeta. \tag{9}$$

The problem now consists of finding the connection between the radiation fluxes $F(\theta)$ and $F(\psi)$. It is obvious that when $r = R$ the identity $F(\theta) \equiv F(\psi)$ holds true. If $R \geq r$, the following relation should apply:

$$I_s(\theta, \Phi) = I(\psi, \zeta). \tag{10}$$

We note, however, that this relation is practically meaningful only if it is assumed that the atmosphere is spherically symmetrical (in this case it indicates that the structure and composition of the atmosphere are independent of the latitude).

Let us write down the following relations, which follow from an examination of Fig. 6:

$$\frac{\sin \Phi}{r} = \frac{\sin \zeta}{R} = \frac{\sin(\theta - \psi)}{\varrho}, \tag{11}$$

$$\varrho = R \cos \Phi - r \cos \zeta. \tag{12}$$

According to (11)

$$\sin \Phi = \frac{r}{R} \sin \zeta = \sin \Phi_m \cdot \sin \zeta.$$

Differentiating the last relation, we obtain

$$\cos \Phi \, d\Phi = \sin \Phi_m \cos \zeta \, d\zeta. \tag{13}$$

From (11) and (12) it follows also that

$$\sin(\theta - \psi) = \frac{\varrho}{R} \sin \zeta = (\cos \Phi - \sin \Phi_m \cdot \cos \zeta) \sin \zeta$$

$$= (\sqrt{1 - \sin^2 \Phi_m \cdot \sin^2 \zeta} - \sin \Phi_m \cdot \cos \zeta) \sin \zeta.$$

This means that the right half of the last relation is a function of ζ only. We introduce therefore the notation

$$\theta - \psi = a(\zeta), \qquad (14)$$

where

$$a(\zeta) = \arcsin\left[\left(\sqrt{1-\sin^2\Phi_m \cdot \sin^2\zeta} - \sin\Phi_m \cdot \cos\zeta\right)\sin\zeta\right].$$

We now find, with allowance for (10), (13), and (14), that

$$I_s(\theta, \Phi)\cos\Phi\, d\Phi = \sin\Phi_m \cdot I(\theta - a(\zeta), \zeta)\cos\zeta\, d\zeta.$$

We thus obtain instead of (7)

$$F_s(\theta) = \sin\Phi_m \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} I(\theta - a(\zeta), \zeta)\cos\zeta\, d\zeta. \qquad (15)$$

According to (9), we have for the radiation flux at the level of the arbitrary top of the atmosphere, at a latitude θ,

$$F(\theta) = \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} I(\theta, \zeta)\cos\zeta\, d\zeta. \qquad (16)$$

The last two formulas provide the sought connection between the radiation fluxes at different altitudes in the atmosphere. For specific estimates it is necessary, however, to have data on the angular distribution of the radiation intensity, otherwise the connection between $F_S(\theta)$ and $F(\theta)$ cannot be determined. Thus, satellite measurements of only the hemispherical radiation fluxes do not determine the radiation fluxes at other levels of interest to us. The solution of this problem can be obtained only from experimental data on the angular distribution of the intensity of the outgoing radiation over the disc of the planet. Even in this case it is necessary to assume in the data reduction that the atmosphere has spherical symmetry, something which is actually never true, owing to the inhomogeneity of the earth's surface and the horizontal inhomogeneity of the distribution of the clouds.

Quantitative estimates of the effect of the aforementioned factors on the relation between the radiation fluxes at different altitudes are very difficult, and have not been carried out so far. Thus, there is still no possibility of estimating the practical importance of these factors. If the character of the angular distribution of the radiation intensity does not influence the relation between the radiation fluxes $F_S(\theta)$ and $F(\theta)$ (this is realized if the radiation is isotropic, i.e., when $I = \text{const}$), they are related by the following extremely simple equation

$$F_s(\theta) = F(\theta)\frac{r^2}{R^2}. \qquad (17)$$

A somewhat more complicated formula is obtained when the radiation receivers are spherical

$$F_s(\theta) = F(\theta)\left(1 - \frac{\sqrt{R^2 - r^2}}{R}\right). \qquad (18)$$

In the derivation of this formula it is assumed that $F = \int I d\Omega$. We note also that formula (17) determines the radiation flux on a flat surface, the normal to which coincides with the direction to the earth's center. A more complicated case of arbitrary orientation was considered by T. Altschuler.[7]

Table II lists for the sake of illustration the calculated outgoing short-wave radiation D_∞ and thermal radiation F_∞ at the different altitudes in the atmosphere. These were calculated by P. Guest[3] for flat and spherical radiation receivers (in the latter case the amount of radiation per unit cross section of the sphere was calculated). All the quantities are in cal/cm^2 min. As can be seen from Table II, the effect of the so-called dilution (rarefaction) of radiation in this altitude interval is quite considerable. There is no doubt that this purely geometric effect, due to the decrease in the angle Φ_m with increasing altitude, should predominate over the influence of the character of the angular distribution of the radiation intensity on the vertical profile of the radiation fluxes, if a large range of altitudes is considered. Thus, to make approximate estimates of the radiation fluxes at different levels from quantities measured on a satellite, we can use the simple relation (17) or (18). It is obvious that the problem of recalculating the fluxes will be most difficult when the height of the satellite is relatively small (300 — 400 km), and the variation of the radiation fluxes with altitude is determined primarily not by the dilution effect, but by the influence of the angular distribution of the intensity of radiation.

The reasoning presented above refers to the connection between hemispherical radiation fluxes at different altitudes. It is obvious that the relation between the radiation flux within the confines of a small solid angle F_Ω and the hemispherical flux at the altitude of interest to us should be determined on the same basis.

3. THERMAL SOUNDING OF THE ATMOSPHERE

Detailed methods of determining the temperature of a body by its thermal radiation are well known from thermodynamics and are widely used in practice. By measuring the thermal radiation of the atmosphere

Table II

	Height (in kilometers)											
	320		480		560		1120		1600		3200	
	D_∞	F_∞	D_∞	F_∞	D_∞	F_∞	D_∞	F_∞	D_∞	F_∞	D_∞	F_∞
Plate	0,33	0,29	0,31	0,28	0,31	0,27	0,26	0,23	0,23	0,21	0,16	0,14
Sphere	0,50	0,45	0,46	0,41	0,44	0,39	0,34	0,31	0,29	0,26	0,18	0,16

with the aid of a receiver mounted on a satellite, it is possible analogously to determine the temperatures of different layers of the atmosphere. This idea was first proposed by J. King.[20,21]

The infrared absorption spectrum of the atmosphere has a very highly pronounced selectivity (see reference 5). Accordingly, the intensity of thermal radiation of the atmosphere depends appreciably on the wavelength. In those regions of the spectrum where the absorption of the infrared radiation is intense ($\lambda < 8\mu$, $\lambda > 12\mu$), only the thermal radiation from the outer layers of the atmosphere will reach the satellite. The main radiating components of the atmosphere are water vapor, carbon dioxide, and ozone. The concentration of all these gases beyond the stratosphere and mesosphere is negligibly small. Thus, in the region of the strong absorption, the radiation will come from different layers of the stratosphere and mesosphere. It is obvious that the thickness of the radiating layer should depend on the intensity of absorption in the portion of the spectrum under consideration. In the region of weak absorption (for example the $8-12\mu$ "atmospheric window") the satellite will receive a "mixture" of radiation from the earth's surface and from a thick layer of the atmosphere. In this case the thermal radiation is a complicated function of the temperature of the earth's surface and of the stratification of the atmosphere.

It is easy to show[5] that the intensity of a monochromatic thermal radiation of frequency ν, passing in the direction of the zenith angle through the upper limit of the atmosphere (in this case it can be assumed that this arbitrary upper limit is approximately 100 km high) is determined by the relation

$$I_\nu(0,\ \mu) = \int_0^\infty E_\nu(\tau)\, e^{-\frac{\tau}{\mu}} \frac{d\tau}{\mu}. \tag{19}$$

Here τ is the optical thickness of the atmosphere, measured from the top of the latter, $\mu = \cos\vartheta$, while $E_\nu(\tau)$ is the intensity of the monochromatic radiation of an absolutely black body at the level τ and is a function of the temperature at this level (Planck's function).

Relation (19) illustrates graphically the fact that the intensity of thermal radiation is a function of the stratification of the atmosphere. From the mathematical point of view, the problem of interest to us reduces to inverting the integral (19), i.e., solving a Fredholm integral equation of the first kind. Let us examine the method proposed by King[21] for the solution of this problem, based on the use of the Volterra approximate method.

We represent Planck's function in the form of the following series:

$$E_\nu(\tau) = \sum_{i=1}^{n} a_i S_i, \tag{20}$$

where

$$S_i = \begin{cases} 1 \text{ in the interval } \tau_{i-1} < \tau < \tau_i, \\ 0 \text{ in the remaining part of the atmosphere.} \end{cases} \tag{21}$$

Here a_i are numerical coefficients, and the summation over the index i is equivalent to a breakdown of the atmosphere into n layers.

Substituting (20) in (19) we obtain

$$I_\nu(0,\ \mu) = \sum_{i=1}^{n} a_i l_i, \tag{22}$$

where

$$l_i = e^{-\frac{\tau_{i-1}}{\mu}} - e^{-\frac{\tau_i}{\mu}}. \tag{23}$$

We choose n values of μ, equal to $\mu = \mu_j$ ($j = 1, 2, \ldots, n$). Putting $I_j = I(0, \mu_j)$, we now write (22) in the form of a set of n equations for the coefficients a_i

$$I_j = \sum_{i=0}^{n} a_i l_{ij}, \qquad j = 1, 2, \ldots, n. \tag{24}$$

Let us extend now the resultant relation to the case of non-monochromatic radiation contained in a frequency band $\Delta\nu$. In this case we have instead of (19) (see reference 5)

$$I(0,\ \mu) = -\int_0^\infty E(\tau) \frac{dP}{d\tau} d\tau, \tag{25}$$

where P is the transmission function, characterizing the fraction of the radiation transmitted by the given absorbing layer (this quantity can be defined also as the probability that a photon emitted in the band $\Delta\nu$ at an optical depth τ will emerge through the boundary of the layer without experiencing absorption).

Substitution of (20) in (25) yields

$$I(0,\ \mu) = \sum_{i=1}^{n} a_i (P_{i-1} - P_i). \tag{26}$$

In analogy with the foregoing, we obtain next

$$I_j = \sum_{i=1}^{n} a_i l_{ij}, \qquad j = 1,\ 2,\ \ldots,\ n, \tag{27}$$

where

$$l_{ij} = P(\tau_{i-1},\ \mu_j) - P(\tau_i,\ \mu_j). \tag{28}$$

Solving Eq. (27) with respect to the coefficient a_i, we can then calculate Planck's function $E_\nu(\tau)$ from the known values of a_i and by using formula (20). Using the well known expression for Planck's function

$$E_\nu(\tau) = \frac{\frac{2h\nu^3}{c^2}}{\exp\frac{h\nu}{kT(\tau)} - 1},$$

we can readily calculate then the dependence of the temperature on the optical thickness $T(\tau)$. This solves our problem.

Obviously we must know in this case the transmission function and the vertical distribution of the concentrations of the substances absorbing and emitting the radiation. Unfortunately, King did not calculate even one specific example to illustrate the effectiveness of the proposed method.

Since the program for satellite measurement of infrared radiation, necessary for the solution of this problem, has not yet been completed, the application of the method described will become feasible only in the future. The foregoing problem can be solved also by measuring the infrared radiation with receivers mounted on rockets or high-altitude balloons. As is well known,[2] measurements of infrared radiation with high-altitude rockets have already been made in the U.S.S.R. According to King's data[21] measurements with high-altitude balloons are planned in the U.S.A., using an infrared photometer with interference filters that transmit a portion of the spectrum near the following wavelengths: 6μ (center of the absorption band of water vapor), 9.6μ (center of the absorption band of ozone), 11μ (center of the "atmospheric window"), and 15μ (center of the absorption band of carbon dioxide).

The idea of thermal sounding of the atmosphere by measuring the outgoing radiation in different regions of the spectrum was recently discussed in references 15, 17, 18, and 32. Kaplan[17,18] proposed to measure for this purpose the outgoing radiation in the 15μ region (carbon-dioxide band) for spectral segments about 5 cm^{-1} wide. The essential advantage of measurements in the region of the carbon-dioxide band is that in this case it is not difficult to specify the concentration of the atmospheric component that absorbs and emits the radiation. It is known that the volume concentration of carbon dioxide in the atmosphere changes very little and amounts to 0.03% on the average. By determining the temperature stratification from measurements of the outgoing radiation in the CO_2 band, we can then use the results of analogous measurements in the water-vapor absorption region to determine the vertical distribution of the concentration of water vapor.

Later on, Kaplan[18] calculated an example to illustrate the possible practical application of his method. The intensities of the ascending thermal radiation were calculated at a level of 50 millibars for a large aggregate of vertical temperature distributions, specified for the 7 isobaric surface levels: 1,000 (earth's surface), 700, 400, 300, 200, and 50 millibars. Inasmuch as the vertical distribution of the temperature is characterized in this case by the values of the latter at the seven levels, it was sufficient to calculate the intensity of radiation for seven frequencies (the concentration was assumed to be 0.26 "cm"/millibar). From a table of the radiation intensities I_{ν_0} calculated for different stratifications, and from the measured values I_ν of the outgoing radiation for the seven frequencies, it is possible to determine the temperature distribution for which $\sum\limits_{(\nu)} (I_\nu - I_{\nu_0})^2$ is a minimum.

For a more accurate determination of the "best" distribution of the temperature, Kaplan proposed also to use the formalism of perturbation theory. By carrying out a series of Gedanken experiments, Kaplan found that in many cases the agreement between the "true" distribution of the temperature and the outgoing radiation determined from "measurement" data was fully satisfactory. In other cases, however, in spite of the idealized character of the calculations, the agreement was found to be unsatisfactory. In the presence of heavy cloudiness, the upper boundary of the cloud layer can be assumed opaque. Obviously, in this case the measurements can give information on the temperature and height of the upper boundary of the clouds. Under real conditions of partial cloudiness and in the absence of sharp transitions from the cloudy and cloudless parts of the atmosphere, the interpretation of the measured outgoing radiation is possible only if the clouds are also tracked simultaneously by television. Estimates made by Kaplan have shown that an essential source of measurement errors can be noise in the electronic devices. Even at a 30:1 signal-to-noise ratio the errors due to the noise reach $3-4°$.

The main difficulty in thermal sounding of the atmosphere with the aid of measurements of the thermal radiation lies in the need of proving that a unique solution of the problem is possible. It is quite obvious that measurement of the outgoing radiation in relatively broad spectral regions in which the absorption coefficient varies greatly will not give the required results, for in this case the "radiating layer" will be too extensive and too varible. This conclusion can be verified by the distribution of the energy in the spectrum of outgoing radiation, as calculated by Kondrat'ev and Yakushevskaya.[6] They investigated the dependence of the spectral composition of the outgoing radiation on the stratification of the atmosphere. From the intensities $I_{\Delta\lambda}$ of the outgoing radiation in different parts of the spectrum, they calculated the effective temperatures using the relation $I_{\Delta\lambda} = (\sigma/\pi)\, T_{eff}^4$ (σ is the Stefan-Boltzmann constant). If the temperature at a given level z is equal to the effective one, then obviously the radiation in this region of the spectrum is generated principally near the level z. If, furthermore, the effective temperature calculated for a definite region of the spectrum corresponds for any stratification to the true temperature at one and the same level z, this means that the outgoing radiation is a single-valued characteristic of the temperature of the atmosphere at the given level. By accumulating a set of values of $I_{\Delta\lambda}$ for different regions of the spectrum, it is possible to determine in this case the temperature at different levels, and consequently, the vertical distribution of the temperature.

Table III. Effective temperatures of the "earth's surface – atmosphere" system

Portion of the spectrum, μ	Stratification III		Stratification II		Stratification I	
	t_{eff} °C	p, millibar	t_{eff} °C	p, millibar	t_{eff} °C	p, millibar
1,08—1.20 (H_2O; 1,1 μ)	12.8	825	—17.2	780	—30,2	890
1,25—1.38 (H_2O; 1,38 μ)	12,3	820	—17,9	765	—30.9	865
1,38—1.50 (H_2O; 1,38 μ; CO_2; 1.4 μ)	12.0	815	—18.2	750	—31,0	870
1,50—1,54 (H_2O; 1.38 μ)	11.8	810	—18.1	760	—30,9	875
1.54—1,67 (CO_2; 1.6 μ)	16.9	950	—17,0	1000	—30,0	900
1,70—1,92 (H_2O; 1.87 μ)	17.0	1000	—19,0	755	—31.0	870
1.92—2,08 (H_2O; 1.87 μ; CO_2; 2,0 μ)	9,5	780	—20.2	735	—32,3	840
2,08—2.15 (CO_2; 2,0 μ)	15.5	885	—18,0	765	—31.1	870
2,27—2.63 (H_2O; 2.7 μ)	3.2	710	—22.5	710	—34,0	800
2.63—2,87 (H_2O; 2.7 μ; CO_2; 2.7 μ)	—9,8	545	—32,5	595	—42.4	595
2,87—2,99 (H_2O; 2.7 μ)	3,2	710	—23,4	700	—34,1	800
2.99—3.57 (H_2O; 3,2 μ)	9,0	755	—18,3	760	—31,0	870
4,00—4.63 (CO_2; 4,3 μ)	1,6	700	—27.5	650	—38,2	690
4.88—8.70 (H_2O; 6,3 μ)	—11,4	525	—28.5	640	—38.0	700
8,70—9.09 (H_2O)	16,0	900	—17,0	1000	—30,3	890
10,55—12,28 (H_2O)	15.7	895	—17,1	1000	—30,0	900
12—13 (H_2O; CO_2; 15 μ)	11,2	800	—19.0	755	—31.2	835
13—14 »	—15.5	485	—36,0	555	—47.5	520
14—15 »	—47,7	200	—54.2	415	—61,2	310
15—16 »	—47,2	205	—53,0	435	—60,0	325
16—17 »	—20,8	440	—36,1	550	—46.2	540
17—18 »	—20,0	445	—30,8	615	—38,0	700

Table III lists the results of the calculations of the effective temperature, given in reference 6 for three different stratifications of the atmosphere, and lists also the altitudes corresponding to the same values of the air temperature. It is seen from Table III that the level of the true temperature of the air, equal in magnitude to the effective temperature, fluctuates over a rather wide range. This means that in this case it is impossible to establish a sufficiently single-valued correspondence between the outgoing radiation in some region of the spectrum and the temperature of the air at a definite level in the atmosphere.

The calculations carried out in reference 6 pertain to relatively broad regions of the spectrum. It is important to ascertain in this connection whether other results will be obtained from calculations of the outgoing radiation for narrower portions of the spectrum. Greenfield and Kellog[15] answer this question in the negative. They calculated for different portions of the spectrum the vertical distribution of the contributions made to the outgoing radiation by atmospheric layers corresponding to a pressure difference of 50 millibars. The calculations were made for the region of the atmospheric "window" 8 – 13 μ, and also for the wavelengths 6.0 and 6.2 μ in the intense absorption band of water vapor (in the last two cases, spectral segments 0.1 μ wide were considered). The calculations for the "window" have shown that the main contribution to the outgoing radiation is made here by the earth's surface. However, the fraction of the radiation from the tropo-

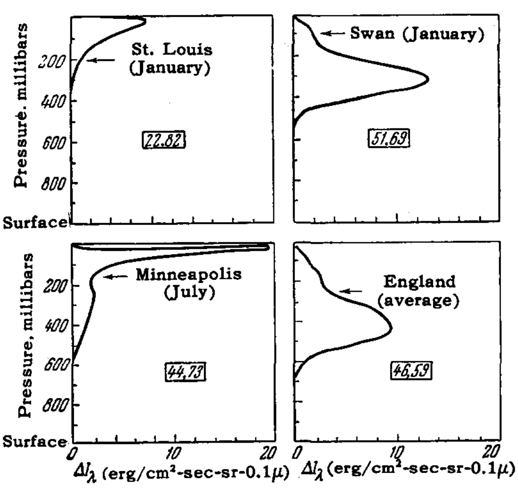

FIG. 7. Outgoing radiation for $\lambda = 6.0\ \mu$ (erg/cm²-sec-sr-0.1μ).

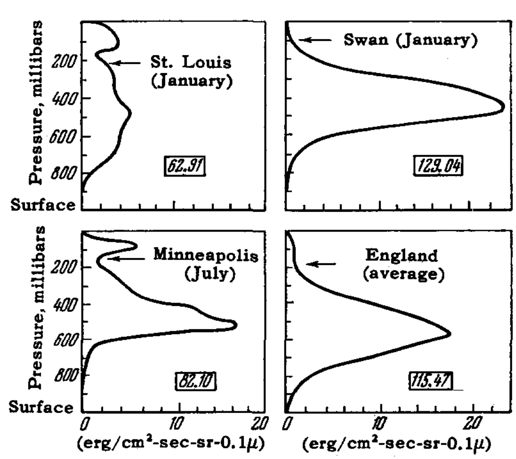

FIG. 8. Outgoing radiation for $\lambda = 6.2\ \mu$ (erg/cm²-sec-sr-0.1μ).

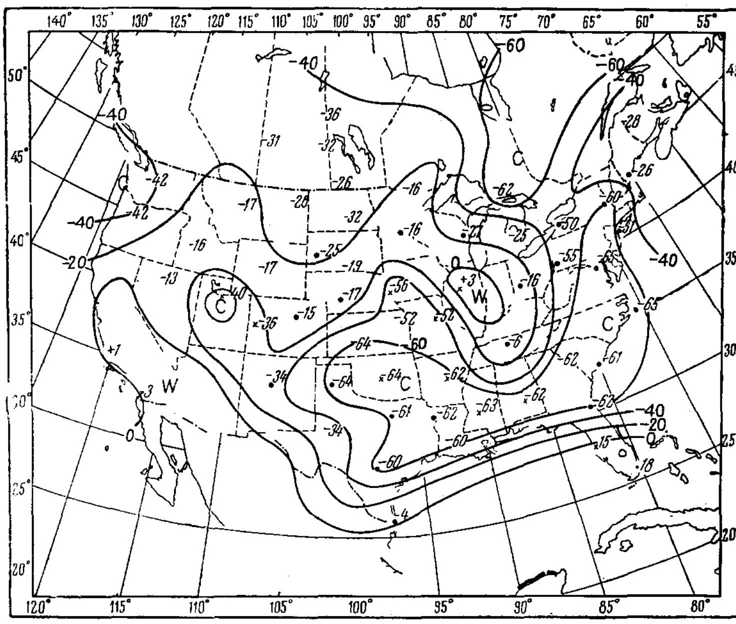

FIG. 9. Geographic distribution of effective temperature (12 noon, February 9, 1959).

sphere is also considerable and ranges from 1/3 to 2/3, depending on the stratification of the atmosphere. The results of calculations for 6.0 and 6.2μ are illustrated in Figs. 7 and 8. The numbers in the rectangle characterize here the total outgoing radiation. The arrows designate the position of the tropopause. An examination of this figure shows that even for the spectrum segments only 0.1μ wide the outgoing radiation is not a single-valued characteristic of the temperature at any level: the vertical distribution of the contribution from different layers of the atmosphere to the outgoing radiation is highly variable and depends on the stratification of the atmosphere. It is important to emphasize that these results pertain to a cloudless atmosphere. Naturally, the presence of clouds aggravates the situation.

Thus, the available quantitative estimates of the possibility of thermal sounding by measuring the outgoing thermal radiation in different regions of the spectrum have not yet yielded the expected results. The main difficulty in solving this problem lies in the strongly pronounced selectivity of the thermal radiation of the atmosphere. The half-widths of the absorption lines (and accordingly of the emission lines) near the earth's surface are less than 0.1 cm^{-1} and decrease with altitude in proportion to the pressure. The influence of selectivity can therefore be small only in limitingly narrow portions of the spectrum, on the order of $10^{-2} - 10^{-3}$ cm^{-1} in width. Unfortunately, measurements of the thermal radiation for such narrow portions of the spectrum are still not feasible.

Although the prospects of determining the vertical distribution of the temperature by measuring the outgoing radiation in different regions of the spectrum hold out little hope at present, this does not mean that measurements of this kind are of no interest at all. To the contrary, the Wexler's results[32] show that the geographic distribution of the outgoing radiation in different regions of the spectrum, obtained by satellite observations, can be a source of very important information on the horizontal inhomogeneity of the temperature field. Particularly graphic results were obtained in the analysis of the geographic distribution of the outgoing radiation in the region of the $8-13\,\mu$ atmospheric "window".

Figure 9 shows the results of Wexler's calculation of the outgoing radiation in the "window" from data of aerological soundings in many points of the U.S.A. and Canada at 12 noon on February 9, 1959. The synoptic surface map for this date is shown in Fig. 10. As can be seen from Fig. 9, the lines of equal effective temperature yield very interesting information on the processes in the atmosphere. The cold center in the northwest is due to the presence of high clouds in the region of a gale in the northwest of the Pacific. Another cold center over the state of Utah is connected with the passage of a cyclone in the region of Colorado. The deep minimum of the effective temperature at the northeast of the U.S.A. reflects the presence there of an extensive territory with upper-layer cloudiness. The wave on the frontal surface in the Missouri region (Fig. 10) can be interpreted with the aid of the map of Fig. 9 as a northern penetration of the cold surface in this region. Many other interesting features that follow from the analysis of Figs. 9 and 10 can also be noted.

Since the regions of heat and cold are at the same time zones of ascending and descending motion, an analysis of the geographic distribution of the effective temperatures can be useful also from the point of view of tracing the large-scale fields of vertical motions.

To conclude the foregoing, it must be stated that at the present time the results of investigations of the

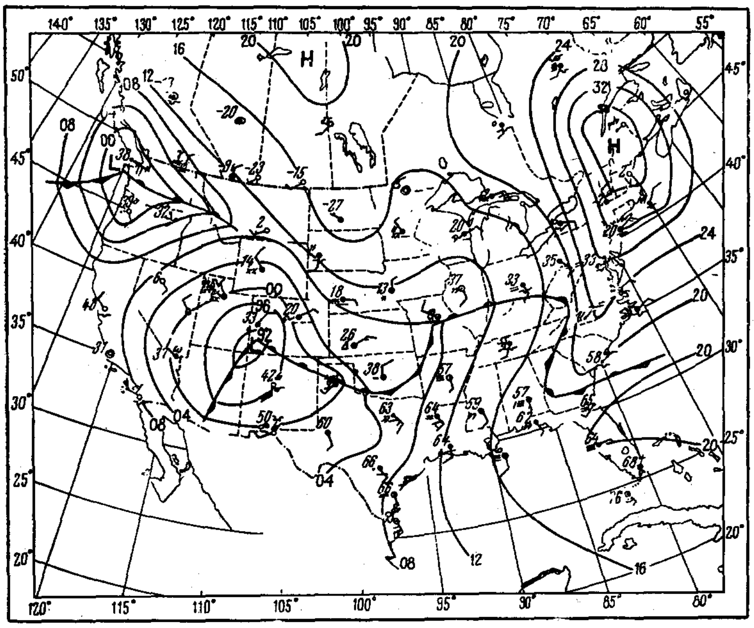

FIG. 10. Synoptic surface map at 12 noon, February 9, 1959.

infrared radiation of the "earth's surface–atmosphere" system can apparently be used most effectively to determine the horizontal thermal inhomogeneity of the earth's surface and the atmosphere. From this point of view, such data can serve as a very valuable supplement to the materials obtained by television tracking of the clouds.

4. RADIOMETEOROLOGICAL INVESTIGATIONS

In principle it is possible to carry out various types of radiometeorological researches with satellites. The greatest interest, however, attaches apparently to the possibility of using centimeter radar to investigate the distribution of precipitation zones. By now radar has become commonplace in meteorological practice. Procedures for radar investigations of cloudiness and precipitation have been quite satisfactorily developed. This gives grounds for assuming that the tracking of precipitation zones with the aid of satellite-borne radar is not far off. Naturally, radar investigations from satellites have many peculiarities, both of operation and observing conditions. Let us examine, following a recent paper by Keigler and Krawitz,[19] the most essential specific features of radar observations of precipitation zones with the aid of satellites. We leave aside such details as weight, dimensions, or power consumption, which are determined by the capabilities of the rocket used to launch the satellite.

Inasmuch as the main problem in radar research is to track precipitation zones at various altitudes, it is necessary to discuss first the region in space that can be covered by radar soundings from a satellite. It is readily seen that the horizontal extent and depth of this region are determined by the aggregate of the following factors: the desired vertical resolution of sounding, the interval between sounding pulses, the permissible information storage and transmission capacity. Let us explain this.

The vertical-sounding resolution depends on the height of the satellite, on the angular beamwidth of radiation (the width of the principal lobe of the antenna directivity pattern), the length of the pulse, and the angle of the direction of the radiation relative to the nadir. Calculations show that the most important factor is the width of the directivity pattern. The requirements imposed on this characteristic of the radar must be rather stringent here. Thus, for example, at a satellite altitude $h = 500$ km and a pulse duration $\Delta t = 1$ sec, the resolving power Δz is less than 1 km at angles $\theta < 40°$ relative to the nadir, if the sounding angular beamwidth is $\beta = 0.1°$. If $\beta = 0.5°$, $\theta = 40°$, and the remaining quantities are the same, we get $\Delta z = 5$ km, which obviously is unacceptable. It is clear therefore that highly directional UHF radiation must be used to scan the atmosphere at angles $\theta < 40°$.

Since the satellite travels at a high altitude, it is very important to take account of the pulse travel time to the sounded element and back. Obviously, the travel time must be shorter than the time of rotation of the antenna from one sounded element to the next. According to reference 19, at a beamwidth $\beta = 1°$ and a satellite altitude $h = 500$ km, the ratio of time the antenna takes to rotate through the beam width to the travel time becomes less than unity when $\theta > 35°$ (the dependence of this ratio on the satellite altitude h is very weakly pronounced in this case). Consequently, the scanning angle to the nadir should not exceed 35°. This circumstance greatly limits the dimensions of the sounded region, since from an altitude of 500 km it is possible, in principle, to scan up to an angle of 125°. Even when the satellite is about 2,000 km high, scanning to the limiting angle θ is still impossible. If $\beta = 0.1°$, the aforementioned limitations are completely eliminated, but the total duration of the scanning process and the volume of the obtained information are greatly increased. Estimates made in reference 19 show that with the best apparatus and with a relatively broad beam ($\beta = 0.5°$, which at a satellite altitude of approximately 1,000 km corresponds to a horizontal resolution of about 10 km at the earth's surface), the volume of the obtained information, with allowance for the use of the limiting value of θ at the equator, amounts to approximately 2×10^7 elements.

Thus, the foregoing considerations show that while radar sounding of extensive portions of the atmosphere from satellites is possible in principle, many factors make this procedure difficult in practice. Let us turn now to a clarification of the specific nature of the physical scope of the radar sounding problem in this case.

The main specific feature of the problem is that in earth-based radar investigations the clouds and precipitation are observed against the sky as a background, but when the clouds are sounded from above reflections from the earth's surface play an important role. In addition, as in earth-based observations, it is necessary to take into account the attenuation of the reflected signal along the path from the target to the radar. As is well known, the power of radio echoes from rain depends on many factors (intensity of the rain, the pulsed power of the transmitter, dimensions of the antenna etc.), and is in particular directly dependent on the frequency of the sounding pulse. Therefore, from the point of view of obtaining maximum reflection, it is expedient to use short wavelengths. Thus, a change from $\lambda = 3.2$ cm to $\lambda = 0.9$ cm increases the intensity of the reflected signal by approximately two orders of magnitude. On the other hand, a decrease in the wavelength entails an increase in the signal attenuation along the path to the target and back. Therefore the optimal wavelength should be determined by the most suitable ratio between the intensity of reflection and the degree of attenuation of the signal.

It is particularly important to account for the attenuation of a signal reflected from the earth's surface and passing through a zone of precipitation and clouds. The reflecting ability of the earth's surface is several orders of magnitude greater than the reflecting ability of the precipitation, and varies widely with the type of underlying surface. Therefore, in spite of the fact that the signal reflected by the earth is greatly attenuated as it returns through the atmosphere (particularly in the presence of precipitation), the power of the radio echo signal from the earth's surface exceeds the power of the "useful" radio echo signal from the precipitation zone (we note that reflection from the earth's surface is useful in the sense that it enables us to determine

Table IV. Characteristics of simplest radar station for satellites

Frequency	10 000 Mc
Transmitter pulse power	80 kw
Duration of sounding poles	5 μsec
Working cycle	10^{-6}
Receiver noise	10 db
Area of sounded zone	2×10^4 m^2 (at distance 1500 km)
Average power	36 w
Antenna	4′ parabola, beamwidth $\beta = 1.75°$
Weight, excluding antenna	20 kg

the distance from the precipitation zone to the earth's surface, instead of having to reckon the distances from the satellite). It is therefore necessary to develop special methods for discriminating between reflections from precipitation and reflections from the earth.

One possible method of such discrimination is a spatial separation of the reflection zones by fixing the arrival of the front of the reflected wave instead of merely registering the amplitude of the reflected signal. Another possibility is to use the spectral dependence of the reflection. As already noted, the reflection of radiation from precipitation is quite selective. On the other hand, reflection from the earth's surface can be regarded as neutral. Thus, simultaneous observation of reflections at two frequencies makes it possible to recognize reflection from precipitation, owing to the difference in the powers of the echo signals for different frequencies. Finally, to discriminate between reflections from precipitation and from the earth, we can use also such phenomena as the difference in the polarization of the reflected signals and the existence of a Doppler frequency shift for the signal from the precipitation zone, due to the vertical velocity of the precipitation relative to the earth's surface.

Summarizing the foregoing, we can state that an investigation of the distribution of precipitation zones with satellite-borne radar is quite difficult. A practical implementation of a complete program of such research will require extensive development of special radiometeorological apparatus. Naturally, the simplest apparatus should be tested and used in the first stage of the investigations. Table IV gives an example[19] of this type of simplest non-scanning apparatus (antenna directed to the nadir). The characteristics of all the units of this apparatus, including the weights and dimensions, make its use fully feasible even at the present time.

5. MEASUREMENTS OF VERTICAL DISTRIBUTION OF OZONE

Very much attention is being paid at the present time to investigations of atmospheric ozone. Numerous measurements have shown that ozone is a very

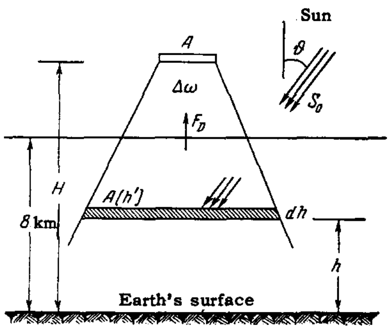

FIG. 11. Diagram showing measurement of scattered radiation on a satellite.

sensitive indicator of the dynamic processes in the stratosphere. It is also clear that ozone is an important factor in the thermal conditions of the stratosphere. As a rule, optical methods are used to investigate the overall content and the vertical distribution of the ozone in the atmosphere. During the IGY, a relatively extensive network of ozone-measuring stations was organized to investigate the laws governing the geographic distribution of ozone. However, in spite of the considerable progress in the research on atmospheric ozone, many problems still remain unclear at present, primarily because ozone-measuring observations cover only a small portion of the earth's surface. It is therefore quite understandable why the prospects of using satellites for ozone measurements are very enticing. It is also obvious that the only procedure suitable for this purpose is the optical method of determining the overall content and the vertical distribution of the ozone. This method of solving this problem was proposed by Singer and Wentworth,[24,25,26] whose results are reported here.

Let us imagine a radiation receiver facing the earth's surface, having an area A and subtending an angle $\Delta\omega$ (Fig. 11). Let the radiation receiver be at at a height H above the earth's surface and let the zenith distance of the sun be ϑ. We assume that it is possible to consider the scattering of the solar radiation in the atmosphere to be of the Rayleigh type and that the effect of multiple scattering can be neglected. Let us calculate the amount of radiation scattered by the atmosphere and picked up by the receiver.

We examine a certain elementary layer of the atmosphere dh at an altitude h above the earth's surface. We replace the real atmosphere by a homogeneous one 8 km high. We introduce further instead of the vertical coordinate h an analogous coordinate h', corresponding to the homogeneous atmosphere, (h' is the thickness of the layer of the homogeneous atmosphere, having the same mass as the layer of the real atmosphere of thickness h). We denote by $d_a = 8 - h'$ the thickness of the layer of the homogeneous atmosphere above the level h'. Let d_0 be the content of the ozone expressed in terms of the layer in centimeters, reduced to the normal pressure and atmosphere, above the level h'. In this case the intensity of the solar radiation S(h') at the level h' can be determined, with allowance for the attenuation due to the molecular scattering and absorption by the ozone, in the following fashion:

$$S(h') = S_0 \exp[-(k_s d_a + k_a d_0)\sec\vartheta], \tag{29}$$

where S_0 is the intensity of solar radiation outside the atmosphere, k_s the coefficient of molecular scattering, and k_a the coefficient of ozone absorption.

The amount of radiation scattered by the layer dh' will be

$$\Delta F(h') = A(h')\, S(h')\, k_s\, dh', \tag{30}$$

where A(h') is the area subtended by the solid angle $\Delta\omega$ at the level h'.

The fraction of radiation $\Delta F(h')$ scattered at an angle ϑ, in the case of Rayleigh scattering, amounts to $3(1+\cos^2\vartheta)/16\pi$ (see reference 4). Thus, the layer dh' scatters in the direction of the receiver (within the limits of the angle $\Delta\omega$) the following amount of radiation:

$$dF(h') = \frac{3}{16\pi} A\Delta\omega (1+\cos^2\vartheta)\, k_s\, S(h')\, dh'. \tag{31}$$

This formula takes into account the obvious relation $A(h')\,d\omega' = A\Delta\omega$, where $\Delta\omega'$ is the solid angle at which the radiation receiver (area A) is seen from the level h'. With allowance for the attenuation along the path from the level h' to the receiver, we have for the radiation incident on a unit area of receiving surface of the instrument

$$dF_D = \frac{dF(h')}{A} \exp[-(k_s d_a + k_a d_0)]. \tag{32}$$

Integrating (32) over all elementary layers dh', we obtain for the flux of scattered radiation on the receiving surface, with allowance for (29) and (31),

$$F_D = S_0 \Delta\omega \frac{3}{16\pi} k_s (1+\cos^2\vartheta) \int\limits_{h'=8}^{0} \exp[-(k_s d_a + k_a d_0)(1+\sec\vartheta)]\, dh'. \tag{33}$$

Calculation of F_D for the region of the Hartley absorption band (ozone absorption band in the wavelength interval approximately from 2200 to 3200 A) shows that in this case the main contribution to the scattering of the radiation is made by the stratosphere, and that the predominant factor is absorption of the radiation by the ozone. Table V lists the calculated altitude of the level above which 90% of the radiation reaching the receiver

Table V. Height (in kilometers) of lower limit of layer scattering the ultraviolet solar radiation

Wavelength (in A)	Zenith distance of the sun			k_a (cm^{-1})
	0°	30°	60°	
2800	41 (0.025)	42 (0.022)	44 (0.017)	46
3000	13 (0.311)	16 (0.294)	20 (0.218)	5

Table VI. Change in intensity of scattered ultraviolet radiation as a function of the vertical distribution of the ozone concentration

Wave-length (in A)	Altitude intervals (kilometers)				k_a (cm^{-1})
	30	21–30	16–21	11–16	
	k_α	k_β	k_γ	k_δ	
2800	−1,00	−0,006	0	0	46
3000	−0,45	−0,62	−0,18	−0,03	5

is scattered. These calculations have been made for different zenith distances of the sun and for two wavelengths, 2800 and 3000 A. The parentheses indicate the ozone contents (in "cm") above the corresponding lower limits of the "radiation layers."

It is seen from Table V that in the region near the maximum of the Hartley band the absorption by the ozone dominates so strongly, that almost all the scattered ultraviolet radiation comes from altitudes above 40 km. Even at $\lambda = 3000$ A (the skirt of the absorption band) the scattered radiation is generated only by the stratosphere. Thus, we can sound the layer of the ozone by measuring the scattered ultraviolet radiation at different wavelengths. This gives rise, however, to the question of the degree to which the scattered ultraviolet radiation is "sensitive" to changes in the vertical distribution of the ozone concentration. Table VI provides an answer to this question.

We denote by $\frac{\Delta\alpha}{\alpha}$, $\frac{\Delta\beta}{\beta}$, $\frac{\Delta\gamma}{\gamma}$ and $\frac{\Delta\delta}{\delta}$ the relative changes in the contents of ozone for the four layers of the atmosphere indicated in Table VI. We now represent the relative change in the flux of scattered radiation reaching the receiver, $\Delta F_D/F_D$, in the following fashion (it can be assumed that this representation is valid for small $\Delta F_D/F_D$):

$$\frac{\Delta F_D}{F_D} = k_\alpha \frac{\Delta\alpha}{\alpha} + k_\beta \frac{\Delta\beta}{\beta} + k_\gamma \frac{\Delta\gamma}{\gamma} + k_\delta \frac{\Delta\delta}{\delta}.$$

Table VI lists the values of the coefficients k_α, k_β, k_γ, and k_δ for these two wavelengths. The table shows clearly that the ultraviolet scattered radiation is quite sensitive to changes in the vertical distribution of the ozone concentration.

Although the results of reference 19 offer some promise, they must still be regarded only as tentative ideas. The weak spot in these calculations is the assumption that there is no aerosol scattering in the stratosphere. Investigations made recently show clearly that the aeorsol scattering of light in the atmosphere is quite considerable and, what is even more important in this case, highly variable.

Naturally, allowance for aerosol scattering will complicate to a great degree the possibility of using measurements of ultraviolet scattered radiation to derive information on the vertical distribution of the ozone. Apparently, the effect of multiple scattering can also be of no little significance. One must therefore think that, at least in the first stage, the most effective would be the method of determining the overall content of the ozone by the absorption of the ultraviolet solar radiation.

If the sun is located near the horizon relative to the satellite, then the sun's rays will pass through a layer of ozone on their path to the satellite. In this case we can determine the total contents of the ozone along the path of the rays by the well developed method of measuring the contents of ozone by absorption of the ultraviolet solar radiation. An example of a successful practical application of such a method is found in reference 1. In this investigation rocket measurements were made of the absorption of solar radiation by molecular oxygen in the region of vacuum ultraviolet at small elevations of the sun (the latter made it possible to obtain considerable values of absorption at high altitudes). The data of these measurements have yielded very valuable results relative to the vertical distribution of the concentration of the molecular oxygen. Obviously, an analogous procedure can be used also with satellites.

In conclusion we note once more that in the present article we considered briefly only certain problems of importance to meteorology, which can be solved with the aid of satellites. There is no doubt that the fast progress of the research in this field will uncover many wider prospects for meteorological investigations with the aid of satellites.

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