ENERGY BALANCE AND THE ROLES OF THE SUN, EARTH, OCEANS AND ATMOSPHERE

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Keywords: albedo, carbon dioxide, diffuse (scattered) radiation, direct solar radiation, effective terrestrial radiation, equinoxes, greenhouse effect, heat balance, latent heat transport, outgoing radiation, radiation balance, sensible heat transfer, radiation, "solar constant," solar radiation, turbulent (eddy) flux.

Contents

Summary

Physical processes occurring in the troposphere determine changes in weather and fundamentally affect the climate conditions in different regions of our planet. These processes include absorption of solar radiation, formation of outgoing long-wave radiation flux, the general atmospheric circulation, and the hydrologic cycle—particularly the formation of clouds and the initiation of precipitation. Solar radiation is the principal source of heat for the atmosphere. Heat energy is transferred from the surface to the atmosphere through long-wave radiation and turbulent heat exchange. The heat consumed by evaporation from the surface of land and water bodies is also released into the atmosphere in the process of water vapor condensation.

The algebraic sum of the radiative energy fluxes coming to and going from the surface of the Earth is called the radiation budget (balance) of the Earth's surface. It is evident that the value of the radiation budget is equal to the difference between the amount of direct and scattered short-wave radiation. The energy of the radiation budget of the Earth's surface is expended in various ways: heating the atmosphere by means of turbulent heat conductivity, evaporation, heat exchange with the deeper layers of the hydrosphere or lithosphere, and so on.

1. Introduction

The Sun is the star closest to the Earth and belongs to a class of yellow stars—dwarfs. The diameter of the Sun equals 1.4 million kilometres, and the mean distance between the Sun and the Earth is 149.5 million kilometres. As a result of nuclear reactions that occur in the Sun, the temperature of its surface is approximately 6000 °C, which causes the radiation of a considerable amount of energy from the Sun. The radiation coming from the Sun to the Earth is the only form of incoming radiant energy that determines the heat balance and the thermal regime of the Earth. Radiation energy incident upon the Earth from all other celestial bodies is so small that it does not noticeably influence the processes of heat exchange occurring upon the Earth. The spectrum of solar radiation corresponds to that of a black body, which is described by Planck's formula. In accordance with the temperature of the Sun's radiating surface, the maximum radiation flux is observed at the wavelength of approximately 0.45 µm, and most of the energy radiated by the Sun falls in the wavelength interval from 0.3 to 2.0 µm.

Moving away from the Sun, the intensity of radiation decreases in inverse proportion to the square of the distance. As the Earth moves around the Sun in its elliptical orbit, the intensity of solar radiation falling on the external boundary of the atmosphere changes during the year according to the change of distance between the Earth and the Sun. The smallest distance between the Earth and the Sun is observed at the beginning of January, when it equals 147 million kilometres. The greatest distance, reached at the beginning of July, equals 152 million kilometres.

2. Solar radiation

2.1. Solar constant

The flux of solar energy in one minute across an area of 1 cm^2 placed perpendicular to the Sun's rays, beyond the atmosphere, and located at the mean distance between the Earth and the Sun, is called the "solar constant". With the changes in the distance between the Earth and the Sun, the actual values of the solar energy flux at the outer boundary of the atmosphere vary from the value of the solar constant. This variation amounts to 3.5% (See *History of Atmospheric Composition*).

For a long time, the value of the "solar constant" was considered approximately equal to 1.9 cal cm^{-2} min^{-1}. It was then established that a noticeable amount of solar energy is absorbed in the upper layers of the atmosphere, and as a result the value of the solar constant was considered to be near 2.0 cal cm^{-2} min^{-1}. In organizing the International Geophysical Year, it was decided to consider the value of the solar constant to be equal to 1.98 cal cm^{-2} min^{-1}. Observations made recently give a slightly smaller value of the" solar constant", namely, 1.94 to 1.95 cal cm^{-2} min^{-1}.

We should indicate that for many investigations it is more important to know, not the true value of the solar constant, but its conditional value, corresponding to the amount of radiation that reaches the upper layers of the troposphere. In contemporary studies, this value of the conditional solar constant is usually considered equal to 1.90 to 1.92 cal cm^{-2} min^{-1}.

The question of the stability of the solar constant with time is of great interest. This question has been discussed in a number of studies and many authors consider that if changes in the "solar constant" do occur, they do not exceed the accuracy of the available measurements. In addition, it should be borne in mind that small variations of the "solar constant" of the order of 1% or even 0.1% may produce a certain influence on weather and climate. For this reason, the question of the stability of the solar constant deserves greater attention.

Knowing the value of the solar constant, one can calculate how much energy penetrates to the Earth's surface in different latitudes in the absence of atmospheric influences on radiation.

2.2. Direct and scattered radiation

In reality, the atmosphere is not a completely transparent medium for solar radiation. A noticeable part of the radiation coming from the Sun is absorbed and scattered in the atmosphere, and also is reflected into space. Solar radiation is especially influenced by clouds, but even under a cloudless sky solar radiation is changed greatly in the atmosphere. Solar radiation is absorbed in the atmosphere by water vapour, water droplets, ozone, carbon dioxide, and dust. Scattering of solar radiation is caused both by the molecules of air and by various admixtures—dust, water droplets and so forth. The flux of direct solar radiation that passes through the atmosphere depends on the transparency of the air and the Sun's elevation that defines the length of the path of the solar rays in the atmosphere .

As indicated above, in addition to changes of atmospheric transparency, cloudiness greatly influences the amount of direct radiation reaching the Earth's surface. With more or less dense clouds direct radiation is equal to zero. The greatest amount of direct radiation flux is recorded under cloudless sky and high atmospheric transparency. In these conditions, the flux of direct radiation incident on a perpendicular surface at sea level can amount to 1000 to 1200 Wm^{-2}. Average noon values of this flux usually equal 700 to 900 Wm^{-2}. With decreasing solar elevation during the daily cycle, the direct solar radiation is noticeably reduced as the optical thickness of the atmosphere increases

The amount of scattered radiation coming to the Earth's surface varies within wide limits, mainly as a function of cloudiness conditions and the Sun's elevation. The theoretical assessment of this radiation flux is rather complicated and does not yield accurate results. Available observational data show that, in many cases, the flux of scattered radiation is comparable in value with the flux of direct radiation incident on a horizontal surface. The greatest amounts of scattered radiation are recorded under cloudy skies. Scattered radiation is greatly influenced by the reflectivity of the Earth's surface. In particular, scattered radiation increases noticeably when the Earth's surface is covered with snow, that reflects a considerable amount of solar energy.

The general picture of the basic solar energy transformations in the geographical envelope of the Earth is as follows. The flux of solar energy at the mean distance between the Earth and the Sun is equal to the value of the solar constant. Because of the spherical shape of the Earth, a unit area at the top of the atmosphere gets, on average, one-fourth of

the total flux—about 340 Wm^{-2}, of which approximately 240 Wm^{-2} is absorbed by the surface of the Earth as a planet. The greater part of this amount is absorbed by the surface of the Earth and a considerably smaller fraction by the atmosphere.

2.3. Outgoing (long-wave) radiation

The Earth's surface, since it is heated due to absorbing solar radiation, is a source of outgoing radiation that transmits heat into the atmosphere. Atmospheric water vapour, dust and various gases that absorb long-wave radiation, detain the long-wave emission from the Earth's surface. Thus, a considerable portion of the long-wave radiation emitted from the Earth's surface is compensated for by the counter-radiation of the atmosphere. The difference between the Earth's surface emission itself and the atmospheric counter-radiation is called the net long-wave radiation. The value of net long-wave radiation is usually several times smaller than the flux of long-wave emission from the Earth's surface that would be observed if the atmosphere were completely transparent to outgoing radiation. On average for the whole surface of the Earth, the net long-wave radiation is much less than the absorbed short-wave radiation. This is a consequence of the so-called greenhouse effect, i.e. the result of the relatively high transparency of the atmosphere to short-wave radiation compared with that for outgoing emission (See *Greenhouse gases, Aerosols, and Ozone Layer*). Therefore, the mean radiation balance of the Earth's surface is a positive value.

2.4. The radiation balance

The energy in the radiation budget of the Earth's surface is expended in various ways: heating the atmosphere by means of turbulent heat conductivity, evaporation, heat exchange with the deeper layers of the hydrosphere or lithosphere, and so on. Terms describing all of the above forms of solar energy transformation must enter the equation of energy (heat) balance for the Earth's surface in addition to the algebraic sum of energy fluxes incident on and going from the Earth's surface. According to the law of energy conservation, this sum must equal zero.

The radiation and energy balances of the Earth's surface are definitely connected with the radiation and energy balances of the atmosphere. The Earth as a planet receives energy from space and emits it back only through radiation. As the mean temperature of the Earth changes little in time, it is evident that the radiation balance of the Earth (the difference between radiation absorbed and emitted into space) is equal to zero (See *History of Atmospheric Composition*).

The radiation budget of the atmosphere (equal to the difference between the radiation balances of the planet as a whole and that of the Earth's surface) must, therefore, on average, be negative and equal in absolute value to the Earth's surface radiation budget. In the atmospheric energy balance, the negative radiation budget is compensated for by an influx of energy from condensation of water vapour when forming clouds and precipitation and by an influx of heat from the Earth's surface due to the turbulent heat conductivity of the lower layers of the atmosphere. In addition to the processes of redistributing solar energy in the vertical direction, powerful processes of horizontal heat redistributing are developed in the geographical envelope. Most important among them is

the transport of thermal energy in the atmosphere and hydrosphere between high and low latitudes due to the heterogeneity of radiative heating of the spherical surface of the Earth. This transport is carried out through macroturbulent heat exchange and sensible heat transfer, as well as latent heat transfer through phase transformations of water.

The processes of solar energy transformation caused by radiative factors enumerated above, in turn significantly change the radiation regime, since this depends considerably on circulation in the atmosphere and hydrosphere, phase transformations of water, and so on. Cloudiness and coverage of the Earth's surface by snow (or ice) produce an especially pronounced effect on the radiation regime. In addition to the "first order" processes of transformations of solar energy which substantially change the radiation and heat regimes, a number of solar energy transformation processes requiring comparatively small amounts of energy and, therefore, not affecting the radiation and heat regimes directly, develop in the geographical envelope of the Earth. Some of these processes are very important for the formation of the biosphere. For example, the process of photosynthesis is based on the transformation of radiant energy into a comparatively stable form of chemical energy.

In studying all the forms of solar energy transformation in the geographical envelope, radiation and energy balance data are of prime importance. Data on the energy and radiation balances at the Earth's surface are especially important for investigating the biosphere since it is the main source of energy for biological processes.

The radiation balance of the Earth's surface R is equal to the difference between the absorbed solar radiation and the net long-wave radiation:

$$R = Q(1-\alpha) - I, \tag{1}$$

where Q is the total short-wave radiation (the sum of direct and scattered radiation), α is the albedo (the reflectivity of the Earth's surface for total radiation) expressed in fractions of unity and I is the net long-wave radiation, i.e. the difference between the radiation emitted by the Earth's surface itself and the absorbed counter-radiation of the atmosphere.

The radiation balance of the Earth-atmosphere system R_s, i.e. the radiation balance of the vertical column through the whole thickness of the atmosphere, may be determined in a similar way. We find in this case that

$$R_s = Q_s(1-\alpha_s) - I_s, \tag{2}$$

where Q_s is the solar radiation incident on the outer boundary of the atmosphere, α_s is the albedo of the Earth-atmosphere system and I_s is the long-wave emission from the outer boundary of the atmosphere into space (outgoing emission).

The radiation balance or budget of the atmosphere, R_a, equals the difference between the values of R_s and R. Using Equation (1) and (2), we get

$$R_a = Q_s(1-\alpha_s) - Q(1-\alpha) - (I_s - I) \tag{3}$$

Reflection of solar radiation from the Earth's surface is determined by the albedo, which is defined as the ratio of the amounts of reflected to incident radiation. Theoretically, the values of albedo might vary from unity for a completely white and reflective surface to zero for an absolutely black surface entirely absorbing the Sun's rays. Available observational data show that the albedo of natural surfaces does vary over a wide range, embracing almost the entire possible interval of values of reflective ability of various surfaces. In experimental investigations of solar radiation reflection, values of albedo for more or less all of the widespread natural surfaces have been found. These investigations have shown that the conditions of solar radiation absorption on land and water differ noticeably.

The conditions of solar radiation absorption by water bodies differ from those at the land surface. Clear water is comparatively transparent to short-wave radiation, and so the Sun's rays penetrating the upper layers of a water body are widely scattered and then to a considerable extent absorbed. It is easily understood that absorption of solar radiation in such conditions should depend on the solar elevation or zenith angle. If the Sun stands high, a great part of the incoming radiation penetrates the upper layers of water and is mainly absorbed there. At low solar elevation, the rays incident on the water surface at small angles are specularly reflected and do not penetrate into the depths of the water body. This leads to an abrupt increase of the albedo.

Both theoretical calculations and experimental investigations show that the albedo of water surfaces for direct radiation varies over a wide range as a result of this dependence on solar elevation. At large solar elevations the albedo for direct radiation does not exceed several hundredths, while with the Sun approaching the horizon the albedo reaches values of the order of several tenths. The albedo of water surface for scattered radiation varies much less and is about 0.10.

The albedo of the Earth-atmosphere system is of a more complicated nature than that of the Earth's surface. Solar radiation received by the atmosphere is partially reflected as a result of backscattering. Under a cloudy sky, a considerable fraction of radiation can be reflected from the upper surfaces of the clouds. With little cloudiness or under clear sky, the albedo of the Earth-atmosphere system depends essentially on the albedo of the Earth's surface. The albedo of clouds depends on the thickness of their layers, its mean value being equal to 0.40 to 0.50. The albedo of the Earth-atmosphere system without clouds is usually greater then that of the Earth's surface, except in those cases when the surface is covered with more or less clean snow.

Both the reflection of short-wave radiation and the emission of long-wave radiation are equally important forms of radiative energy loss. The basic features of the process of heat transfer by long-wave radiation can be expressed using the Stefan-Boltzmann law which states that the flux of radiation from an absolutely black body equals σT^4, where T is the absolute temperature in the Kelvin scale and σ is the Stefan constant, equal to $5.67 \cdot 10^{-8} \mathrm{Wm}^{-2}\mathrm{K}^{-4}$. Available experimental data show that radiation from the Earth's natural surfaces is, in general, rather close to radiation from a black body at the corresponding temperature, the ratio of the observed values of radiation to black-body radiation in most cases amounting to 0.90 to 1.00.

The maximum intensity of long-wave radiation from the Earth's surface is a function of wavelength and depends on the temperature of the radiating surface. The major part of the radiated energy occurs at wavelengths from five to several tens of micrometers. A considerable part of the long-wave radiation flux emitted by the Earth's surface is compensated for by the atmospheric counter-radiation. The long-wave emission of the atmosphere under clear sky is basically determined by the content of water vapour and carbon dioxide. The effect of atmospheric ozone is usually relatively less important. Water vapour, as well as carbon dioxide, absorbs long-wave radiation mainly in certain spectral bands so that the long-wave emission from the atmosphere is rather spectrally heterogeneous. The most intensive absorption of radiation by water vapour takes place in the range of wavelengths from 5 to 7.5 µm, while in the 9 to 12 µm spectral range the absorption of radiation is relatively small. Carbon dioxide has several bands of absorption, of which the band with wavelengths from 13 to 17 µm is of greatest importance. It should be mentioned that the atmospheric content of carbon dioxide varies comparatively little, while the amount of water vapour can be very different depending on meteorological conditions. Therefore, variations in air humidity are a substantial factor influencing radiation from the atmosphere.

Usually the atmospheric counter-radiation increases greatly with cloudiness. Low and middle level clouds are, as a rule, sufficiently dense so that they radiate like an absolutely black body at the corresponding temperature. Under such conditions, the net long-wave radiation is mainly defined by the difference between the temperature of Earth's surface and that of lower surface of the clouds. If this difference is small, the net radiation also approaches zero. High clouds, because of their low density, usually radiate less than a black body. Since the temperature of such clouds is comparatively low, their effects on net long-wave radiation are insignificant compared with the influences of low and middle clouds. Thus, the net long-wave radiation from the Earth's surface depends mainly on the surface temperature, the water vapour content in the air, and cloudiness. Due to these factors, the net long-wave radiation may vary from values close to zero up to several hundred watts per square metre.

Radiation from the Earth-atmosphere system is that portion of the surface emission that passes though the atmosphere, plus atmospheric radiation. Under clear sky, the Earth's surface emission at wavelengths from 9 to 12 µm plays a significant role in the radiation from the system. In conditions of continuous cloudiness, radiation from the top surface of clouds (which depends on the cloud-top temperature) is of principal importance. Since this temperature is usually much lower than the temperature of the Earth's surface, it is evident that cloudiness considerably reduces the output of heat to space through outgoing radiation.

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Biographical Sketch

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