ENERGY BALANCE OF THE EARTH
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INTRODUCTION

During the last three decades the development of scientific methods of short-range weather forecasting has been a central task in meteorology. Today the problem of understanding the climate and its fluctuations is becoming one of the main issues.

Some 50 years ago a scientist who tried to understand the behaviour of the atmosphere on the time scale of a few days must have felt the situation rather perplexing. The basic physical laws governing the changes in the atmosphere were known; the mathematical problem of weather forecasting had been formulated by N. Bjerknes as early as the beginning of this century. However, these laws were in a much too general form and could neither be solved nor that stage used to yield much information. One could make a long list of factors that were likely to be of some relevance for the development of changes in the atmosphere on time scales of a few days, but the more fundamental ones had not been identified. Today we are much wiser. We know, for example, that the essential content of the general equations for the purpose of short-range weather forecasting is in principle very simple (conservation of potential vorticity) and we have learned how to tackle the problem.

As regards the problem of climate and the fluctuations in the atmosphere for periods of one week or longer, we are again in the same bewildering situation. We can make a long list of factors that must be of some importance, but we certainly do not know which are the really important quantities (counterparts of potential vorticity) determining the behaviour of the earth-atmosphere system at these longer time scales. What are the proper questions to be asked? What observations should be made?

In this kind of situation a fundamental task is to obtain as clear a picture as possible about the way in which different physical balance requirements are fulfilled in the fluid envelope of the earth. A task of perhaps highest priority in this respect is the investigation of the energy balance of the earth.

Some of the basic long-term features of the global energy balance have been known for about 50 years. Recently, radiation measurements from satellites together with conventional data from the atmosphere and oceans have enabled us to study not only the long-term mean energy balance of our planet but also the seasonal cycle, which is the most conspicuous longer-term fluctuation in the atmosphere and the oceans. In this paper particular emphasis is placed on these new results, but before discussing them it is appropriate to review a few classical concepts of the annual-mean heat balance of the earth and to point out those factors for which the quantitative estimates today deviate considerably from those found in textbooks a few decades ago.

THE ANNUAL-MEAN HEAT BALANCE OF THE EARTH

The earth receives its energy from the sun. At the top of the atmosphere and at the mean solar distance, the rate of this energy supply, averaged for the whole surface of the earth, is 340 W m$^{-2}$ ( = $\frac{1}{4} \times$ solar constant). The numerical value of this basic quantity has still an uncertainty of about 1 per cent. There are many reasons for this uncertainty. One of these is that the determination of the solar constant has so far only been made from inside the atmosphere (either from mountain observatories as with measurements initiated by Abbot in the Smithsonian Institute in 1900, or from balloons like those made by Soviet scientists in the
1960s (Kondratyev 1972)). Needless to say, extrapolation of these measurements beyond the atmosphere involves quite a number of assumptions and diminishes the reliability of the results. Another reason for the uncertainty is that the solar constant may not be a constant but may slightly vary due to processes in the sun. For example, there is some (but not conclusive) evidence that it may vary with the well-known sunspot activity, the postulated range of variability being 2 to 2.5 per cent. In the future, monitoring the incident solar radiation from satellites will obviously be a task of high priority.

A part of the solar energy received by the earth is reflected back into space, principally by clouds, and does not in any way influence the processes on the earth.

The earth's surface and the atmosphere also radiate at a rate according to their temperature. Because the amount of energy on the earth does not vary much over periods of many years, the energy lost to space in the form of long-wave radiation must on average be very close to the amount of incoming solar radiation absorbed by the earth. Locally, however, this is not the case. Mainly due to the sphericity of the earth the lower latitudes receive more energy and are therefore bound to be warmer than the higher latitudes. Horizontal temperature differences cause horizontal pressure differences and these in turn imply some kind of circulation in the fluid envelope of the earth. The instantaneous state of the huge ventilation system that develops determines the distribution of weather; the statistics of these instantaneous states determines the climate. The physics requires that, on the average, radiation surplus (absorbed solar energy larger than outgoing terrestrial radiation) must be associated with higher temperatures (and thus with lower latitudes) and radiation deficit with lower temperatures (and thus with more polar latitudes). Furthermore, the ventilation system must be such that it carries energy from the areas of radiation surplus to those of radiation deficit.

Let us denote 340 W m\(^{-2}\) by 100 and briefly remind ourselves, in the light of the latest estimates, how these 100 units are used, on average, over the globe, and over a period of many years.

Figure 1 is a classical scheme showing the annual-mean energy budget for the earth as a whole. The left-hand side shows the budget of short-wave radiation, the middle part the budget of long-wave radiation and the right-hand side the exchange processes between the surface and the atmosphere. The first figure of this type was presented by Baur and Philipp (1935), although the first numerical values concerning the radiation budget of the planet earth were published prior to this by Simpson (1928).

According to the present view, the mean planetary albedo of the earth is somewhat less than 30 per cent (28 in Fig 1). The main contribution (19 units) comes from the clouds. Air dust and haze are estimated to contribute 6 units and the earth's surface about 3 units. The estimates of the planetary albedo have shown a steady decrease in the course of time: Simpson (1928) 43, Baur and Philipp (1935) 42, London (1957) 35, London and Sasamoni (1971) 33. The value 28 in Fig 1 is based on satellite measurements, which indicate that especially in tropical latitudes the albedo is smaller, and hence our planet is darker, than previously believed.

The planetary albedo is of course a highly variable quantity both in space and time. At present, we know practically nothing about the long-term (e.g. inter-annual) variability of cloudiness and planetary albedo. This kind of basic information will obviously be provided by satellite measurements in the coming years.

The atmosphere is relatively transparent to solar radiation. From the 72 units which are not reflected or back-scattered to space, only 25 units are absorbed by the atmosphere and the rest (47 units) by the underlying surface.
Fig 1. The annual-mean heat budget of the earth (from Schneider and Dennett, 1975)
In the region of long-wave radiation, the atmosphere absorbs practically all of the radiation emitted by the surface (the so-called green-house effect!), and the long-wave radiation going to space originates mainly from the cloud tops and the upper layers of the atmospheric water vapour and CO$_2$. Indeed, these have for a long time been known to be the major contributors to the terrestrial radiation and were already included by Simpson (1928) in his classical calculations.

The net radiation budget (short wave + long wave) of the atmosphere is negative (-29 units) and that of the surface positive by the same amount. For energy balance to exist there has to be a vertical energy transfer from the earth's surface to the atmosphere. This energy transfer is associated mainly with the transfer of water vapour (24 units) and only to a small extent with the direct transfer of sensible heat.

Baur and Philipps (1935) estimated the flux of sensible heat to be on average from the atmosphere to the surface. The reason for the wrong direction in the early estimates of this flux was due essentially to the misconception that, because the atmosphere is stably stratified on average, the average heat flux should be downward.

Figure 1 refers to the average long-term mean condition over the whole globe. The following discussion is concerned mainly with horizontal energy fluxes, the effects of which disappear when the earth is considered as a whole. For this purpose we need a scheme and equations describing the energy budget at a particular location.

**A SCHEME FOR THE LOCAL ENERGY BUDGET**

Consider a vertical column (Fig 2), which includes an air column extending from the top of the atmosphere down to the surface, and a column of underlying surface (which may be ocean (O), land (L), ice and snow (I) or all of them) down to a level where vertical energy fluxes can be neglected. The atmosphere contains internal energy (proportional to the temperature), potential energy (determined by the position of air particles in the earth's gravitational field), and kinetic energy. In addition, by virtue of its water vapour content, the atmosphere has a plentiful supply of latent heat that is released in condensation processes. The oceans also have these energy forms, except latent heat.

Let us denote (following Oort and Vonder Haar 1976) the rate of local change of total energy, or the storage of total energy, in the atmospheric part of the column by $S_A^t$ in the ocean part by $S_O$, in the land mass part by $S_L$, and in the ice part by $S_I$. Let us further denote the vertical energy flux by $F$; $F_A^t$ will then denote the downward energy flux at the top of the atmosphere, i.e., the radiation balance (incoming solar radiation - reflected solar radiation - long-wave radiation emitted to space). At the top of the atmosphere there are no other fluxes other than those associated with radiation and therefore

$$F_A^t = F_T$$

The downward energy flux at the bottom of the atmosphere, denoted by $F_{BA^t}$, consists of three parts

$$F_{BA^t} = F_{BA}^R + F_{BA}^{SH} + F_{BA}^{LH}$$

where $F_{BA}^R$ = the net downward radiation flux at the surface,

$F_{BA}^{SH}$ = the downward flux of sensible heat at the surface,

and $F_{BA}^{LH}$ = the downward flux of latent heat at the surface.
If we represent the horizontal transport of total energy in the atmosphere by $T_A$ and that in the oceans by $T_O$, we can write the energy balance for the entire column as

$$S_A + S_O + S_L + S_I = F_{TA} - \text{div} \ T_A - \text{div} \ T_O.$$  \hspace{1cm} \ldots 3

The energy balance for the atmospheric part is then

$$S_A = F_{TA} - F_{BA} - \text{div} \ T_A$$ \hspace{1cm} \ldots 4

and for the underlying surface

$$S_O + S_L + S_I = F_{BA} - \text{div} \ T_O.$$ \hspace{1cm} \ldots 5

Eqn 3 is naturally the sum of Eqns 4 and 5.

If we had good global observation covering the atmosphere, the hydrosphere (oceans and land waters), and cryosphere (ice and snow) of the earth, and if all these observations fitted nicely together in the framework of Eqns 3 to 5, we could say that we know how the energy balance of the earth is maintained.

For the time being, data from all subsystems are (in the global domain) more or less incomplete, and we do not know very accurately the long-term mean values of different terms in Eqns 3 to 5, even in the case where the volume in Fig 2 is a latitudinal zone. In particular, the horizontal heat transport in the oceans, $T_O$, is known only very poorly.

As I stated earlier, I shall concentrate mainly on the horizontal energy transports and, in particular, on the meridional energy fluxes, which are responsible for energy exchanges between different latitudinal zones in the atmosphere and oceans.
If we know \( F_{OA} \) and \( F_{BA} \), either from direct measurements or from indirect calculations, and if we can neglect the storage terms (for long-term mean conditions we are safe in doing so) we can determine \( T^0 \) and \( T_A \) by integrating Eqns 3 to 5 with respect to latitude and using the condition that all the mean meridional fluxes must vanish at the poles. The fluxes obtained in this way will be called required fluxes (i.e., those required by the heat sources and sinks as determined by \( F_{TA} \) and \( F_{RA} \)). On the other hand, we should in principle be able, by using the observed three-dimensional distributions of different variables, to calculate a more direct estimate for the fluxes; we shall call these observed fluxes. Historically, the 'required' meridional fluxes were the first to appear in the literature and, in fact, are still the only ones we have for the oceans. If our data and our methods of estimating different terms were good, 'required' fluxes should, of course, be equal to the 'observed' fluxes.

Because the 'required' meridional transport of energy in the oceans is one of the main items to be discussed later, it is appropriate to consider in more detail how one can proceed to get estimates of \( T^0 \). To do this, it is crucial to obtain accurate estimates of \( F_{BA} \), the net downward energy flux at the surface. One way of determining \( F_{BA} \) over ocean areas has been to obtain separate estimates of the three terms in the expression for \( F_{BA} \) (see Eqn 2). Some components of \( F_{BA}^R \) (e.g., the incoming solar radiation) can be measured although estimates of its geographical distribution depend heavily upon the climatological formulae. For estimation of \( F_{BA}^H \) over the oceans, climatological formulae exist that relate this flux to the mean wind speed and the mean temperature difference between air and water. Similarly, climatological formulae exist that relate \( F_{BA}^L \) to the mean wind speed and the mean vapour pressure difference between the water and the air.

Extensive calculations of the various terms in \( F_{BA} \) have been made, for example, by a group headed by Budyko in the Main Geophysical Observatory in Leningrad. The atlas of global distribution of the various components of \( F_{BA} \) prepared by this group (Budyko 1963) has been one of the most frequently cited publications in the climatological literature and most 'required' flux determinations are based on this atlas.

**ANNUAL-MEAN MERIDIONAL ENERGY TRANSPORT**

Figure 3 shows the annual-mean 'required' meridional energy fluxes on the globe. Considering first the total transport by the atmosphere and the oceans, \( T^0 + T_A \), we notice a poleward flux in both hemispheres with a quasi-symmetry with respect to the equator. The maximum poleward fluxes, \( (T^0 + T_A)^{max} \), are found close to 35° with a magnitude of about \( 5 \times 10^{15} \) W. The first determination of the required meridional energy fluxes was done by Simpson (1928), who estimated \( (T^0 + T_A)^{max} \) to be \( 4.1 \times 10^{15} \) W. The corresponding value obtained by Baur and Philipp (1935) was \( 4.4 \times 10^{15} \) W. Taking into account the fact that even today the uncertainty concerning the long-term mean value of \( (T^0 + T_A)^{max} \) is ± 10 per cent we notice that these old estimates by the pioneers in the area of the global energy balance were extremely good.

The value \( 5 \times 10^{15} \) W is not very informative as such and it may be useful to remember that the implied divergence/convergence of this flux could cool/warm the atmosphere in low/high latitudes at a rate that is -1°C day⁻¹. (It may also be of some interest to note that \( (T^0 + T_A)^{max} \) is 500 times (not more!) the rate of the present energy consumption by mankind \( (10 \times 10^{12} \) W).)
Fig. 3  Annual mean 'required' northward energy fluxes.

- total energy flux \( T_0 + T_A \)
- oceanic energy flux \( T_O \)
- latent heat flux \( L \)
- 'dynamic' energy flux \( (T_A) \)

(From Palmen and Newton 1969.)

Turning now to \( T_0 \) in Fig. 3, we notice a considerable poleward flux of energy in the oceans; in subtropical latitudes \( (10^\circ, 20^\circ) \) the oceans are seen to account for about a half of the total flux. There are many uncertainties involved in determination of the 'required' oceanic flux by the method described in the previous section. For example, the validity of the climatological formulae used in determination of the different components of \( F_{BA} \) is questionable. The value of \( F_{BA} \) is in many places a small residual between \( F_{BA}^R \) and \( (F_{BA}^{SH} + F_{BA}^{LH}) \) and therefore its estimation is liable to large relative errors. For these reasons, the \( T_0 \)-curve in Fig. 3 can be considered only as a rough estimate. The calculations for the southern hemisphere are particularly uncertain.

The difference between \( (T_0 + T_A) \) and \( T_A \) is \( T_0 \). The 'required' annual mean flux of latent heat in the atmosphere can be estimated from the annual mean distribution of precipitation and evaporation; this flux is given in Fig. 3 by the dotted line. It indicates in both hemispheres a considerable flux of latent heat from the sub-tropics, where the evaporation is larger than precipitation, towards the pole and the equator, where the contrary is true.

The rest of the energy flux must occur in the atmosphere in the form of 'dynamic' energy. We notice that the 'required' values for this quantity indicate two maxima, one in low and another in middle latitudes.
If the annual-mean 'required' fluxes are as shown in Fig 3, what about the 'observed' fluxes? As regards the atmospheric fluxes over the northern hemisphere we can say that a fair agreement exists between the estimates shown in Fig 3 and those estimates that are obtained from aerological observations. No corresponding comparison has been possible for the oceans.

Studies of the 'observed' atmospheric fluxes have also provided information concerning the mechanisms of the meridional energy transport in the atmosphere. It appears that in the meridional flux of all energy forms in the atmosphere, the mean meridional circulation predominates in low latitudes while different kinds of large-scale eddies are the primary mechanism in the middle and high latitudes.

Figure 3 roughly represents the state of our knowledge concerning the annual-mean energy fluxes at the end of the 1960s. The 1970s have brought into the picture radiation data from satellites and new compilations of conventional meteorological and oceanographic observations.

Meteorological satellites have provided the opportunity to measure directly some of the components of the net radiation at the top of the atmosphere ($F_{TA}$). So far one has not been able to measure and monitor the total incoming solar radiation but one can measure the part of solar radiation reflected by the earth, and also the long-wave radiation emitted to space by the earth's surface and the atmosphere. If one assumes the value of the solar constant to be known, the net radiation at the top of the atmosphere can be determined from these measurements. Large difficulties are encountered in processing the raw observations and the work is very laborious. For example, degradation of the sensors in space and the possible diurnal bias for satellites in sun synchronous orbits create problems, and there exist temporal and spatial sampling gaps in the data. Nevertheless, the estimates of the net radiation $F_{TA}$ obtained from satellite measurements are independent of the earlier estimates of this quantity.

Figure 4 shows the 'required' total energy transport by the atmosphere and the oceans, $T_0 + T_A$, for the annual-mean conditions over the northern hemisphere. These were determined by Vonder Haar and Oort (1973) from the satellite measurements of the net radiation at the top of the atmosphere ($F_{TA}$) during 1962-1970. One should not be led to assume that the monitoring of the net radiation $F_{NA}$ is already continuous; due to many difficulties associated with the satellite measurements only a part of the raw data can be used. Figure 4 also shows the 'observed' energy transport in the atmosphere ($T_A$) determined from aerological observations throughout the period 1958-1962. Because the storage terms can be neglected for the long-term annual-mean conditions, the oceanic heat transport can be determined thus: $T_0 = (T_0 + T_A) - T_A$. This estimate of $T_0$ does not therefore include the rather inaccurate determinations of the components of $F_{TA}$ from climatological observations. An estimate for the error limits for $T_0$ is indicated in the figure.

It appears that the new estimate of $F_{NA}$ is considerably larger than earlier ones (see Fig 4). The main reason for this is that satellite observations indicate a smaller albedo in the tropics than previous conventional observations, and hence more absorption of solar energy takes place in the tropics and there is consequently a larger total transport across, say 30°N.

**SEASONAL VARIATION OF THE ENERGY BUDGET IN DIFFERENT LATIDINAL ZONES OVER THE NORTHERN HEMISPHERE**

As regards the long-term annual-mean meridional energy fluxes over the northern hemisphere, Fig 4 represents the present level of our knowledge. But what can be said about the energy budget of the fluctuations that occur on different time scales both in the atmosphere and the oceans?

Oort and Vonder Haar (1976) have studied the energy budget of the seasonal variation and some results from their study are given below.
Fig 4  New estimates of the annual-mean poleward energy fluxes over the northern hemisphere.

- 'required' total energy transport \( (T_A + T_Q) \), determined from the satellite measurements of the net radiation \( F_{TA} \) at the top of the atmosphere.
- 'observed' energy transport in the atmosphere \( T_A \).
- 'required' energy transport in the oceans (with uncertainty limits denoted by shading).
- Earlier estimate of the 'required' energy transport in the oceans.

(from Vonder Haar and Oort 1973)

Fig 5  Net radiation at the top of the atmosphere \( F_{TA} \) (watt m\(^{-2}\)).

(from Oort and Vonder Haar 1976).
Figure 5 shows the mean seasonal variation of $F_{TA}$, based on the satellite observations during the period 1964-1971, and indicates that practically the whole hemisphere is heated ($F_{TA} > 0$) during the summer and cooled ($F_{TA} < 0$) during the winter.

Figure 6 shows the rate of storage of energy in the atmosphere ($S_A$) and in the oceans ($S_O$). It is seen that the atmospheric energy content increases ($S_A > 0$) from the beginning of February to the end of July and decreases during the rest of the year, the amplitude of $S_A$ being largest in the high latitudes. Whereas $S_A$ can be determined relatively accurately from observations, $S_O$ cannot, although this is clearly the more important of the two terms. Nevertheless, the estimates of $S_O$ given in Fig 6 are based on the largest set of oceanic temperature data ever used in similar studies.

Figure 7 shows the 'observed' northward flux of energy in the atmosphere ($T_{IA}$) and the 'required' flux in the oceans ($T_O$). The former is determined from aeronautical observations and the latter is calculated in principle from Eqn 3 using the estimates of $F_{TA}$, $S_A$, and $S_O$. The rate of heat storage in ice and land ($S_I$ and $S_L$) have been neglected as relatively small terms but despite this and despite many uncertainties in the calculations, Fig 7 probably contains the best available estimates of the seasonal variation of the meridional energy flux in the atmosphere and oceans of the northern hemisphere.

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Fig 6  Rate of energy storage in the atmosphere ($S_A$) and in the oceans ($S_O$) (from Oort and Vonder Haar 1976).
The atmospheric energy transport shows a rather regular seasonal cycle with maximum poleward flux during the cold season, when the intensity of the atmospheric ventilation system is largest, and minimum during the warm season. Moreover, there is a small cross-equatorial energy flux from the summer hemisphere to the winter hemisphere.

The 'required' oceanic energy transport is in general polewards and its magnitude is comparable with the atmospheric flux. There is a large annual variation of $T_o$ in middle and low latitudes and at the equator there appears to be a large cross-equatorial energy flux in the oceans. The negative values in high latitudes are likely to be spurious.

Comparison of $T_a$ and $T_o$ in Fig 7 shows that in the meridional transport of energy over the northern hemisphere, the oceans dominate in low latitudes and in the subtropics whereas in middle and high latitudes the atmosphere clearly plays the dominant role.
FUTURE STUDIES

On the basis of the results presented in earlier sections it is obvious that in future climatic studies the heat transport in the oceans and the energetic interplay between the oceans and the atmosphere must have a high priority.

One question is likely to come into the mind of the southern hemisphere reader. It concerns the role of oceans in the poleward transport of energy in the southern hemisphere, which to a much larger extent than the northern hemisphere is covered by oceans. Unfortunately, the accuracy by which the 'observed' TA is known for the southern hemisphere is so low, that for the time being the method of Oort and Vonder Haar (1976) is not likely to give useful information on these areas. The First GARP Global Experiment (FGGE) in 1978-79 will hopefully provide the first opportunity to apply the method in the southern hemisphere.

A really fundamental task for the future is to obtain estimates for the 'observed' oceanic heat transports. It may be that the total 'observed' T0 will never be estimated to the same accuracy as TA. Even so, it may be possible to estimate some components of T0 with reasonable accuracy from the existing data, for example, the contribution from the time-averaged flow.

So far in this discussion, only meridional energy fluxes, averaged with respect to longitude, have been considered. The climate certainly varies with longitude and an important future task is to investigate how the energy balance is maintained in different parts of the world. In doing this both zonal and meridional components of the fluxes, as well as heat sources and sinks, are needed as a function of latitude and longitude.

Figure 8 shows the annual-mean net radiation (Fa) at the top of the atmosphere in 1969-70, as determined by Raschke et al. (1973) from radiation measurements on the Nimbus 3 satellite. As one might expect the main variation of Fa (λ, φ) is meridional, although there are significant deviations from axial symmetry. The most noticeable deviation appears over the Sahara where, in the middle of otherwise positive values, Fa < 0. To a smaller extent, the same is true for Australia and South America. In fact, if one subtracts from Fa the zonal-mean value, one finds in general a positive/negative anomaly over oceans/continents. This implies that there must, on average, be a net energy transport in the atmosphere from the oceans to the continents. An important future investigation must be to study, on the basis of aerological observations, in what form and by what mechanisms this net energy transport is accomplished. This kind of study would seem particularly interesting and feasible for the Australian area, which in the light of satellite observations is an anomalous region and has a good network of aerological stations.

CONCLUDING REMARKS

It is appropriate to close with a reference to meteorological history. The upper part of Fig 9 shows the net radiation at the top of the atmosphere as determined by Simpson (1929); these calculations were published in the Memoirs of the Royal Meteorological Society almost 50 years ago. The lower part of the figure shows the same quantity as determined by Raschke et al. (1971) from radiation measurements on the Nimbus 3 satellite during the latter part of July 1969.

Many of the general features in the two figures are similar. For example, the O-line of Fa TA runs in Simpson's calculations between 10° and 15°S, in the satellite measurements around 10°S. At 40°S Simpson's calculations indicate values close to -0.15 units (cal cm⁻² min⁻¹). The corresponding values from satellite measurements vary between -0.13 and -0.18: good correspondence! If one further takes into account the fact that for the annual-mean meridional energy flux across 35°N, Simpson obtained a value 4.1 × 10¹³ W, which is almost as good as the present-day estimates, one has really to appreciate the contribution of this pioneer in the field of the energy balance of the earth.
Fig 8 Annual-mean net radiation at the top of the atmosphere \( F_{TA} \) from measurements on Nimbus 3 satellite (from Raschke et al. 1973).

Of course, if we look at the two maps in Fig 9 in more detail, we find discrepancies. We notice, for example, that Simpson's map shows large positive values over North Africa, where satellite measurements indicate negative values. Over Australia, Simpson's map implies a positive anomaly compared with the average latitudinal conditions whereas satellite measurements show that the anomaly is in fact negative. Thus, perhaps we need not feel frustrated: we have learned something new during the last 50 years.

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REFERENCES


Fig 9(a) Net radiation at the top of the atmosphere (cal cm$^{-2}$ min$^{-1}$) in July as calculated by Simpson (1929).

Fig 9(b) Net radiation at the top of the atmosphere (cal cm$^{-2}$ min$^{-1}$) as determined by Raschke et al. (1971) from measurements on Nimbus 3 satellite, 16 to 31 July 1969.


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