The greenhouse Earth: A view from space

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SUMMARY

The natural greenhouse effect of the Earth is strongly influenced by the radiative effects of water vapour and clouds in the atmosphere, which control the energy absorbed from the sun, and that lost through thermal emission to space. Any perturbations to the climate balance, for example through so-called ‘radiative forcing’ due to increasing CO₂ amounts, variations in solar constant, or other causes, can be amplified by the feedback processes that involve water in its various phases. The radiative cooling of the Earth in the absence of clouds has recently been shown to be dominated by emission from upper-tropospheric water vapour, in the far infrared portion of the spectrum, and this is illustrated: observations of this radiative flux, and of the distribution of water vapour in the upper troposphere, are urgently needed. The role of clouds is discussed, and it is noted that their response to global warming is not presently unambiguously determined with available models, due to the complexity of competing processes: again, as in the cloud-free case, more accurate global observations are needed. The paper is illustrated by data from satellite experiments, most notably the Earth Radiation Budget Experiment sponsored by NASA.

KEYWORDS: Climate feed-back processes Clouds Earth radiation budget Global warming Greenhouse effect Radiative forcing Water vapour

1. INTRODUCTION—THE VIEW FROM SPACE

The Earth system interacts with the rest of the universe by the exchange of energy in the form of electromagnetic radiation. The incoming supply of energy comes, of course, from the sun, which emits radiation as a thermal ‘black body’ at a temperature of about 6000 K: other radiative inputs to the Earth, from the stars and from sunlight reflected by the moon, are negligible in comparison. Over time, the Earth has arrived at a balance between this energy received from the sun and energy thermally emitted to space. We shall see later that a simple calculation indicates that the effective emission temperature of the Earth, when viewed externally from space, is about 255 K. This is, of course, a lot cooler than typical temperatures at the Earth’s surface, the difference being due to the so-called greenhouse effect (hereafter GHE). We will be considering aspects of the physics underlying the GHE of the planet Earth, and how new space observations are coming to be used to obtain a better understanding of those physical processes. In particular we shall be looking at the critical role of the three phases of water, all of which are found in the atmosphere: vapour, liquid (as water clouds) and solid (as ice clouds).

The atmosphere acts as a ‘moderator’, which influences both the solar-input and the thermal-output radiation streams. The radiative processes in the atmosphere that give rise to this moderating effect are complex, and far from fully understood. In the visible part of the spectrum, as Fig. 1 shows, the atmosphere in cloud-free areas is quite transparent. However, it is important to remember that, whereas our eyes are tuned to quite a narrow
Figure 1. The spectroscopy of the atmosphere and equivalent black-body curves for the sun and the Earth. (a) The black-body curves corresponding to 6000 K and 255 K, normalized to give equal areas (energies). (b) The percentage absorption spectrum of the atmosphere in a vertical path from the surface to the top of the atmosphere: principal absorptions are labelled. (c) As (b), except from an altitude of 11 km to the top of the atmosphere. (d) The absorption spectra for individual atmospheric gases which make up curve (b).
range of the electromagnetic spectrum (between about 0.4 and 0.7 \( \mu m \)), there are huge ranges of spectral bandwidth which we cannot see. Part of this 'hidden' spectrum is the thermal infrared (IR) at wavelengths between about 3 and 100 \( \mu m \). At these wavelengths the atmosphere is much less transparent.

Figure 2 shows a well-known set of Meteosat images. The visible image (measured at wavelengths between 0.5 and 0.9 \( \mu m \)) at top left shows the high transparency of the atmosphere which we have just mentioned, and the brightness (i.e. high reflectivity) of the clouds. In the bottom left image, however, which was taken in the mid IR between 10.5 and 12.5 \( \mu m \), despite being a relatively transparent part of the spectrum (it is known as the atmospheric 'window' region—see Fig. 1), we can see a much stronger influence of the atmosphere: thus, while we can see down to the surface in regions of subsiding dry air (e.g. over the Sahara), in regions of high moisture and cloudiness atmospheric attenuation is strong. At bottom right the image looks even stranger, yet we will see that this image is
perhaps more representative of the greenhouse Earth than the familiar visible image. This is the Earth that visitors with eyes tuned to just 10 times the wavelength of our own eyes would see (in fact this Meteosat channel accepts radiation in the range 5.2 to 7.1 \( \mu \text{m} \)), and it is no longer possible to see the surface: all the radiation is coming from within the atmosphere, and is arising from the molecular vibrations and rotations of water vapour in the atmosphere. In what follows, we will see that water vapour and other gases, radiating at mid- and far-IR wavelengths where the atmosphere has strong absorption features, are responsible for a large part of the loss of energy to space by the atmosphere itself.

2. Physics of the greenhouse Earth

If we consider in a little more detail the explanation of the GHE, we may again refer to Fig. 1. Figure 1(a) shows the spectral radiance from a 6000 K black body, representing the sun, and a 255 K black body, representing the effective temperature of the Earth (see later): the two curves have been normalized to give equal areas (equal energy in and out of the Earth system). Figure 1(b) shows the absorption spectrum due to the full thickness of the Earth's atmosphere. It is plotted so that a value near 0% indicates high transmission through the atmosphere, while 100% indicates total absorption through the full thickness of the atmosphere. Note the low absorption mentioned earlier in the visible region, the strongly absorbing vibration–rotation bands in the mid IR, the atmospheric 'window' from 8–12 \( \mu \text{m} \), and the high absorption in the far IR beyond about 20 \( \mu \text{m} \). Note also that the strongest absorption bands are due to water vapour, \( \text{H}_2\text{O} \), carbon dioxide, \( \text{CO}_2 \), ozone, \( \text{O}_3 \), and to a lesser extent other gases (as shown in Fig. 1(d), which breaks down the total absorption of the full atmosphere into the effects of individual gases). Figure 1(c) shows the absorption of the atmosphere in a vertical path above an altitude of about 11 km, that is the absorption spectrum of the stratosphere and above. In this case the absorption is weaker, owing to both the reduced absorber amounts and the reduced total pressure of the atmosphere. Note that some of the more abundant atmospheric constituents, such as molecular oxygen and nitrogen, are spectroscopically relatively inactive, and so do not show up in Fig. 1.

The GHE arises because incident short-wave (SW) solar radiation is easily transmitted to the ground, while outgoing long-wave (LW) thermal radiation is partially absorbed by the atmosphere, and trapped within the atmosphere–surface system. It has to be stressed that this GHE is a natural part of the climate of the Earth, and accounts for the difference between the effective external radiative temperature of the planet and the warmer surface temperature noted earlier. Concern arises over two things, namely: how the natural GHE may vary with time, and how man's activities might modify this natural GHE. Our understanding of the physics lying behind both these questions is not yet adequate to be able to make reliable predictions of the future, largely because of the extreme complexity of the system (the climate) under study. We can see from Fig. 1(d) though, that water vapour has a very dominant influence on the GHE.

Figure 3 shows the globally averaged components of the radiation budget, expressed as a percentage of the incoming solar energy. Note the SW components of the absorption by the atmosphere (including clouds), and the reflection and backscatter of about 30% back to space. About 70% of the radiative energy loss to space comes from the LW region of the spectrum, and is due to emission from the surface, gases and clouds. We shall be particularly concerned with the LW emission from water vapour and clouds, and the SW properties of clouds. It should be noted that the numbers quoted in Fig. 3 are far from certain in some cases. For example, the absorption of SW radiation by clouds has been recently suggested (Cess et al. 1995; Ramanathan et al. 1995) to be much higher than
previously believed to be the case (nearer 10% than the previously held view of a few %), and the SW and LW properties of industrial and volcanic aerosols are now known to be very significant (Charlson et al. 1992). It is also important to recognize that through direct absorption of LW radiation by gases and clouds, and by the deposition of latent heat during cloud formation, a large amount of energy is deposited within the atmosphere from the surface.

A simple calculation serves to illustrate how we may estimate the equilibrium energy balance and effective radiative temperature, $T_{\text{eff}}$, of the Earth seen from space (see Houghton 1986). If $F_s$ is the solar flux at a distance $R_e$ from the sun, a fraction $(\pi r^2(1 - A)F_s/R_e^2)$ is intercepted by the cross-sectional area of the Earth, $\pi r^2$ ($r$ is the Earth’s radius and $A$ is the average planetary albedo (reflectivity). According to the Stefan–Boltzmann law, thermal energy at a rate of $\sigma T_{\text{eff}}^4$ is lost from each square metre of surface area $4\pi r^2$ (where $\sigma$ is the Stefan–Boltzmann constant and assuming an emissivity of 1). This allows us to derive $T_{\text{eff}}$ for the planet, as seen from space, by equating the incoming and outgoing rates-of-change of energy:

$$T_{\text{eff}}^4 = \frac{(1 - A)F_s^2}{4\sigma R_e^2}.$$  \hspace{1cm} (1)

Using values of $A = 0.3$, $F_s = 1370$ W m$^{-2}$, $\sigma = 5.67 \times 10^{-8}$ J m$^{-2}$K$^{-4}$s$^{-1}$, and $R_e = 1$ astronomical unit, gives us the value of $T_{\text{eff}} \approx 255$ K quoted earlier. The fact that the Earth’s surface temperature is greater than this, by about 25–30 K on average, is due to the natural GHE of the Earth.

3. **Forcing, Feedback and the Earth Radiation Budget (ERB)**

(a) **Forcing and feedback processes**

It is helpful, when thinking of how the balance of the GHE might be disturbed, to consider separately ‘forcing’ and ‘feedback’ (see Fig. 4). Forcings are processes which act as external agents to the climate system, such as changes in solar input to the Earth, the loading of the atmosphere with volcanic ash and aerosols, or, indeed, the introduction of rising levels of IR active gases such as CO$_2$ through anthropogenic activities.
Radiative gases (changes in concentration of $CO_2$, $CH_4$, etc)
Aerosols (natural or industrial)
Volcanic debris
Solar variations

Water vapour*
Clouds*
Snow/ice albedo
Land surface albedo
Ocean circulation & heat storage

Forcing processes $\Delta T$ Feedback processes

Figure 4. Radiative forcing and feedback processes. $\Delta T$ represents a change in surface temperature arising from the radiative forcings and the resulting feedback processes. The asterisks indicate the processes of particular concern in this study.

Figure 5. Calculations using the GENLN2 code (Edwards 1992) of the greenhouse effect for a moist tropical atmosphere. Solid curve: black-body radiance leaving the surface ($E^*$ in text). Broken curve: upwelling radiance leaving the 10 mb surface (essentially the top of the atmosphere). Dotted curve: difference between solid and broken curves, or the greenhouse-effect parameter $G$ (see text).
Figure 6. The greenhouse-effect parameter $G$ (see text) for two atmospheres: solid curve, for the moist tropical case shown in Fig. 5; broken curve, for a cold sub-arctic standard atmosphere.

Figure 7. Calculations using a 1-D radiative–convective model (due to Shine (1992)) without any feedback, of the effect of fractional changes in a number of parameters ($\text{CO}_2$, water vapour and ozone mixing ratios, and fractional cover of low-, middle- and high-level cloud). A perturbation factor of 2.0 means a doubling of that variable.
Figure 8. The outgoing long-wave radiation (OLR) calculated using the Shine (1992) model, by removing the water vapour from 50 mbar thick layers and noting the change in OLR. The units are mW m$^{-2}$ per 10 cm$^{-1}$ interval of spectrum. Abscissa gives spectral wave number (= reciprocal of wavelength), and ordinate gives pressure in the atmosphere. Case of a moist tropical standard atmosphere. (Sinha and Harries 1995).

Figure 9. As for Fig. 8, but using a cold sub-arctic standard atmosphere.

As a direct effect, these forcings may influence atmospheric temperatures and circulations. As a result of temperature changes, however, other radiatively active changes may take place: for example, the humidity of the atmosphere may change because of greater evaporation from the oceans. An increase of water vapour would cause an enhancement even further of the GHE and further warming; this is an example, therefore, of a positive feedback effect. Other feedbacks may involve the amount, height or type of cloud; a change in snow/ice cover with a consequent reduction of planetary albedo; changes in vegetation
Figure 10. Cooling rates (in m K day⁻¹ per 10 cm⁻¹ spectral interval) calculated from flux divergences for the moist tropical atmosphere case shown in Fig. 8.

Figure 11. The water vapour mixing ratio from 0.01 mb (about 80 km) to just below tropopause heights, the lowest altitudes where retrievals proved possible, for sunset 29 August to 7 October 1994 from the HALOE experiment on NASA's Upper Atmosphere Research Satellite. Units are volume mixing ratio (parts per million by volume).
and land-surface albedo; and, of course, rather slow, but potentially huge, changes in the flow and storage of energy by the oceans. Some of these feedback processes are indicated in Fig. 4.

A large number of feedback processes are known to occur in the Earth's climate system, which explains the difficulty in making a simple and quick assessment of the effects of global change. In particular, the uncertainties associated with those feedback processes involving water in its different forms cause considerable uncertainty in predictive climate models.

(b) The ERB: some definitions

The ERB is defined as the net energy flux at the top of the atmosphere (TOA) per unit area in a column of atmosphere, and is given by:

\[ R = Q^* - F^* \]

\[ = ((1 - A) F_s / 4) - F^*. \]

where \( Q^* \) is the downward flux of energy from the sun at the TOA, and \( F^* \) is the upward flux of terrestrial radiation, also at the TOA. \( F_s \) is the solar flux at 1 astronomical unit distance from the sun. The factor \( 1/4 \) arises because of the sphericity of the Earth.

At least to the accuracies of current measurement, the annual global mean of \( R \) is equal to zero (in other words, the Earth is assumed to be in net radiative equilibrium, averaged over the globe and over a year (indicated by overbars)), so that

\[ \bar{Q}^* = (1 - \bar{A}) F_s / 4 = \bar{F}^*. \]

Using values of \( F_s \approx 1370 \text{ W m}^{-2} \) and \( \bar{A} \approx 0.3 \), yields \( \bar{Q}^* = \bar{F}^* \approx 240 \text{ W m}^{-2} \).

Radiative forcing is defined as a change in \( R \) due to any perturbation. A forcing perturbs the balance between \( Q^* \) and \( F^* \); over time we expect the climate to respond to restore balance.

It is customary to define two LW GHE parameters, \( G \) and \( g \), as follows: an absolute GHE parameter, \( G \):

\[ G = E^* - F^*. \]

where \( E^* \) is the upward flux emitted by the Earth's surface. \( G \) is, therefore, simply the difference between the flux that leaves the surface and that which actually leaves the planet at the TOA. A related parameter is the normalized GHE parameter:

\[ g = G/E^* = (E^* - F^*)/E^*. \]

One cause of confusion is, sometimes, that though clouds and water vapour represent the largest GHE components, they act as feedbacks rather than forcings.

It is also common practice to define a 'sensitivity parameter', \( \lambda_i \), for a particular process, \( i \), in terms of the change in (surface) temperature, \( \Delta T_i \), caused by a change in flux, \( \Delta F_i \), due to that process (Houghton et al. 1994):

\[ \Delta T_i = \lambda_i \Delta F_i. \]

Having given these definitions, we can now describe the general detailed features of the GHE of the Earth's atmosphere. Figure 5 shows the results of a calculation carried out by a group at Imperial College (Brindley, personal communication), using a line-by-line high-resolution radiative-transfer model of the atmosphere developed by Edwards (1992).
The figure shows the LW radiation at the TOA for a warm, moist atmosphere (a tropical model atmosphere). The solid curve gives the black-body curve which describes the rate of energy leaving the Earth’s surface, $E^1$. The broken curve gives the upwelling energy rate leaving the 10 mb level (effectively the TOA). Spectral features due to H$_2$O (0–600 cm$^{-1}$, and 1500–1600 cm$^{-1}$), CO$_2$ (600–750 cm$^{-1}$), and O$_3$ (1000–1080 cm$^{-1}$) can be seen. The dotted curve is simply the difference between the first and second curve: i.e. the GHE parameter, $G$, defined earlier. While the major influence of CO$_2$ is clear, it can be seen that the effects due to H$_2$O cover a very much wider spectral range, and cannot be ignored. In fact the sheer intensity of the CO$_2$ absorption band at 15 $\mu$m actually means that the atmosphere is optically thick near the centre of this band: this means that increases of CO$_2$ in the atmosphere, while very important, do not have such a strong effect on surface temperature as the same fractional increase of water vapour.

Figure 5 also illustrates the fact that in the window between about 800 and 1200 cm$^{-1}$ there is weaker, but significant absorption. This comprises weak absorption lines due to water vapour and other gases, plus a broad-band continuum thought to arise from the effects of the far wings of strong H$_2$O lines throughout the IR (for example, see Clough et al. 1992).

4. WATER VAPOUR AND CLEAR-SKY EFFECTS

Figure 6 shows this same parameter, $G$, i.e. a measure of the strength of the GHE, for two different standard atmospheres, one warm and humid (the tropical standard atmosphere used in Fig. 5), the other cold and dry (sub-arctic standard atmosphere). In each case the upward flux at the 100 mb level, effectively the TOA, has been calculated. We can perhaps see more clearly here just how the water vapour spectrum dominates the GHE, especially in the warm, humid case. However, it must be noted that, even in the colder, much drier case of the sub-arctic standard atmosphere, the water vapour effect is still relatively strong.

The water vapour influence is seen in three separable spectral regions: (i) the pure rotation band long wave of about 20 $\mu$m (i.e. wave numbers less than about 500 cm$^{-1}$); (ii) the $v_2$ (symmetric bending) vibration–rotation band centred at 6.3 $\mu$m (about 1600 cm$^{-1}$); and (iii) the weaker, continuum absorption in the 8–12 $\mu$m (1250–800 cm$^{-1}$) range, which becomes an important mechanism for cooling the lower atmosphere in tropical regions, as we shall see later.

To illustrate the dominant effect of water vapour in another way, Fig. 7 shows further calculations (Sinha, personal communication), using a one-dimensional (1-D) radiative–convective model (Shine 1992), without any feedbacks: in other words, these are the direct radiative effects on the surface temperature of a given fractional change in any component (2.0 meaning a doubling). While we must put something of a health warning on the absolute values of each effect, because of the lack of any feedbacks, and because the model does not have complete detail of all possible processes, particularly for clouds, nevertheless, the broad results are borne out by more complex calculations. Note that, for a given change (e.g. $\times$2), water vapour has a large positive effect (which would give rise to a positive feedback on any water vapour increase due to global warming). CO$_2$ has a smaller direct effect, while cloud effects vary from a strong negative response for low, bright liquid-water clouds, to a small positive response for ice clouds such as cirrus.

One of the most interesting and significant results in recent years, however, has been the recognition (Clough et al. 1992) that in the radiative cooling of the Earth to space one very important component is the emission to space from the pure rotation band in the far IR, from water vapour in the upper troposphere. Figure 8 (Sinha and Harries 1995)
Figure 12. Measurements of the outgoing long-wave radiation (OLR) (W m$^{-2}$) to space made by the Earth Radiation Budget Experiment (Harrison et al. 1988), averaged for the month of March 1986. Top: total (clear + cloudy) OLR; bottom: clear-sky OLR.

illustrates this point well. It shows a calculation for a cloud-free tropical atmosphere, using the 1-D model mentioned earlier. The plot may be difficult to take in if not familiar with it: it shows altitude in the atmosphere on the left, and spectral wave number along the x-axis. The numbers shown are the contribution of each 50 mb-thick layer to the outgoing long-wave radiation (OLR) (in units of mW m$^{-2}$ per 10 cm$^{-1}$ interval, calculated by removing the water vapour in each layer in turn, and noting the change in OLR). Thus, we can see the contribution to the GHE at a given level and in a given part of the spectrum.

It is immediately apparent that the biggest contribution to the OLR is at pressures between 200 and 600 mb, in the far IR between about 20 and 40 μm (500 and 250 cm$^{-1}$). Note the influence of CO$_2$ at about 650 cm$^{-1}$, and O$_3$ at about 1000 cm$^{-1}$, which, because
Figure 13. The clear-sky outgoing long-wave radiation (W m\(^{-2}\)) at the top of the atmosphere for March 1982, by the Hadley Centre CLERA project (see text).

Figure 14. As Fig. 13, except clear-sky long-wave divergence (W m\(^{-2}\)).
they are not removed in each layer, show up as gaps in the figure. We can again see the influence of the three spectral bands mentioned earlier (pure rotation, vibration–rotation, continuum). Before Clough et al.'s seminal paper in 1992, the true significance of this very-long-wave emission to space had not been fully recognized.

Figure 9 (also from Sinha and Harries 1995) shows a similar calculation, but for a cold, dry atmosphere (note that the effect of a near-isothermal temperature inversion near and at the surface is to make the OLR insensitive to removal of water vapour layer by layer below about 800 mb). It is clear that, even under these cold/dry conditions, the H_2O pure rotation band still is a dominant effect in the radiative cooling of the atmosphere.

Similar calculations have been used to prepare Fig. 10, in this case to derive the vertical flux divergences, and so the amount of energy lost by radiation in each atmospheric layer (the cooling rate in m K day^{-1}—Goody 1964). This figure dramatically demonstrates that there are essentially two domains: firstly, lower tropospheric humidity (LTH) controls the cooling to space from the lower layers of the troposphere, at the relatively transparent wavelengths of the atmospheric window: and, secondly, upper tropospheric humidity (UTH) controls the cooling to space from the upper layers of the troposphere, via the optically strong pure rotation band below about 600 cm^{-1}, and to a lesser extent in the ν_2 band.

A major concern must be that observations in the atmosphere throughout the far IR of the spectral fluxes at different levels or at the TOA, even on a local basis, are not available, and are certainly not available globally from space. It is critical that we obtain data to test the performance of our theories. A group at Imperial College is planning an airborne experiment to make in-situ measurements of upward and downward radiances in collaboration with the Meteorological Office and its Hadley Centre, using the Meteorological Research Flight's C-130 aircraft.

Knowledge of the distribution of UTH is also very limited. There have been a few aircraft measurements (notably by the NOAA* Aeronomy Lab as we heard from Dr Adrian Tuck in the Symons Memorial Lecture in May 1995). These have been shown to be of very high accuracy, but are very localized. Global measurements in the lower stratosphere have been made by limb sounders, for example, the Halogen Occultation Experiment (HALOE) on the NASA† Upper Atmosphere Research Satellite (UARS) (Russell et al. 1993). Figure 11 shows an example of HALOE measurements of water vapour for the period 29 August to 7 October 1994, illustrating that considerable structure is seen from the low stratosphere to the mesosphere, but that little information is generally available in the troposphere, because (owing to the choice of spectral bands) spectral lines are saturated. Also, while these data are available on a quasi-global basis, and at high accuracy (e.g. see Harries et al. 1996), they possess limited vertical and horizontal resolution, an especially serious limitation when we are concerned with the tropopause region, where vertical (and sometimes horizontal) gradients of mixing ratio can be very large. Data are also available from operational meteorological nadir sounders, but the vertical resolution achievable with vertical-viewing nadir sounders is poor. It has to be said that the exploitation of satellite data for the troposphere is in its infancy, and is bound to receive much more attention in coming years. It should also be stressed that much more effort is required to make measurements of the spectrum of radiative fluxes in the atmosphere, and of the humidity structure of the upper troposphere.

What of observations of the clear-sky ERB from space? The primary source during the 1980s has been the Earth Radiation Budget Experiment (ERBE) and Nimbus 7 satellites

* National Oceanic and Atmospheric Administration.
† National Aeronautics and Space Administration.
(Harrison et al. 1988). An example of the satellite data is shown in Fig. 12. The top frame shows the ERBE-measured LW emission flux to space averaged for the month of March 1986, for all cases, cloudy and cloud-free. The bottom frame shows the equivalent figure for cases of cloud-free skies only. The global/annual average OLR is, as we have seen, near 240 W m\(^{-2}\), but considerable regional and time variations occur in the observations. In the cloud-free case there is a broad zonal appearance, which is driven by the atmospheric temperature distribution, with small regions of higher levels of OLR in sub-tropical areas where descending, dry air is allowing more radiance from lower, warmer layers to escape to space. In the northern hemisphere (NH) the warming effect of the oceans versus the land is apparent. In the cloudy data, the much reduced cooling to space above high tropical cloud systems is apparent, which emphasizes the sub-tropical descent zones mentioned earlier.

Several studies have been carried out using the ERBE data. For example, a number of intercomparisons of general-circulation models (GCMs) with ERBE data exist. These include several studies by Stephens and colleagues on the relation of hydrology to atmospheric radiation, which have shown how powerful the use of global satellite data can be (Stephens and Greenwald 1991; Stephens et al. 1993). One of the most illuminating efforts in recent years has been the work at the Hadley Centre by the group led by Slingo. A powerful method, called SAMSUON (Simulation and Analysis of Measurements from Satellites using Operational nAlysis), has been developed, in which atmospheric data from operational analyses are used as input to a radiative-transfer code to simulate OLR from the atmosphere (Slingo and Webb 1992). These, and other, studies indicate agreement between models and the ERBE on a global basis of perhaps \(\pm 5-10\) W m\(^{-2}\), which is probably near the accuracy of the experimental observations (though there is no definitive published report on those accuracies).

Recently the same team developed the CLERA (Clear-sky Longwave from ECMWF Re-Analyses) project. The European Centre for Medium-Range Weather Forecasts (ECMWF) is engaged in a major exercise to re-analyse available atmospheric data from 1979 to 1993, including the assimilation of satellite (TOVS)* humidity and temperatures. Re-analyses have 1° horizontal resolution, and 19 levels in the vertical. These new data are being validated at the Hadley Centre by simulating the 3-D structure of clear-sky LW fluxes and cooling rates, for comparison with ERBE data, in a study of the mechanisms controlling clear-sky GHE, water-vapour feedbacks, and in order to validate climate models.

Figure 13 shows clear-sky OLR for March 1982, calculated using the CLERA process. Generally high values of OLR are observed in the tropics, with a decrease towards the poles due mainly to the effect of decreasing temperature. However, the effects of moisture dominate in the tropics as evidenced by minima over the convective regions (caused by high values of UTH, and thereby emission from high, cold layers of the atmosphere), and maxima in sub-tropics (UTH minima allow radiation from lower, warmer layers to escape to space). The great value of the CLERA approach is that not only TOA fluxes may be calculated, but other, internal, atmospheric parameters may be examined. In Fig. 14, for example, the clear-sky LW atmospheric divergence is shown, which is the difference between the previous data and the calculated net surface LW radiation. The divergence is, of course, related to the net heating rate, or the available atmospheric energy (Houghton 1986). Generally speaking, the divergence is highest in the tropics and lowest over the poles: studies show that the divergence correlates well with total column moisture, except in the tropics where the effect of UTH is to make the vertical profile of water vapour more relevant to the OLR.

* TIROS (Television Infra-Red Observation Satellite) Operational Vertical Sounder.
Figure 15. Cloud long-wave forcing, estimated from Earth Radiation Budget Experiment measurements, for January 1986.

Figure 16. As Fig. 15 but for cloud short-wave forcing.

Figure 17. As Fig. 15 but for cloud net forcing.
5. CLOUDS

Clouds have an important effect on both the SW reflection properties of the Earth, and the LW GHE of the atmosphere. Despite this importance, however, we are still unsure about some of the major questions, though space observations are having a powerful effect on our understanding of the role of clouds. Conventionally, the radiative effect of clouds is discussed in terms of the 'cloud radiative forcing', \( C \), even though cloud effects are feedback processes in principle. \( C \) is the difference between the TOA energy budget (net flux \( R = Q \downarrow - F \uparrow \), see Eq. (2)) for a given column for clear and cloudy cases:

\[
C = R - R_{\text{clear}} = (Q \downarrow - F \uparrow) - (Q_{\text{clear}} \downarrow - F_{\text{clear}} \uparrow)
\]

\[
= (Q \downarrow - Q_{\text{clear}} \downarrow) + (F_{\text{clear}} \uparrow - F \uparrow)
\]

\[
= C_{\text{SW}}(-\text{ve}) + C_{\text{LW}}(+\text{ve}),
\]

where the cloud forcing has been divided into a SW and LW component indicated by subscripts.

ERBE data once more have been widely used for cloud-radiation studies. Figure 15 (Harrison et al. 1993) shows \( C_{\text{LW}} \) for one particular month, January 1986. We see very high values of \( C_{\text{LW}} \) over the 'warm pool' region of the western Pacific, where strong convection occurs; near-zero values over the sub-tropics (dry, descent, cloud-free); and high values over NH storm-tracks. Figure 16 shows the corresponding \( C_{\text{SW}} \), with large (−ve) values in the summer hemisphere (southern); low values in cloud-free zones (e.g. over the Sahara); and low values in the far north where no sunlight occurs in January. Figure 17 shows the net cloud forcing, \( C = C_{\text{LW}} + C_{\text{SW}} \), formed by the combination of Figs. 15 and 16. This parameter ranges from \(-140 \text{ W m}^{-2}\) to \(+50 \text{ W m}^{-2}\); the global average is about \(-20 \text{ W m}^{-2}\). Recent GCM model predictions of \( C \) give values between about 0 and about \(-30 \text{ W m}^{-2}\). We might note in passing that in Fig. 17 the high values (up to about \(40 \text{ W m}^{-2}\)) observed in the North Atlantic are due to high values of \( C_{\text{LW}} \) (due to storm tracks) balanced against almost zero \( C_{\text{SW}} \) (due to little sunshine). Note also that, despite huge variations in both \( C_{\text{LW}} \) and \( C_{\text{SW}} \) terms (c. \(100 \text{ W m}^{-2}\), there is near perfect balance around the equator.

One important feature to recognize (and over which there has been some confusion in some papers) is that though the absolute value of \( C \) may be negative, it is not true that any change, \( \Delta C \), in \( C \) (say in response to global warming or any other forcing) would necessarily also be negative. Because many complex changes of cloud properties, e.g. height, or temperature, are possible in response to global changes, it is not currently possible to know whether \( \Delta C \) would be positive or negative.

Another important development in satellite data on clouds has been the International Satellite Cloud Climatology Project (ISCCP–Schieffer and Rossov 1983). ISCCP has produced a merged geostationary/polar orbiter data set, from imagers and sounders, in the visible and IR spectral regions, which includes IR radiances, cloud amount, optical thickness, and top pressure. These are now being used by researchers in studies of cloud-radiation processes.

Just as for the clear-sky case, there are a number of unresolved issues in the problem of clouds and radiation, some of which are mentioned here:

(i) The net radiative effect of different cloud types depends on their properties and altitude; cloud response to global warming is very complex, involving possible changes in cloud amount, altitude (temperature of emitting surfaces), water and ice content, and microphysical properties.
(ii) Many models currently predict a decrease of cloud amount, and a rise in average cloud-top height as a response to global warming. This is far from a secure conclusion however, due to the many uncertainties.

(iii) The effect of changes in atmospheric aerosols is not certain, due to both the direct radiative effects and the indirect effects arising from the formation of clouds on aerosol particles.

(iv) Previous mention has been made of the controversial issue, currently unresolved, concerning the SW absorption properties of clouds: two recent publications (Cess et al. 1995; Ramanathan et al. 1995) indicate that there might be considerably larger absorption at short wavelengths than previously expected (order 10% rather than a few %), and the case was argued that this was necessary for ‘closure’ of the radiation budget as evidenced by a number of measurements. This remains a controversial suggestion (e.g. Li et al. 1995).

(v) The situation concerning clouds and their radiative effects on the climate system has been summarized by Hartmann (1993) as follows: ‘The net effects of clouds under global warming are not currently predictable’. Improvements are needed not only in our ability to model cloud processes, but also we need considerable improvement in our ability to measure the 4-D distribution of radiation, cloud and humidity.

6. Summary

The GHE of the Earth may be ‘forced’ by changes in composition, or in solar radiation, but is very strongly controlled by feedback processes, especially those involving water vapour and clouds. For comparison, these feedback processes can give rise to surface radiative fluxes of many tens of W m$^{-2}$, compared with about 4 W m$^{-2}$ due to global warming through CO$_2$ doubling. The effects are, therefore, extremely important, and must be very accurately modelled in large-scale climate-prediction models if considerable errors are to be avoided. Observations from space have been used successfully to study a number of important radiative driving and feedback effects, but it is suggested that the exploitation of global satellites is in its infancy.

Water vapour cooling to space occurs principally from: the lower troposphere through the atmospheric ‘window’ between 8 and 12 $\mu$m wavelength, upper-tropospheric water vapour via the pure rotation band beyond 20 $\mu$m, and the 6 $\mu$m vibration–rotation band. Despite these effects being major ones in the energy loss of the planet to space, there are very few experimental observations by which one can test theoretical ideas on water-vapour distributions, radiative fluxes, and cooling rates. Therefore, observations of both the water-vapour distribution in the upper troposphere, and the radiative fluxes in the far IR are badly needed.

Current studies of the clear-sky GHE using assimilation schemes with radiation codes are able to bring global satellite observations and calculations into agreement within about 5–10 W m$^{-2}$. Regionally, larger differences exist, rising in places to several tens of W m$^{-2}$. Improved observations from space are urgently required.

Global observations of the influence of clouds on the radiation balance have provided considerable new information: for example, it is clear that the LW greenhouse forcing varies from about 0 to 100 W m$^{-2}$, (e.g. over the ‘warm pool’ region); the SW albedo forcing varies from about $-140$ to $+50$ W m$^{-2}$, with highest reflection of solar radiation in summer storm tracks, and over convective regions. Global mean cloud forcing, $C$, is $\approx -20$ W m$^{-2}$. The negative value of $C$ does not mean that clouds necessarily produce a
negative feedback to global warming: the sign of a change in \( C \) following any forcing will depend on many competing processes, and currently cannot be predicted with certainty.

Observations from space are beginning to make a quantitative impact on GHE studies: but improved observations are needed with higher absolute accuracy, better sampling, and the capability of deriving 'internal' quantities (water vapour and cloud distributions, cloud optical properties, radiative fluxes). As a final remark, we might note that three promising new observational developments are:

(i) To fly a radiation-budget instrument on a geostationary satellite to obtain crucial information about diurnal variations of key radiative processes. A project entitled the Geostationary Earth Radiation Budget experiment is proposed for flight on the Meteosat second-generation satellites, beginning in the year 2000.

(ii) A cloud profiling radar is under study as a means of obtaining the vertical distribution of cloud properties on a near-global basis.

(iii) The European Space Agency is studying an Earth Radiation Mission: this is an attempt to build a complete mission (satellites, other observations, theory, models) to address the Earth's GHE—one of the most challenging and important scientific problems facing mankind today.

The GHE of the Earth is how it is in large part because of water on the planet. Water in the atmosphere, as vapour, water cloud and ice cloud (as well as water in the oceans and cryosphere), plays a major role in controlling our climate, providing many positive feedback paths, and perhaps fewer negative ones. Do we understand climate stability? Aside from somewhat generalized views of the intrinsic stability of such a complex system, it could be argued that we do not. Improved space observations will, I believe, continue to have a major role to play in solving this fascinating global problem. I hope that through this brief review of aspects of the GHE of the Earth, and of the way we can use space data to study it, I have been able to illustrate the importance of radiation studies in our subject of meteorology and atmospheric science.

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