1

The balance of energy

Solar radiation

The Sun emits around 74 million watts (W) of electromagnetic energy from each square metre (m^2) of its surface. The energy arriving at the outer boundary of the Earth's atmosphere, at right angles to the solar beam, is about 1,366 W m^{-2} (watts per square metre). This value is known as the *solar constant*, although it is not quite constant, for reasons discussed later.

Figure 1.1 shows the spectrum of electromagnetic radiation emitted by the Sun, and also those parts absorbed during its passage through the Earth's atmosphere and the radiation received at the Earth's surface. Because the Sun is rarely directly overhead, and is out of sight at night, the average amount of energy received at any one point on the surface is much less than the solar constant; the global average at the outer boundary of the Earth's atmosphere being about 342 W m^{-2} (or about one-quarter of the solar constant).

Passage of radiation through the atmosphere

During its passage to the ground (Figure 1.2), around 67 W m^{-2} (or 19.6% of the 342 W m^{-2} input) is absorbed by the atmosphere, warming it slightly in the process. A further 77 W m^{-2} (22.5%) is reflected and scattered back to space by the atmosphere, clouds, aerosols, smoke from many sources, microscopic living organisms, dust from sandstorms, volcanic ash, micrometeorites from space, and salt crystals from the sea; the amount finally reaching the surface being 198 W m^{-2} (58% of that received just outside the outer boundary of the atmosphere). A further 30 W m^{-2} (9%) is immediately reflected back to space from the ground, the proportion reflected being known as the *albedo* of the surface, its magnitude depending on the nature of the surface. This means that the amount of energy lost back to space is about 107 W m^{-2} (31.3% of the input) which, together with that absorbed

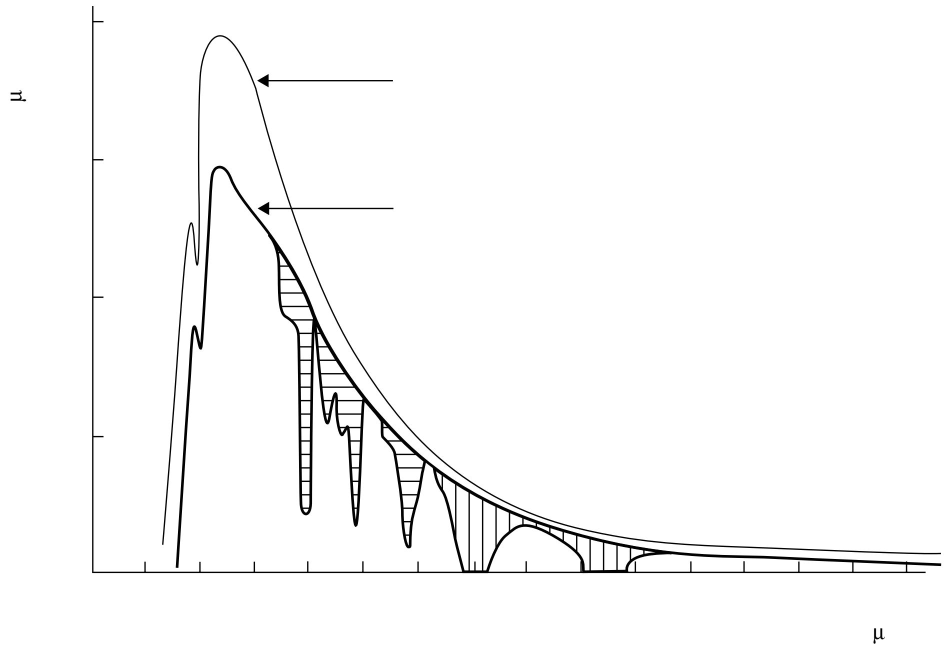

Figure 1.1 Spectrum of solar radiation at the outer limits of the atmosphere and on the ground (at sea level). *Vertical shading* indicates absorption by water vapour and carbon dioxide. *Horizontal shading* indicates absorption by water vapour and oxygen. Oxygen and ozone absorb in the UV below 0.4 μm (not shaded).

during its passage through the atmosphere, leaves 168 W m^{-2} 'available' at the surface (or 49.1% of the energy arriving at the outer boundary of the atmosphere). This is an average, as the levels available vary greatly with the time of day and season as well as with the nature of the surface. What happens to this 168 W m^{-2} depends on surface conditions – on whether it is land or sea, city or countryside, on the nature of the soil and of the vegetation, and on how much water is available.

Sensible heat flux

Initially the Sun's energy heats the Earth's surface, be it land or sea. Some of the warmth is then transferred to the air, first by molecular diffusion across what is known as the *laminar boundary layer* (a layer only a millimetre or so thick, which clings by molecular attraction to most surfaces). Beyond this, in the *turbulent boundary layer*, transfer of the warmed air away from the surface occurs through the action of the wind and eddies within it – a process that is much more effective than the initial very slow molecular diffusion. Through the action of wind turbulence alone, the atmosphere would be mixed rapidly to the top of the troposphere, taking the heat with it.

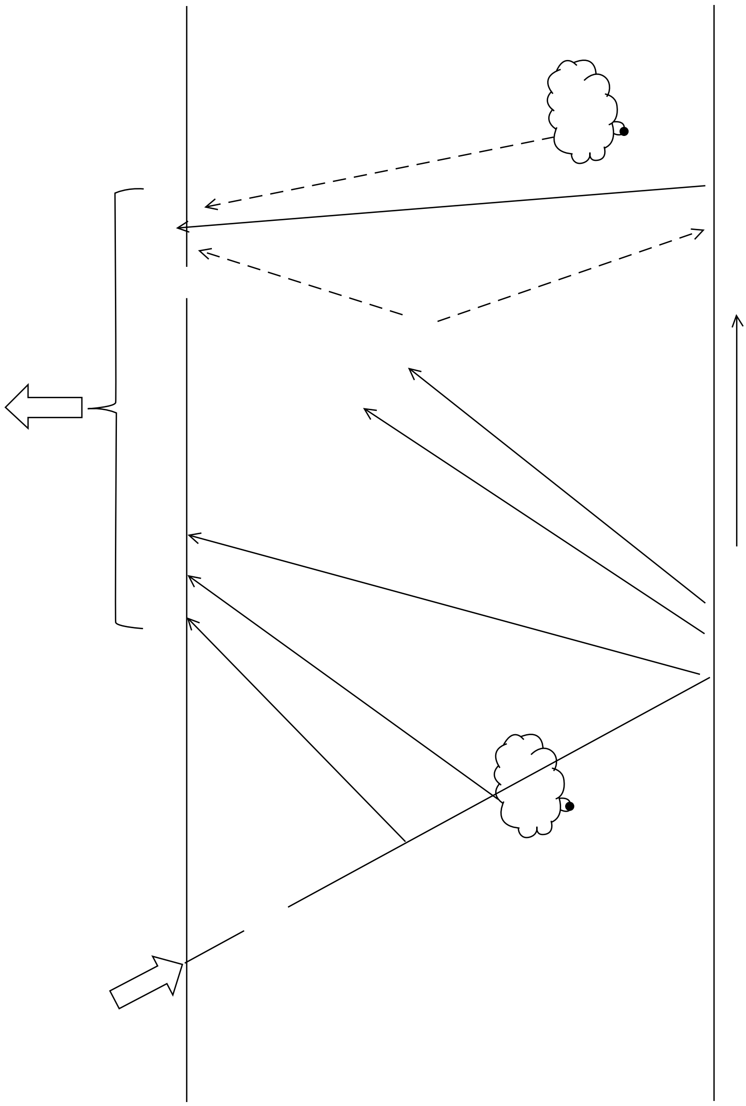

Figure 1.2 Energy exchange of solar and infrared (IR) radiation within the atmosphere. See text for discussion.

The atmosphere is made up of several layers (Figure 1.3): with the *troposphere* being the lowest layer, in which we live and in which most clouds exist, extending from the surface to a height of between 8 and 20 km, depending on latitude; its upper limit is known as the *tropopause*. The transfer of warmed air away from the surface is known as the *sensible heat flux* ('sensible' in that it can be sensed, or measured, as a change in temperature; 'flux' referring to the rate of passage of the heat energy upwards). The sensible heat flux consumes, on average, 24 W m^{-2} (14.3%) of the available 168 W m^{-2} solar input, but this varies greatly with time, place and conditions.

Soil heat flux

Over the land, a small proportion of the heat from the warmed surface is also transferred downwards by conduction as the *soil heat flux*, this generally being returned to the surface during the night or over the course of a season. The rate of transfer of the warmth downwards is influenced by the amount of water in the soil, by the soil's pore and particle sizes, and by the presence of vegetation. Many biological processes are influenced by soil temperature, from the activity of micro-organisms to the germination of seeds and plant growth. Over the oceans somewhat different processes occur (see the section *Radiation processes over the oceans* later in this chapter).

Latent heat flux

In addition to warming the atmosphere, a proportion of the available solar energy arriving at the Earth's surface also evaporates some, or all, of the available water, producing water vapour. The resultant vapour mixes with the warmed air and is transported away by the same processes. This is known as the *latent heat flux* ('latent' because the heat is hidden – there being no change of temperature associated with the process, just a change of state from liquid to gas). For reasons explained later, this hidden energy is important. Evaporation takes place from the sea, from open lakes and rivers, from moist soil, from wet vegetation and as transpiration from plants through their stomata; the whole process often being called *evapotranspiration*. An average of 78 W m^{-2} is used up in this way, although the amount depends very much on circumstances, in particular on the amount of water actually available, on the temperature of the air and water, on how humid the air already is, on wind speed, on the texture and colour of the plants and soil, and on the many different combinations of these factors. Over the sea, for example, the unlimited supply of water produces a somewhat different relationship between sensible and latent heat flux than that found above a desert.

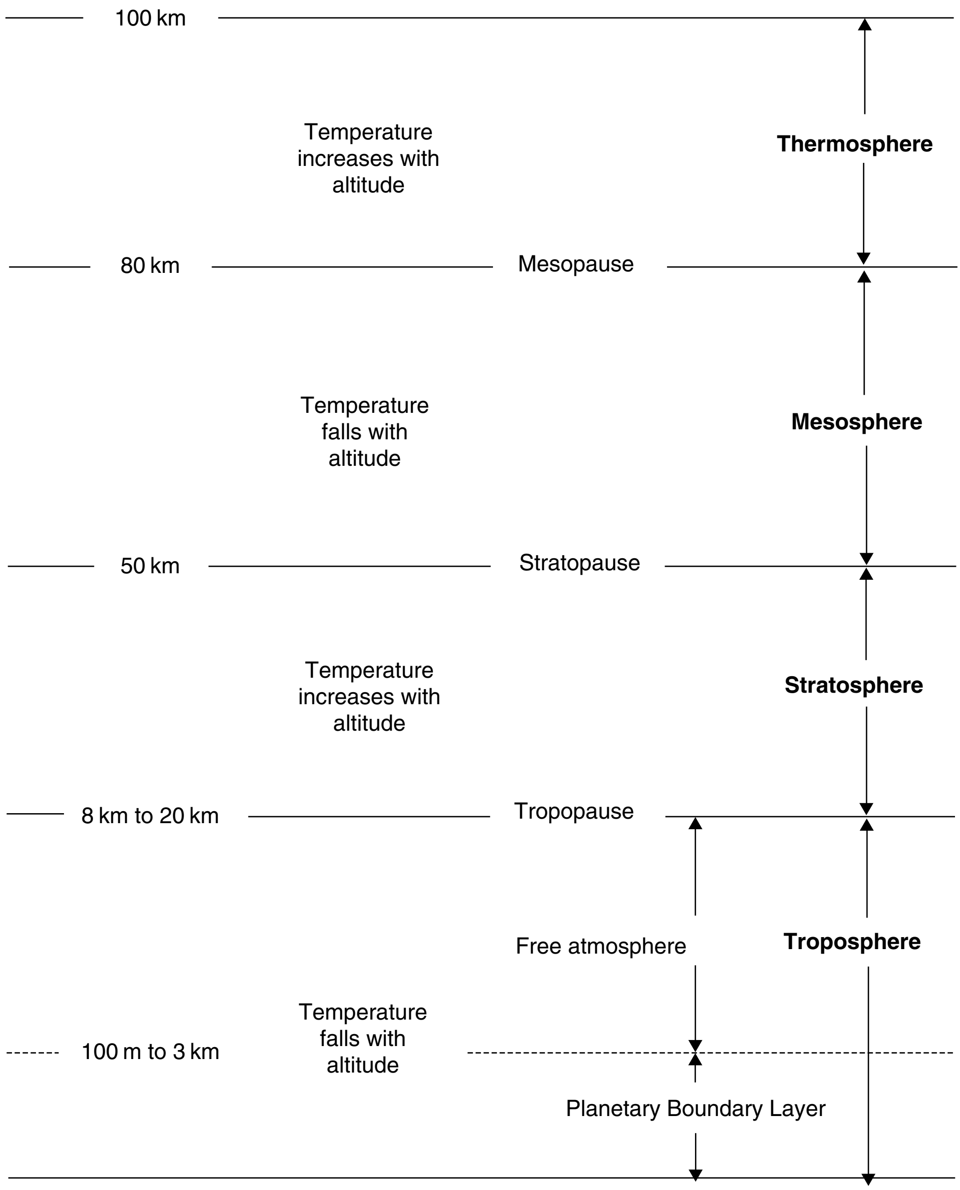

Figure 1.3 Structure of the atmosphere.

Convection and clouds

As well as being transferred away from the ground by wind turbulence, the warmed air is also transferred upwards by the process of *convection*: warmer, and therefore less dense air rising into the relatively cooler, denser air above, taking the water vapour with it. To balance out this upward movement, cooler air descends elsewhere, thereby thoroughly mixing the atmosphere. This mixing occurs right up to the tropopause (Figure 1.3). (There is also a process termed *advection*, which is simply the movement of a fluid in any direction, such as sideways as in the blowing wind; it means 'to transport away'.)

The next piece of reasoning is important. Because air pressure reduces with height, rising packets of air (often known as *thermals*) expand as they ascend and, in so doing, they cool by virtue of the first law of thermodynamics – the conservation of energy. (Those interested in a mathematical explanation of why this happens can refer to Boyle's law and Charles's law in Appendix A.) The process cannot be understood intuitively, but it is not necessary to be able to follow the (simple) maths in order to accept the fact that the expansion of a gas cools it, and vice versa (for instance, a bicycle pump heats up when inflating a tyre, due to the air inside being compressed). Until the maths was done, the whole process was something of a mystery to scientists (Strangeways 2007). Provided that the rising thermals do not mix (much) with the air that they are rising through, the rate of cooling is known as the *dry adiabatic lapse rate* and is between 6° and 10°C for each kilometre increase in height. When mixing of the air occurs, it is termed *diabatic* or, more usually, the negative term *non-adiabatic*. 'Lapse' refers to the fact that the temperature decreases with height.

Because of this cooling, a height will eventually be reached where the temperature of the rising thermal becomes the same as the surrounding air, whereupon ascent stops. However, the water vapour which has risen with the warmed air may condense back into water (as cloud droplets) if the air cools to the *dew point* (see Appendix B). The important point here is that when the water vapour condenses back to the liquid form, all of its 'hidden' (latent) energy, absorbed in the evaporation process at ground level, is now released, thereby warming the air. Thus heat is transported upwards in the guise of water vapour as well as in the form of warmed air, provided, that is, that condensation occurs. Moreover, once condensation starts to occur, the resultant warming stops the air from cooling as fast as it would have done if there had been no condensation, and this has the effect of causing the thermal to continue rising further than it would have done if the dew point had not been reached. The lapse rate of moist air, known as the *wet adiabatic lapse rate*, is thus less than that of dry air (about 4.5–5.0°C per kilometre rise) because the heat released by condensation partly reduces the rate of cooling. Under certain atmospheric conditions, termed 'unstable', the thermal can continue to rise right to the top of the troposphere, where it meets the tropopause.

The tropopause

The tropopause is the boundary between the top of the troposphere and the bottom of the stratosphere (Figure 1.3), at which point rising thermals stop ascending because they meet a 'ceiling' or inversion. The ceiling results from the fact that, in the stratosphere, temperature increases with height – in contrast with it decreasing with height in the troposphere – thereby reducing or preventing convection. At the tropopause, therefore, the air and any clouds within it spread out sideways, travelling downwind, producing the familiar anvil shape of a thundercloud (Strangeways 2007).

The rise of temperature with altitude within the stratosphere is due to heating from above by the absorption of ultraviolet (UV) wavelengths by the ozone layer – this action giving off heat. The ozone itself is formed in the first place by the UV radiation splitting oxygen molecules (O_2) into single atoms of oxygen, these then combining with O_2 molecules to form O_3, or ozone. The UV then subsequently splits the O_3 back into O_2 and O, with the process cycling continuously. Most of the ozone is concentrated at an altitude of between 15 and 35 km, where its concentration is between 2 and 8 ppm (parts per million), although the ozone extends from 10 km to 50 km. The amounts are very small; if all the ozone was collected together at surface pressure it would be only a few millimetres thick.

The tropopause occurs where the heating by conduction from above, in the ozone layer, and the heating by convection from below balance out. Because the surface is colder in polar regions, the tropopause is lower there. But even within the troposphere, convection may be halted by inversions before the tropopause is reached, because under some conditions, such as an anticyclone, warm air can occur above colder air below, producing what is termed a 'stable' atmosphere.

As noted earlier, the latent heat transfers just under 78 W m^{-2} of the available 168 W m^{-2} arriving at the Earth's surface, leaving, on average, 66 W m^{-2} still unaccounted for. What happens to this remaining energy?

Infrared radiation and the greenhouse effect

Until Sir Benjamin Thompson, Count Rumford, wrote his classic paper (Thompson 1804) on infrared (IR) radiation, no-one was aware of its existence, nor of the important role that it played in meteorological processes. In particular, no-one could explain why dew formed on clear nights. What Count Rumford found, by experiment, was that all objects above a temperature of absolute zero (see the section *The absolute, or thermodynamic, temperature scale* in Chapter 2) emit IR radiation and, in so doing, are cooled. Infrared radiation is that part of the electromagnetic spectrum (see Appendix C) which has a wavelength longer than that of visible light but shorter than that of microwaves. Everything above a temperature of absolute zero emits IR. (When an object is very hot, some of the radiation emitted is in the visible spectrum, the hotter the bluer, the cooler the redder.) At lower temperatures, the radiation is restricted to the IR and microwave spectra; the actual range of wavelengths and their intensity depends on the temperature of the object. The IR spectrum stretches from 0.7 μm to 1 mm, divided into near-, mid- and far-IR (Appendix C).

The remaining 66 W m^{-2} of the available energy at the surface is radiated from the warmed ground in the form of this IR radiation, with the amplitude and wavelength depending on the temperature of the ground. Most gases are not affected by the passage of IR radiation through them and, if the atmosphere was composed entirely of oxygen and nitrogen, the IR radiation would pass through as if they were not there and would be lost to space; there would be no greenhouse effect. But some gases do react with IR by absorbing the radiation. In Earth's atmosphere the main greenhouse gases (GHGs) are water vapour (WV), carbon dioxide (CO_2) and methane, with WV being by far the most powerful and prevalent GHG, accounting for about 95% of the total greenhouse effect, carbon dioxide accounting for 3.6% and methane and all the others put together accounting for 1.4%.

The GHGs in the atmosphere absorb some of the IR emitted from the surface, and warm up. But it is in the nature of molecules that absorb radiation at a given frequency also to radiate it at the same frequency, so the warmed GHGs re-radiate some of the received heat in all directions, part of it downwards towards the surface, causing the atmosphere below, and the surface, to warm up. Some of the upwardly re-radiated IR is lost to space and some is absorbed by the atmosphere higher up; it is a very circular process at all altitudes. The amount of IR radiation received back at the surface, as *back radiation*, from all of the GHGs mentioned above, is about 324 W m^{-2}, which, when added to the 66 W m^{-2} remaining from the incoming energy budget (see above), amounts to 390 W m^{-2}. Continuing the circular process, this radiation is now radiated back upwards.

Because the planet must be in thermal equilibrium, the energy lost to space must balance that coming in from the Sun. Since the input from the Sun is 342 W m^{-2}, of which 107 W m^{-2} is lost as reflected solar radiation, the infrared losses must make up the difference; that is 342 – 107 = 235 W m^{-2}. This is made up of three parts: a direct loss of 40 W m^{-2} through a clear atmospheric 'window' (a band of wavelengths transparent in that part of the IR spectrum), a loss of IR from cloud tops of 30 W m^{-2}, and a loss of 165 W m^{-2} from the atmosphere, totalling the required balance of 235 W m^{-2}. This means that the energy retained by the greenhouse effect is 390 – 235 = 155 W m^{-2}, keeping the surface and lower troposphere warmer than it would otherwise be by about 21°C. Without the GHGs, the average temperature of the planet would be −6°C. Thus, with the GHGs, the average is 15°C.

Since atmospheric temperature, overall, decreases with altitude, the radiation emitted becomes less intense with increasing height. Thermal balance is, therefore, achieved by the temperature of the Earth's surface and its lower atmosphere increasing sufficiently for the cold upper atmosphere on the edge of space to be warm enough to radiate exactly the required 235 W m^{-2}. So it is the drop of temperature with altitude that makes the greenhouse process work. It would not happen if the temperature was the same all the way up. Nor would it happen if the atmosphere was composed only of oxygen and nitrogen, since they do not absorb IR radiation. Indeed, if this were so, life would probably not have developed on Earth, for the surface would be at an average temperature of −6°C. We should, therefore, be thankful for the presence of water vapour and carbon dioxide; our existence depends on them. (While the solar input only occurs during the day, the IR radiation is continuous, day and night.)

The greenhouse effect has been going on ever since the Earth's atmosphere formed 4.5 billion years ago, when it contained water vapour, hydrogen, hydrogen chloride, carbon dioxide, carbon monoxide and nitrogen. Through its interaction with rocks and early living organisms, the primitive atmosphere slowly evolved, with photosynthesising plants eventually changing some of the carbon dioxide into oxygen. The atmosphere reached its present composition only about 280 million years ago, since when it has changed but little. Its fixed part is now made up of 78.09% nitrogen, 20.95% oxygen, 0.93% argon and 0.03% carbon dioxide (by volume), along with a few trace gases.

The water vapour component of the atmosphere, however, varies considerably across time and space. The amount of water vapour that air can hold (the *relative humidity*: see Appendix B) reaches a theoretical maximum of about 3%, although normally it is less, with 2% representing very moist air. The presence of water vapour in the atmosphere causes complications when considering the greenhouse effect, because it is a considerably more powerful GHG than carbon dioxide and a great deal more plentiful, consequently producing a much greater greenhouse effect than carbon dioxide (about 95% of the total).

The *enhanced greenhouse effect* is the increase in retained heat above the natural mean of 155 W m^{-2} (see above) as carbon dioxide levels increase, amounting to about 4 W m^{-2} for a doubling of the pre-industrial concentration. Although there is no doubt whatsoever about the physics and maths of the greenhouse effect, CO_2 does not act on its own and there are many other factors involved in the workings of the climate. Consequently a very simple, one-to-one, direct link between the amount of CO_2 present in the atmosphere and the temperature at the Earth's surface is most unlikely to exist. It is, however, most probable (all else being equal) that an increase in temperature will increase evaporation and thus the amount of water vapour in the atmosphere. Being a GHG, the water vapour might then increase the temperature further. In climate models water vapour is treated simply as a positive feedback in isolation, resulting from temperature increases due to CO_2 increases. However, the increased water vapour might increase low cloud cover and this would reflect more solar radiation resulting in negative feedback and cooling. Some models include this effect, others do not (see *Clouds,* page 149). The intertwined processes producing Earth's climate are complex.

Measuring the radiation

Sunshine duration

It was not possible to measure the output of the Sun precisely until there were satellites above the atmosphere equipped with suitable sensors. Prior to this, measuring the intensity of solar radiation at the Earth's surface was all we could do. Until the last 50 or so years, the most common way of measuring solar activity was by the Campbell–Stokes sunshine recorder developed in the nineteenth century (Strangeways 1998, 2003) (Figure 1.4). Although elegant and well designed, it gives only an indication of whether the Sun is shining brightly or is obscured by clouds. It tells us only indirectly about the radiation intensity: only if the time of day, date and latitude are known can a fair estimate of intensity be calculated. It is also difficult to interpret the burnt tracks on the recorder charts, especially if clouds obscure the Sun frequently for short periods of time. The recorder does, however, give a useful record of cloudiness, but only during the daytime. Modern electrical equivalents of this old device have been developed to enable sunshine duration to continue to be measured by unattended automatic weather stations (AWSs).

Solar radiation

To measure the intensity of solar radiation, instruments known as *pyrheliometers* automatically track the passage of the Sun across the sky, in the same way as a telescope on an equatorial mount tracks stars. The second-by-second variations in intensity of the radiation are measured by sensing the temperature of a black surface exposed to the Sun at the bottom of a tube, relative to a similar black surface screened from the Sun (thereby compensating for air temperature changes). The tube has a field of view that encompasses just the Sun's disc and a small annulus of sky around it. The variations in intensity caused by changes in the atmosphere would swamp any changes occurring in the amount of energy