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Phil. Trans. R. Soc. A 2012 **370**, doi: 10.1098/rsta.2012.0004, published 6 August 2012

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The runaway greenhouse: implications for future climate change, geoengineering and planetary atmospheres

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The ultimate climate emergency is a ‘runaway greenhouse’: a hot and water-vapour-rich atmosphere limits the emission of thermal radiation to space, causing runaway warming. Warming ceases only after the surface reaches approximately 1400 K and emits radiation in the near-infrared, where water is not a good greenhouse gas. This would evaporate the entire ocean and exterminate all planetary life. Venus experienced a runaway greenhouse in the past, and we expect that the Earth will in around 2 billion years as solar luminosity increases. But could we bring on such a catastrophe prematurely, by our current climate-altering activities? Here, we review what is known about the runaway greenhouse to answer this question, describing the various limits on outgoing radiation and how climate will evolve between these. The good news is that almost all lines of evidence lead us to believe that is unlikely to be possible, even in principle, to trigger full a runaway greenhouse by addition of non-condensable greenhouse gases such as carbon dioxide to the atmosphere. However, our understanding of the dynamics, thermodynamics, radiative transfer and cloud physics of hot and steamy atmospheres is weak. We cannot therefore completely rule out the possibility that human actions might cause a transition, if not to full runaway, then at least to a much warmer climate state than the present one. High climate sensitivity might provide a warning. If we, or more likely our remote descendants, are threatened with a runaway greenhouse, then geoengineering to reflect sunlight might be life’s only hope. Injecting reflective aerosols into the stratosphere would be too short-lived, and even sunshades in space might require excessive maintenance. In the distant future, modifying Earth’s orbit might provide a sustainable solution. The runaway greenhouse also remains relevant in planetary sciences and astrobiology: as extrasolar planets smaller and nearer to their stars are detected, some will be in a runaway greenhouse state.

Keywords: runaway greenhouse; radiation limit; water vapour; dangerous climate change; tipping point; Venus

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One contribution of 12 to a Discussion Meeting Issue ‘Geoengineering: taking control of our planet’s climate?’.

1. Introduction

We live in a time of rapid climate change. The warming that the Earth is experiencing now as a result of emissions of greenhouse gases from fossil fuel burning is unprecedented in human history, and the pace of the change is very rapid compared with most past climate change as deduced from palaeoclimate records. It is natural to ask just how bad the problem could be: could humanity's collective failure to control our influence on climate lead to warming so cataclysmic as to render the planet uninhabitable?

Wide attention has recently been drawn to such a concern by one of our most eminent climate scientists, James Hansen: '... if we burn all reserves of oil, gas, and coal, there's a substantial chance that we will initiate the runaway greenhouse. If we also burn the tar sands and tar shale, I believe the Venus syndrome [the runaway greenhouse] is a dead certainty' [1, p. 236]. In this study, we begin by reviewing the physical basis of the runaway greenhouse in order to directly address this issue. We then consider how geoengineering could be appropriate to address a threat such as a moist or runaway climate state. Lastly, we discuss the runaway greenhouse for other planets.

It is a common misconception that the runaway greenhouse is a simple extension of the familiar water vapour feedback. As the planet warms, more water evaporates. Water vapour is a greenhouse gas, so this enhances the warming (a positive feedback). Accelerating this, by implication, would boil the entire ocean. Historically, this possibility was discussed by Sagan [2] and Gold [3], but some of the earliest modern-era climate models [4] showed that while water vapour feedback is an important positive feedback for Earth, it is not a runaway feedback.

In fact, the physics of the runaway greenhouse is rather different to 'ordinary' water vapour feedback. There exist certain limits that set the maximum amount of outgoing thermal (longwave) radiation that can be emitted from a moist atmosphere. In the ordinary regime in which the Earth resides at present, an increase in surface temperature causes the planet to emit more radiative energy to space, which cools the surface and maintains energy balance. However, as a limit on the emission of thermal radiation is approached, the surface and lower atmosphere may warm, but no more radiation can escape the upper atmosphere to space. This is the runaway greenhouse: surface temperature will increase rapidly, finally reaching equilibrium again only when the surface temperature reaches around 1400 K and emits radiation in the near-infrared, where water vapour is not a good greenhouse gas. Along the way, the entire ocean evaporates.

Simpson [5] was the first to identify a radiation limit, finding a tropospheric limit, which was seen as a paradox. Komabayashi [6] and Ingersol [7] both analytically derived stratospheric limits that gave rise to the modern concept of the runaway greenhouse. Rasool & de Bergh [8] applied a grey model to Venus. Spectrally resolved numerical calculations of the full runaway were performed by Pollack [9], Watson *et al.* [10], Abe & Matsui [11] and Kasting [12]. Over two decades later, these 1980s era calculations have yet to be substantially revised. Nakajima *et al.* [13] used a grey radiative-convective model to develop the theoretical framework of the various radiation limits, which is followed here. Pierrehumbert [14] examined the local runaway greenhouse in Earth's tropics, highlighting the importance of dry air columns for cooling. Rennó [15] studied the transition towards runaway conditions (up to 330 K) using a radiative convective

model with an explicit hydrological cycle, finding multiple equilibria. Pujol *et al.* [16–18] elaborated a semi-grey radiative–convective model finding—for example, the existence of multiple equilibria during warming. Sugiyama *et al.* [19] examined the role of relative humidity. Ishiwatari *et al.* [20] conducted the only study of the runaway greenhouse in a general circulation model (GCM) to date, using a simplified model with the grey radiative transfer scheme from Nakajima *et al.* [13]. Ishiwatari *et al.* [21] compared these GCM results with an energy balance model to examine climate states from low-latitude glaciation to the runaway.

2. Physics of the runaway greenhouse

In this section, we review the physics of the runaway greenhouse in detail. Our theoretical framework follows Nakajima *et al.* [13]. While Nakajima *et al.* [13] were very formal and mathematical, our description here is more qualitative. We begin by describing three different ‘limits’ on the outgoing longwave radiation. While we find it useful to group these together under a single umbrella term, we emphasize that the physical character of them is rather different. Limit 1, the *black body upper limit*, is the maximum radiation that any planet of given surface temperature can emit; this varies with temperature. It does not lead to a runaway greenhouse, but is included for context. Limit 2, the *moist stratosphere upper limit*, is the maximum amount of radiation that can be transferred by a moist stratosphere; this is invariant with temperature, and so can lead to a runaway greenhouse. Limit 3, the *moist troposphere asymptotic limit*, is the amount of radiation that a thick, pure water vapour atmosphere will emit; the radiation from any thick water-rich atmosphere will tend towards this asymptotic limit. It is invariant with temperature, and so can lead to a runaway greenhouse. We then describe the changes that occur when increasing temperature in both runaway and moist greenhouses.

Understanding the runaway greenhouse requires familiarity with many aspects of atmospheric radiative transfer and thermodynamics. Given the interdisciplinary interest in this topic, we summarize the necessary background material in appendix A. For those who skip appendix A, most of our discussion refers to an atmosphere transparent to solar (shortwave) radiation and grey with respect to thermal (longwave) radiation. Longwave optical depth is measured from the top of the atmosphere. Most emission to space is from where optical depth is of order unity, so emission depends on the temperature structure in this region. In the context of Earth, we refer to water vapour as a condensible greenhouse gas, carbon dioxide as a non-condensable greenhouse gas and radiatively inactive gases as background gas.

(a) Radiation limit 1: optically thin atmosphere/black body upper limit

For an airless body (the Moon, for example), or an atmosphere with no greenhouse gases or clouds, radiative emission from the surface will escape to space unattenuated. The flux to space at the top of the atmosphere is the black body emission from the surface. This flux,

$$F_{\text{TOA}(\text{thin limit})}^{\uparrow} = \sigma T_{\text{surf}}^4, \quad (2.1)$$

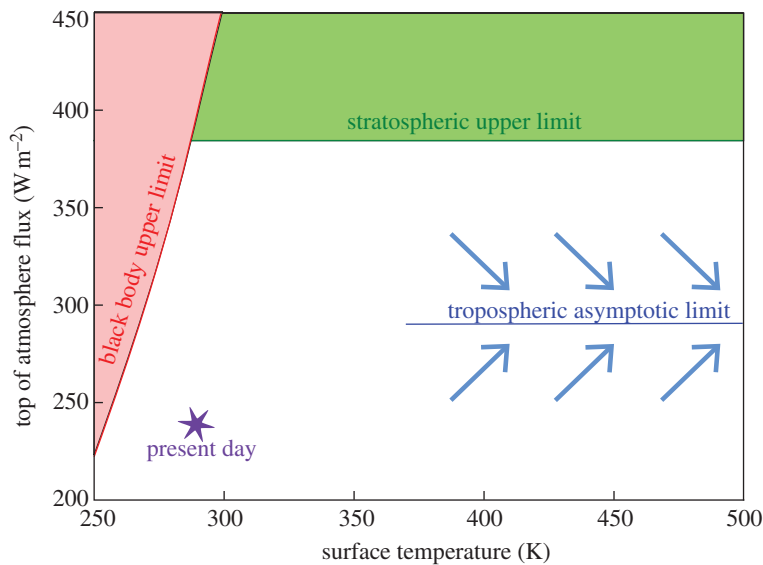


Figure 1. Radiation limits (solid lines) as a function of surface temperature, after Nakajima *et al.* [13]. Inaccessible regions are shaded. All of the white area can, in theory be occupied depending on the amount of non-condensable greenhouse gas (e.g. carbon dioxide) present, but at higher temperatures, outgoing flux will tend towards the tropospheric limit, as indicated by the arrows. Present day mean surface temperature is 289 K with an outgoing thermal flux of 239 W m^{-2} . Note that the change in temperature with flux is equivalent to the climate sensitivity, so the horizontal lines of the radiation limits imply infinite climate sensitivity, hence a runaway greenhouse.

is a radiation limit: adding any greenhouse gas (i.e. making the optical depth non-negligible) will cause less radiation to be emitted to space.

In figure 1, we show this and other radiative flux limits as a function of surface temperature. Because the outgoing thermal flux (plotted) must balance incoming solar flux, and if we assume that the atmosphere is transparent to shortwave, we can substitute absorbed solar flux for any given planet (about 239 W m^{-2} for Earth) for thermal flux and thus read surface temperature.

(b) *Radiation limit 2: moist stratosphere upper limit*
(Komabayashi–Ingersoll limit)

Historically, this is the limit discovered by the first workers to explicitly describe the runaway greenhouse [6,7], though as we will describe later, this limit is effectively impossible to achieve in practice on Earth. This limit is also the least intuitive.

Consider an atmosphere with a convective troposphere and a radiative equilibrium stratosphere, the boundary between which is the tropopause. With the atmosphere in contact with an ocean, the troposphere will be moist; to calculate a radiation limit, we assume that it is saturated with water vapour. Assuming a constant mixing ratio of stratospheric water vapour, the mass of this can be calculated directly from the partial pressure of water at the tropopause, which is the saturation vapour pressure for the tropopause temperature. We use

optical depth as the vertical coordinate. This depends, to first approximation, on the mass of the absorber, so the optical depth of the stratosphere is set by the amount of water vapour it contains, which is set by the tropopause temperature.

In radiative equilibrium, the upwards longwave flux must equal the downwards shortwave flux throughout. Two consequences of this are that the temperature increases with optical depth (this increase is rapid once optical depth $\tau > 0.1$) and that a higher absorbed shortwave flux will mean that the stratosphere is warmer throughout.

A limit on outgoing radiation from the stratosphere arises because the radiative equilibrium temperature profile needs to match the optical depth of the stratosphere that is set by the saturation vapour pressure at the tropopause. Intuitively, one might think first of the tropopause temperature, then of the associated outgoing flux, but the radiation limit is easier to demonstrate by starting with the outwards flux. For some given flux, we follow the radiative equilibrium temperature–optical depth profile until it intersects the temperature–optical depth curve derived from the saturation vapour pressure (figure 2). The intersection gives the tropopause. If the outgoing longwave flux is large, no solution can be found—the radiation limit is then the highest flux for which a solution can be found (around 385 W m^{-2} here).

One can understand this with a thought experiment in which we warm the tropopause. A cool tropopause means little water in the stratosphere, a small optical depth and only a small flux is required to maintain radiative equilibrium. Increasing the tropopause temperature increases the amount of water vapour and hence the optical depth of the stratosphere, requiring a larger outgoing flux to maintain equilibrium. We can think of this intuitively as a stable regime: the consequence of increasing temperature is an increased outgoing flux that would maintain equilibrium. This regime operates while the stratosphere is optically thin ($\tau < 0.1$), up to a tropopause temperature of about 255 K. At $\tau \approx 0.1$, the rate of increase of radiative equilibrium temperature with respect to optical depth becomes larger than the rate of increase of the tropopause temperature with respect to optical depth (from the saturation vapour pressure). Consequently, when tropopause temperature is increased, the new equilibrium has a lower outgoing longwave flux. This should be unstable: decreasing the outgoing flux would cause further warming, further decreasing the outgoing longwave radiation. Thus, we have an increase in the runaway temperature.¹

Thus, the simplest way to think of the Komabayashi–Ingersoll limit is to say that the stratosphere must be optically thin ($\tau \lesssim 0.1$), and that the radiation limit can be calculated assuming the tropopause is at $\tau \sim 0.1$.

(c) *Radiation limit 3: moist troposphere asymptotic limit*
(Simpson–Nakajima limit)

As the surface temperature rises, the amount of water vapour in the atmosphere increases, from water being a minor constituent (at present), to

¹In addition, as we move to an optically thick stratosphere, our physical assumptions break down. The temperature decrease with height would exceed the moist adiabatic lapse rate and convection would occur, causing the tropopause to move up. A very warm tropopause is unlikely in the real world.

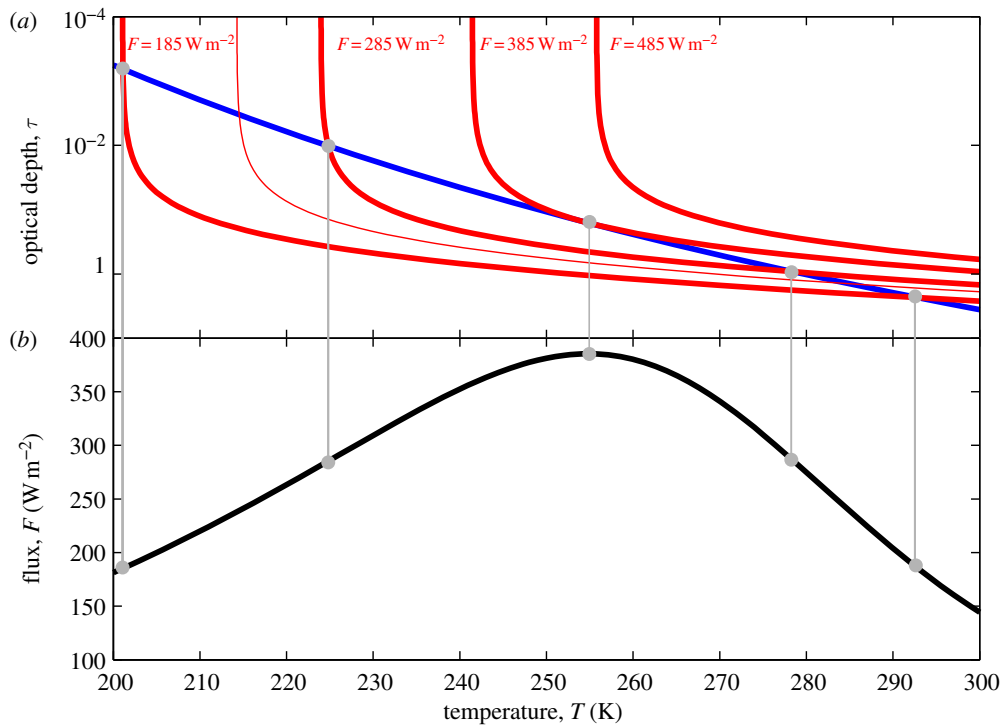


Figure 2. Top of atmosphere flux with tropopause temperature. (a) Reverse scale of optical depth on y -axis, so upwards on plot is increasing altitude. Bold red lines are solutions of the grey atmosphere model for various fluxes, showing the temperature profile of the stratosphere. Thin red line shows current top of atmosphere flux for comparison. The blue line is the optical depth of the stratosphere derived from saturated vapour pressure at the tropopause. Valid solutions for the tropopause are at intersections of red and blue lines. (b) Resulting change in top of atmosphere flux with temperature. Optical depth and grey atmosphere equations are $\tau_{\text{trop}} = k_v e_s(T_{\text{trop}})1/gM_v/\bar{M}$ and $\sigma T^4 = 1/2F_{\text{TOA}}^\uparrow(3/2\tau + 1)$, where k_v is a grey absorption cross-section for water vapour, $e_s(T_{\text{trop}})$ is the saturation vapour pressure at the tropopause temperature, g is acceleration owing to gravity, M_v/\bar{M} is the ratio of the molar masses of water vapour and the mean air, and T and τ are any temperature and optical depth pair on the profile [7,13].

a major constituent and ultimately to the dominant atmospheric constituent (figure 3). For example, the boiling point of water at the Earth's surface pressure of 10^5 Pa is 373 K (100°C). The boiling point is, by definition, the temperature at which the saturation vapour pressure of water equals the atmospheric pressure, so with a surface temperature of 373 K, the saturation vapour pressure of water is 10^5 Pa , the total surface pressure would be $2 \times 10^5 \text{ Pa}$ and the mixing ratio of water vapour would be 0.5 (now that surface pressure has increased, the boiling point would be higher than 373 K and the entire ocean will *not* boil: a liquid ocean remains until the critical point is reached at $T_c = 647 \text{ K}$).

In general, the tropospheric temperature structure follows the moist adiabatic lapse rate. As the water vapour mixing ratio tends towards 1, this simplifies to the saturation vapour–pressure curve. Raising the surface temperature will

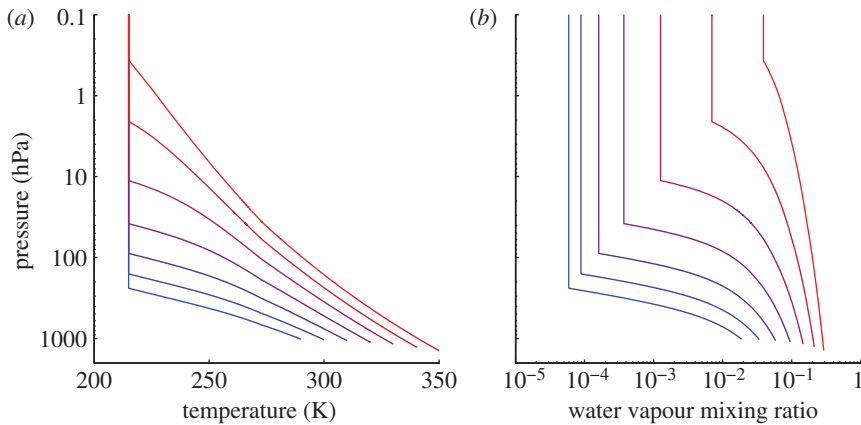


Figure 3. (a) Change in temperature structure and (b) moisture structure for warming atmospheres. A background pressure of 10^5 Pa is assumed, equal to that of Earth's atmosphere.

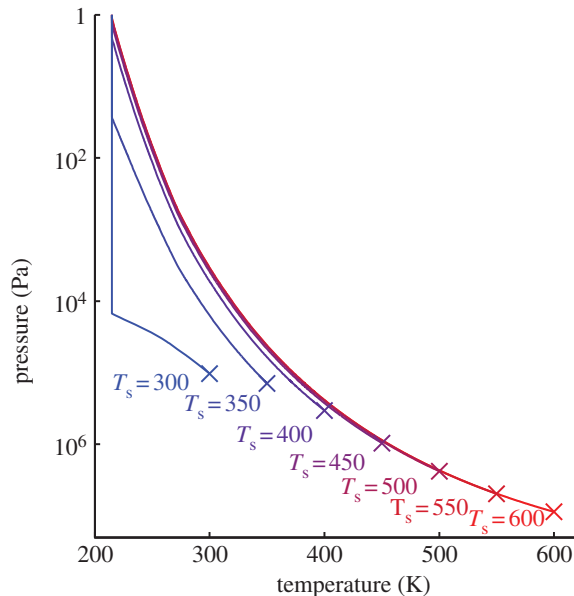


Figure 4. Temperature structure of the atmosphere with increasing surface temperature. A background pressure of 10^5 Pa is assumed, equal to that of Earth's atmosphere.

evaporate more of the ocean—we can think of this as adding extra gas to the bottom of the atmosphere—but will not change the overlying atmospheric structure (figure 4).

Effective emission to space takes place around $\tau = 1$. For the thick water vapour atmosphere we are now considering, this will be in the upper troposphere. Because the temperature structure is fixed, so too is the amount of radiative flux that can be emitted. Hence, a radiation limit emerges: increasing surface temperature adds extra gas to the bottom of the atmosphere and increases surface pressure, but

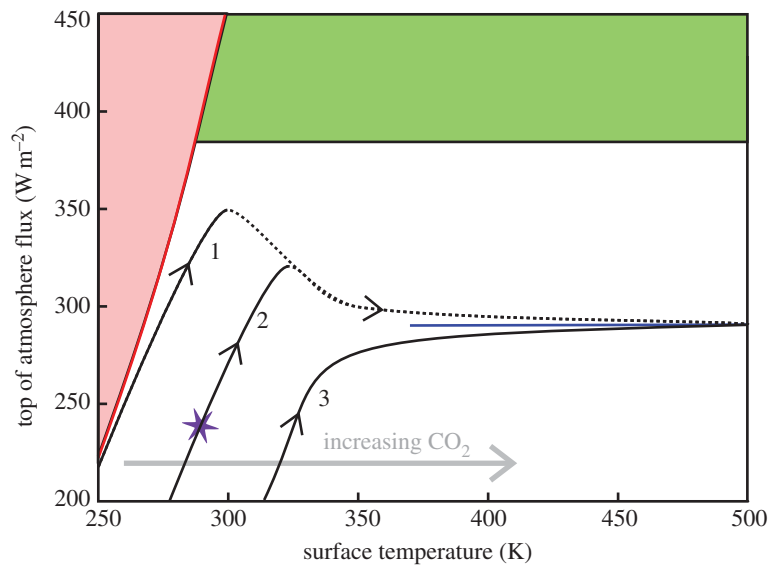


Figure 5. Increase in outgoing thermal flux as a function of surface temperature, after Nakajima *et al.* [13]. Black lines marked from 1 to 3 show how the top of atmosphere flux changes with increasing surface temperature for successively higher concentrations of a non-condensable greenhouse gas (e.g. carbon dioxide). Line 2 corresponds to Earth's present amount of non-condensable greenhouse gases and lines 1 and 3 are illustrative of lower and higher concentrations, respectively. All lines are for an amount of background, non-condensable and non-radiatively active gas similar to Earth's (see fig. 6 of Nakajima *et al.* [13] for other background gas inventories). Radiation limits are shown in colour (see figure 1 for labels). Fig. 9 of Kasting [12], derived from a spectrally resolved model, has similar features.

does not change the temperature structure of the upper troposphere from which radiation is emitted to space. The radiative flux to space is decoupled from surface temperature, with the radiation limit

$$F_{\text{TOA}(\text{trop limit})}^{\uparrow} \approx \sigma T_{\tau=1}^4. \quad (2.2)$$

We refer to this as the moist troposphere limit or Simpson–Nakajima limit, for Simpson [5] who first found the limit and for Nakajima *et al.* [13] who described the relevant physics.²

Calculating the value of this radiation limit requires numerical evaluation of a radiative transfer code. Example values are 290 W m^{-2} from a grey calculation [13] and 310 W m^{-2} from a spectrally resolved calculation [12].

Note that a cool atmosphere that is not pure water vapour, but contains some background non-condensable gas as Earth does will ‘overshoot’ the moist troposphere asymptotic limit (figure 5). This is because the presence of background gas decreases the water vapour mixing ratio, increasing the moist

²In practice, this was the limit found in previous numerical calculations [9–12], but was wrongly described there as the Komabayashi–Ingersoll limit that should correctly refer to the stratosphere. Also, Ingersoll [7] noted that a convective troposphere would not be able to deliver the required flux to meet the Komabayashi–Ingersoll limit, but did not sufficiently elucidate the physics.

adiabatic lapse rate (a smaller water mixing ratio permits a bigger moist adiabatic lapse rate because there is more non-condensable gas to take up latent heat of condensation from the water), such that the temperature around an optical depth of unity is higher and more radiation can be emitted. Also, with a larger lapse rate but a prescribed surface temperature, the vertically averaged atmospheric temperature is lower, so the column mass of water vapour is lower [13].

Of theoretical interest, but probably not relevant to Earth, is an analogous dry adiabatic limit. In the preceding discussion, we have considered an atmosphere where the background atmosphere (10^5 Pa for Earth) is orders of magnitude smaller than the amount of water vapour available (approx. 200×10^5 Pa for Earth), so the saturation water vapour mixing ratio in a warm atmosphere will always be big enough to make the moist adiabatic lapse rate markedly smaller than the dry adiabatic lapse rate. By contrast, an atmosphere with a very large reservoir of non-condensable gas (two or three orders of magnitude more than Earth; Venus, for example) would still have small water vapour mixing ratios in a warm atmosphere, so the moist adiabatic lapse rate would be no lower than the dry adiabatic lapse rate, owing to the large reservoir of non-condensable gas available to absorb the latent heat released. With the atmospheric profile tending towards the dry adiabat, the resulting radiation limit is higher than the moist adiabatic limit and is almost as high as the stratospheric limit [13].

(d) Increasing temperature

If we imagine increasing the solar flux absorbed by the planet, then the outgoing longwave flux must increase to compensate. How will the temperature adjust to provide this increased flux? The radiation limits provide guidance, but we must interpret radiative–convective model output (from Nakajima *et al.* [13]) to see how these limits interact (figure 5). It turns out, perhaps unsurprisingly, that the path surface temperature takes depends strongly on the amount of non-condensable greenhouse gas, such as carbon dioxide that is present.

If the atmosphere has a negligible amount of carbon dioxide, and a low surface temperature, then the atmosphere is optically thin and follows the black body limit. The temperature and outgoing flux increase together in this way while the atmosphere is not optically thick, overshooting the moist troposphere limit. The flux approaches, but never reaches, the stratospheric limit (or dry tropospheric limit). As the atmosphere becomes warmer, there is more water vapour aloft (see figure 3), so $\tau = 1$ is higher and thus colder, hence less flux is emitted to space. The outgoing flux then asymptotes to the moist troposphere limit (figure 5). Clearly, decreasing outgoing flux with increasing temperature is an unstable situation, so the onset of the runaway greenhouse is at the turning point of the flux with respect to temperature.

Now consider an increased carbon dioxide abundance. With a fixed absorbed solar flux, more carbon dioxide gives a warmer surface; this is the familiar greenhouse effect. With higher carbon dioxide, the outgoing flux overshoots the moist troposphere limit less. Thus initiation of the runaway greenhouse is at a higher surface temperature, but requires a lower flux. With a suitably large carbon dioxide inventory, there is no overshoot and the moist troposphere limit is approached monotonically (figure 5). In no case is the runaway initiated with a flux lower than the moist troposphere (Simpson–Nakajima) limit.

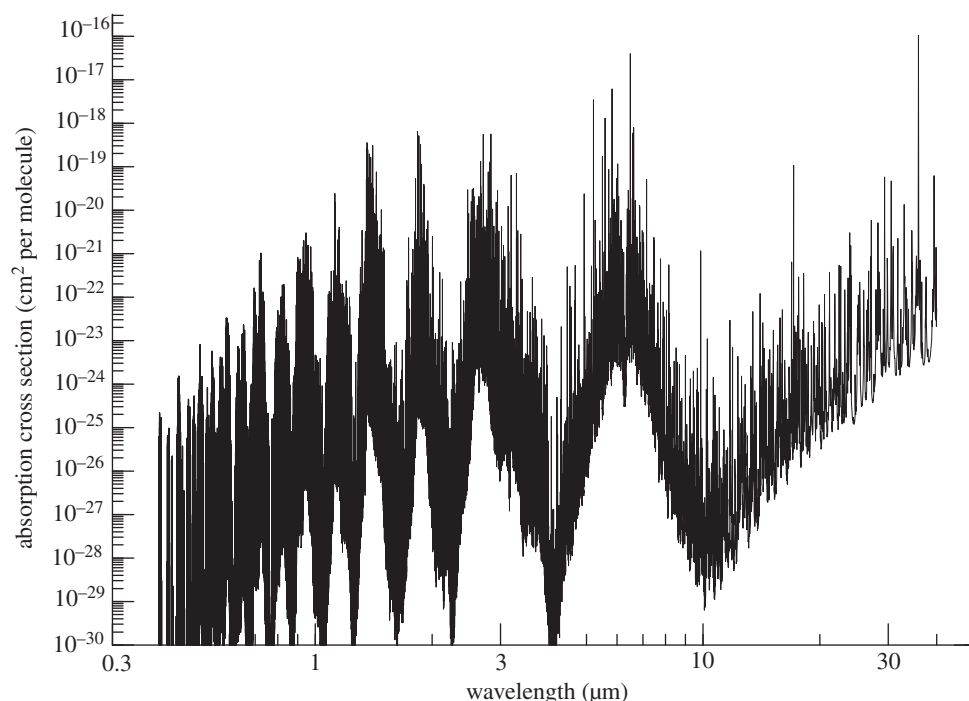


Figure 6. Absorption spectrum of water vapour (0.3–40 μm) shown at 220 Pa and 260 K. Note the ‘window’ regions where the absorption coefficient is low and the general decline of absorption coefficient at a shorter wavelength.

In the simplified system, where the atmosphere is transparent to solar radiation and grey with respect to thermal radiation, the runaway greenhouse occurs when increasing surface temperature is accompanied by constant or decreasing outgoing longwave flux. This implies that the surface should warm infinitely. There is an end however, and temperature will stabilize again. To explain this, we need to relax our assumption that water vapour is a ‘grey’ absorber: in reality, it is quite far from being so. The absorption spectrum of water vapour (figure 6) shows that there are major ‘window’ regions between 8 and 12 μm and around 4 μm , where absorption is weaker than elsewhere, so radiation can escape from the surface or lower atmosphere directly to space. Today, the wavelength of maximum emission is 10 μm and much of the radiation that escapes from Earth does so through the 8–12 μm window. With more water vapour in the atmosphere, the 8–12 μm window will become optically thicker, absorbing more radiation and preventing escape to space. However, as temperature increases, the wavelength of maximum emission decreases in an inverse proportion (Wien’s displacement law), and radiation can escape through the shorter wavelength window regions. This allows escape from the runaway greenhouse. Numerical calculations show that this occurs when the surface is around 1400 K [11,12] when the corresponding wavelength of maximum emission would be 5 μm . Making the analogy to our original grey case, allowing emission through the 4 μm window can be thought of as analogous to reducing the grey absorption coefficient of the atmosphere when the surface temperature exceeds 1400 K.

1400 K is well above the critical temperature of water (647 K) at which liquid and vapour phases cease to be distinct, so the oceans all evaporate during the runaway greenhouse (the exit temperature from the runaway is basically fixed by the position of water vapour windows, and so does not depend much on the volume of the oceans or whether they have all evaporated).

Relaxing the assumption that the atmosphere is transparent to shortwave radiation causes some qualitative changes as water vapour absorbs solar radiation. As the water vapour inventory increases with temperature, there is initially an increase in absorption (decrease in planetary albedo). However, as the atmosphere becomes thicker, increased Rayleigh (molecular) scattering dominates and net absorption decreases (planetary albedo increases). Consequently, the runaway greenhouse would be initiated later, as solar absorption is typically decreasing as thermal emission asymptotes [12].

(e) The moist greenhouse

So far, we have considered only the strict case of a runaway greenhouse. However, there exists an intermediate atmospheric state, the so-called moist greenhouse [12] in which water vapour is a major constituent of the troposphere and the stratosphere becomes moist (figure 3). For the troposphere, the explanation is simply the increase in saturation vapour pressure with temperature: with a 320 K surface, for example, the water vapour mixing ratio at the surface would be 10 per cent. Understanding the stratosphere requires slightly more physics. We can assume that the stratospheric water vapour mixing ratio is determined by the mixing ratio at the tropopause. This is known as the cold trap; the minimum in temperature causes a minimum in saturation vapour pressure that keeps water out of the stratosphere. However, with a warmer surface, the tropopause moves upwards and to a lower pressure, but stays at a fixed temperature. Water vapour mixing ratio is the ratio of the vapour pressure to ambient pressure, so with a lower ambient pressure, the tropopause mixing ratio becomes higher and the cold trap less effective. Increasing surface temperature from 290 to 300 K, 310 K or 330 K gives increases in the stratospheric water mixing ratio of factors of 1.5, 2.7 and 21, respectively.

In principle, a moist greenhouse is a stable climate—just one much hotter than today. This is not a cause for comfort however, as the differences relative to today would be profound. With a warming climate, the latitudinal temperature gradient would weaken (likely melting any year round ice) and the atmospheric circulation patterns would be quite different (compare figs 8 and 9 of Ishiwatari *et al.* [20]). Furthermore, the present stratosphere begins at around 100 hPa and the ozone maximum is around 10 hPa. In our simple model (figure 3), which tends to put the tropopause too low, by the time the surface temperature is 330 K, the tropopause height would have risen to engulf the ozone maximum. This could destroy the ozone layer via catalytic ozone destruction by water-derived hydrogen radicals [22,23]. Even before this, much more water in the stratosphere owing to a weaker cold trap will rather change the stratospheric chemistry. A moist greenhouse would also increase climate sensitivity. More water in the stratosphere will increase the rate of hydrogen escape [24], which, although not relevant on human time scales, is a major process in planetary evolution.

Getting to a moist greenhouse simply requires warming, not exceeding radiation limits: more greenhouse gases could certainly get us there. Kasting & Ackerman [25] found that somewhat over 10 000 ppmv carbon dioxide was required.

3. Could anthropogenic global change cause a moist or runaway greenhouse?

(a) *Runaway greenhouse*

Figure 5 illustrates how increasing the carbon dioxide inventory of the atmosphere affects the change in outgoing longwave radiation with temperature, using a grey atmosphere model. Presently, Earth's absorbed solar flux is smaller than all of the radiation limits described earlier, and the surface temperature adjusts so that outgoing longwave radiation matches the absorbed solar flux. More carbon dioxide means that the surface must be warmer to provide the same outgoing flux—this is the familiar greenhouse effect.

The runaway greenhouse occurs only when the outgoing longwave flux reaches a radiation limit. The fundamental point is that adding carbon dioxide does not increase the outgoing longwave flux, so it cannot cause a runaway greenhouse. While this result comes from simple models, a qualitatively similar result has been obtained from spectrally resolved radiative transfer codes (see fig. 9 of Kasting [12]): even 100 bar of carbon dioxide does not give a radiation limit [12,25].

And yet, because of the extreme seriousness of the consequences, this theoretical answer is insufficiently satisfying. While we can construct simple models to describe the limits of how the atmosphere behaves (and verify them with numerical calculations to an extent), we actually have little idea of exactly how the atmosphere will change when subject to extreme heating. Is there any missed physics or weak assumptions that have been made, which if corrected could mean that the runaway is a greater risk? We cannot answer this question with confidence.

One major uncertainty is relative humidity. We have assumed a saturated troposphere as the end member that makes a runaway greenhouse most likely, with the aim that uncertainty here would be how much *less* likely the runaway greenhouse would be. If Earth's tropics were saturated, they would be in a local runaway greenhouse, whereas in reality, columns of unsaturated air allow radiation to escape [14]. How relative humidity will change with warming is very poorly understood, but will have a first-order effect on temperature change. Explicit inclusion of a hydrological cycle in models has given rise to multiple equilibria [15], which adds an additional threat.

Likewise, we have not discussed clouds (the physics is hard enough without them). Presently, the greenhouse effect of clouds is about half the albedo effect [26]. A typical argument, following Kasting [12], is that in an optically thick atmosphere, the cloud greenhouse effect would diminish as non-cloud optical depth would be so high and the albedo effect would dominate further, making the runaway greenhouse less likely. This is not a watertight argument: for example, with much more water vapour aloft, high clouds might become thicker and more widespread, causing net warming. We do not know what will happen.

There are uncertainties in the radiative transfer too. The last full, spectrally resolved, treatments of the problem [11,12] are over two decades old. Much of the uncertainty relates to the so-called continuum absorption of water vapour (in window regions, both 8–12 μm and 4 μm , there is more absorption than one would expect considering nearby water vapour lines). This is probably caused by far-wing line absorption by water lines that depends on self-collisions of water molecules. Hence, this depends on vapour pressure squared, and is highly relevant to our problem. Understanding and empirical constraints on this continuum have improved, but remain unsatisfactory.

Are there alternate ways that the radiation limit could be reached? Could decreased albedo increase the absorbed solar radiation sufficiently? Taking the lowest radiation limit to be 285 W m^{-2} , one would need to reduce the planetary albedo from 0.3 to 0.16 to bring the absorbed solar radiation up to the radiation limit. In radiative forcing terms, this is 45 W m^{-2} . Changing all land surface to ocean (which is darker) would give a forcing of 5 W m^{-2} [27]. Clouds are more important: removing all low cloud would give a forcing of 25 W m^{-2} . Marine stratus depends strongly on boundary-layer processes that are not represented well in all models. While albedo changes alone would unlikely be sufficient to cause a runaway greenhouse, a major reduction in cloud reflection would pose a hazard. In a warmer climate, more atmospheric water vapour would increase the amount of incoming solar radiation absorbed in the atmosphere, also decreasing planetary albedo.

As a second alternative, we consider whether anything might lower the radiation limit. Given that the runaway greenhouse threshold depends on absorption in the 8–12 μm [12] and 4 μm [28] windows, it seems plausible that a greenhouse gas that absorbed strongly here (carbon dioxide does not) could lower the radiation limit. This has not been investigated. Alternatively, more extensive high cloud cover could block radiation through the windows (clouds *do* approximate well to a grey absorber). If such clouds had a larger particle size than present, their albedo could remain low.

Although it seems quite unlikely, a case of more widespread and larger droplet size high clouds or good gaseous absorbers in the water vapour window and a reduction in low clouds could, in principle, increase the absorbed solar radiation and decrease the radiation limit. Without better understanding whether such changes are possible, we cannot totally rule out a runaway greenhouse.

(b) *Moist greenhouse*

The earlier-mentioned considerations all apply to the case of a strict runaway greenhouse. A transition to a moist greenhouse (§2e) or other hot climate state is not excluded by theory and must be seen as a potential threat, until proved otherwise. In our understanding, this is the physically correct interpretation of the severe hazard of which Hansen [1] warns.

The question here is simply how much could human action increase the strength of the greenhouse effect? Kasting & Ackerman [25] found that, with carbon dioxide as the only non-condensable greenhouse gas, over 10 000 ppmv would be needed to induce a moist greenhouse. This is likely higher than could be achieved than by burning all the ‘conventional’ fossil fuel reserves—though the actual amount of fossil fuel available is poorly constrained, especially when one

includes ‘exotic’ sources such as tar sands (which are already being exploited) and methane clathrates. Greenhouse gases other than carbon dioxide, cloud or albedo changes could all contribute further warming. Likewise, the exhibition of multiple equilibria in the relevant temperature range [15,17] complicates matters.

(c) *Could we predict this?*

If we did approach a moist or runaway greenhouse, how would we know? If the approach to a radiation limit is monotonic, then increased climate sensitivity might serve as a warning: close to a limit, any forcing (e.g. from more carbon dioxide) would cause a much enhanced temperature change. To put bounds on this, models used for the Intergovernmental Panel on Climate Change reports generally give a climate sensitivity to doubling carbon dioxide of around 3 K with a likely range of 1.5–4.5 K. Ensemble experiments in which uncertain parameters are varied over reasonable ranges yield a distribution of sensitivities with a long tail out to over 10 K [29]. Such high sensitivity is inconsistent with our knowledge of palaeoclimate and the model cases that provide the extremes do not seem likely (due to poor representation of contemporary climate): thus if such a sensitivity was observed, it would indicate moving to a markedly different climate state. However, figure 5 (following Nakajima *et al.* [13]) and various other studies of the approach to the runaway greenhouse [15,17,19,20] indicate that there are strong nonlinearities and some bifurcation points (transitions between different steady states) and multiple equilibria on this approach. The approach to a runaway greenhouse is not sufficiently understood, so it is not obvious what we should look for as a warning.

Ideally, we would want numerical climate models to robustly resolve the transition to a much hotter atmosphere. However, most such models have been developed for fairly small perturbations from the existing climate, and their wider applicability may be limited by the obvious unavailability of data to tune the model to, and by simplifications made to reduce computational cost (e.g. the performance of the radiative transfer schemes used in some contemporary GCMs deteriorates at or soon after doubling carbon dioxide from present day concentrations [30,31]). Considering an atmosphere that substantial fractions of which can condense introduces many additional challenges. A new generation of model may well be needed.

4. Geoengineering solutions to a runaway greenhouse

Suppose that the runaway greenhouse did prove to be a risk, either in the foreseeable future for humanity (if our earlier-mentioned analysis is erroneous) or for our remote descendants as solar output increases. Geoengineering would be necessary to keep Earth habitable, but how would this best be done? Three principles might guide action: (i) the threat of a runaway greenhouse would come from external heating (probably solar), not from greenhouse gases, (ii) this would be a permanent threat, and (iii) climate sensitivity would be very high.

From point (i), it is clear that the geoengineering strategy would be to reduce the amount of sunlight absorbed. From points (ii) and (iii), it is equally clear that this strategy would need to be long-lived and robust. Temporary reflectance-altering strategies such as the addition of aerosols to the stratosphere might

provide a quick fix in dire emergency, but would not answer the need for permanence. A more robust approach would be to deploy sunshades at the inner Sun–Earth lagrange point (L1). A scheme described by Angel [32] would use millions of small sunshades at the L1 point that would block 1.8 per cent of the incident solar radiation (equivalent to about 6 W m^{-2} when averaged over the Earth). However, the small ‘flyers’ would also have a limited lifetime, probably only measured in decades [32], so constant maintenance would be required.

If the risk of a runaway greenhouse is distant in time, we can consider solutions beyond our present engineering capability. To counter the brightening Sun, one could move the Earth’s orbit outwards (astronomical engineering). Physically, this could be performed by altering the orbit of an asteroid to perform gravity assists around Earth and Jupiter, transferring orbital energy from Jupiter to Earth [33]. This would be ideal as it would directly address the cause of the problem and would not require continual maintenance.

5. The runaway greenhouse on other planets and exoplanets

The theory relating to the runaway greenhouse was developed in the planetary sciences, concerning the evolution of Venus and planetary habitability. Reaching a radiation limit is seen as setting a hard limit on the inner edge of the circumstellar ‘habitable zone’ (the region in which an terrestrial planet would likely to be habitable for life as we know it [34,35]) and is thus relevant for the design of exoplanet characterization missions.

A runaway greenhouse is caused by an increase in external heating. One heating source is solar: stellar luminosity increases with age (e.g. the Sun is 30% more luminous now than when it entered the main sequence), so time is sufficient. Alternately, if the orbit of a planet evolves inwards, a radiation limit could equivalently be reached. Theoretically, a decrease in albedo could mimic increased solar flux (e.g. deglaciation from total ice cover while solar flux is high [21]). The second commonly discussed heating source is a large asteroid impact [11]. Impact heating is effectively delivered instantly, giving an immediate transition to runaway conditions; these would be transient with the atmosphere slowly condensing afterwards.

For insolation-driven runaways (or a planet subject to relentless bolide impact), the ultimate escape from a runaway greenhouse would be loss of water to space by disassociation owing to photolysis and then escape of hydrogen to space. Given that the atmosphere would be dominated by water vapour, hydrogen escape would be energetically limited (hydrodynamic escape), and the time scale for loss of an Earth-size ocean of water would be a few hundred million years [36].

These considerations are not entirely hypothetical: Venus’ atmospheric state is thought to be explained by a runaway greenhouse and subsequent hydrogen escape. Its present atmosphere is 92 bars, consisting 96.5 per cent carbon dioxide and 3.5 per cent nitrogen. Water vapour is present at only approximately 30 ppmv [37]. Being nearer to the Sun, Venus receives nearly twice the energy from the Sun as Earth does, so will have exceeded the radiation limits at some point in its history. A hundredfold enrichment of the deuterium–hydrogen ratio on Venus relative to Earth is evidence of the subsequent hydrogen escape [38].

The radiation limits vary with the acceleration owing to gravity, g (bigger for heavier planets). We focus here on the moist troposphere asymptotic limit (Simpson–Nakajima limit), as it is the lowest and most commonly encountered limit. To first-order approximation, the optical depth of a layer depends on the mass of absorbing gas in that layer. With this approximation, the required mass of absorber in the layer between the top of the atmosphere ($\tau = 0$) and the effective emission level ($\tau = 1$) is constant. If g is increased, the pressure at the bottom of this layer (at $\tau = 1$) will similarly increase ($p = \text{force/area} = \text{mass} \cdot g/\text{area}$). With the vertical temperature structure following the saturation vapour–pressure curve, higher pressure at the effective emission level corresponds to a higher emission temperature, hence a higher limiting flux.³ A Mars-size planet would have a lower limiting flux and be more susceptible to a runaway greenhouse, whereas a few Earth-masses planet would have a higher limiting flux and be less susceptible to a runaway greenhouse.⁴

6. Conclusion

The runaway greenhouse is well defined in theory: as the surface warms, the atmosphere becomes optically thick with water vapour, which limits the amount of thermal radiation that can be emitted to space. If the planet absorbs more solar energy than this limit, runaway warming ensues. The practical limit on outgoing thermal radiation occurs when the atmospheric structure tends towards the saturation vapour pressure curve of water as the atmospheric composition tends towards pure water vapour, giving a limit of around 300 W m^{-2} . Earth presently absorbs around 240 W m^{-2} of solar radiation. Increasing carbon dioxide concentration will make surface warmer with the same outgoing thermal flux. Following this theory, we are not near the threshold of a runaway greenhouse. However, the behaviour of hot, water-vapour-rich atmospheres is poorly understood, and an in-depth study of these is necessary.

In the event that our analysis is wrong, we would be left with the situation in which only geoengineering could save us. The only useful methods would be those that would reduce the amount of solar energy that the Earth absorbs.

Lastly, we emphasize that nothing in this study should detract from the fact that anthropogenic greenhouse gas emissions are a major threat to global society and ecology. While we make the technical point that these emissions are unlikely to cause a ‘runaway greenhouse’, very severe and dangerous climate change is a real possibility. Abrupt change to a hot ‘moist greenhouse’ state, which though not technically a ‘runaway’ would nonetheless be dire, is not excluded. The imperative to cut greenhouse gas emissions remains.

³Optical depth does actually depend on pressure too owing to ‘pressure broadening’ of absorption lines. Thus, in a more complete treatment, the limiting flux varies more slowly with g than indicated here. See Pierrehumbert [28] for a quantitative treatment of this.

⁴A semi-analytical treatment for the moist troposphere limit has recently been developed by Pierrehumbert [28], by substituting the vertical structure given by the saturation vapour pressure curve into a two-stream radiative transfer equation. Coefficients for radiative transfer terms are tuned to runs of a radiative transfer code. This formulation will likely be useful for ad hoc calculations for applications of the problem (e.g. calculating a bound on the inner edge of the circumstellar habitable zone for variable size extra-solar planets)

We thank Kevin Zahnle and Ray Pierrehumbert for enlightening discussions on the runaway greenhouse, Kevin Zahnle, Naomi Vaughan, Megan Smith and Tyler Robinson for comments on the manuscript and Jim Kasting and an anonymous reviewer for constructive reviews. We also thank Richard Freedman for plotting figure 6. C.G. was funded by NASA Planetary Atmospheres programme (grant no. NNX11AC95G) and through the Virtual Planetary Laboratory Lead team of the NASA Astrobiology Institute. A.J.W. was supported by a Royal Society Research Professorship.

Appendix A. Technical background

In this appendix, we summarize the background physics required to understand the runaway greenhouse, for the benefit of non-experts. Key terms are emphasized for easy reference. Further background can be found in various textbooks. For example, Wallace & Hobbs [39] at an introductory level, Curry & Webster [40] for reference to thermodynamics and Pierrehumbert [28] for an accessible, intermediate level, coverage of radiative transfer and discussion of the runaway greenhouse.

We are concerned with how the atmosphere absorbs and emits radiation. Two basic properties here are the *absorptivity*, \mathcal{A} , and *emissivity*, ε . Absorptivity is the fraction of the incident radiation at any given wavelength that is absorbed by some atmospheric layer, and emissivity is the fraction of the maximum radiation that the layer could emit (the black body radiation, to be discussed shortly), which is actually emitted. For any given wavelength, absorptivity equals emissivity (*Kirchhoff's law*). A body that is a perfect absorber and emitter ($\mathcal{A} = \varepsilon = 1$) at all wavelengths is described as a *black body*. The wavelength distribution of emitted radiation from a black body depends solely on the temperature of the body. This distribution is called the *Planck function*. Two properties of this function are most relevant to the atmosphere. First, the integrated radiative emission (the radiative flux) is given as $F = \sigma T^4$ (the *Stefan-Boltzmann law*), where $\sigma = 5.67 \times 10^{-8}$ is the Stefan-Boltzmann constant and T is the temperature. Thus, the emitted radiative flux is a very strong function of temperature. Second, the peak in the Planck function (the wavelength of maximum emission) is inversely proportional to the temperature of the body (*Wien's displacement law*).

The Earth's surface and the Sun correspond roughly to black bodies with temperatures of 289 and 5900 K, which have their respective Wein peaks at $10\ \mu\text{m}$ (thermal infrared radiation) and $0.5\ \mu\text{m}$ (visible radiation). Hence, terrestrial radiation is often referred to as thermal or *longwave radiation*, and solar radiation as *shortwave radiation*. It is common to treat these two wavelength regions separately as their behaviour in the atmosphere is rather different. In the shortwave region, reflection dominates over absorption. Around 30 per cent of incident solar radiation is reflected by clouds, the surface and by molecules (*Rayleigh scattering*), but only a small fraction is absorbed. Neglecting this absorption is a common approximation. In the longwave, scattering (reflection) is negligible but absorption is very important. This absorption is carried out both by clouds and by gases (*greenhouse gases*). While radiative absorption has a very rich wavelength dependence, a lot of theoretical progress can be made with the simplifying assumption that each wavelength region (longwave and shortwave) can be described with a single emissivity. This is the *grey atmosphere* approximation, and is used for most of the discussion in this study.

Now, consider the energy balance of Earth. Absorbed solar radiation must equal emitted thermal radiation at the top of the atmosphere, so we can write $S/4(1 - \alpha) = \sigma T_e^4$, where $S = 1368 \text{ W m}^{-2}$ is the solar constant (the amount of solar energy incident at the top of the atmosphere), the one-quarter factor accounts for the ratio of the area of the disc on which solar energy is absorbed to the Earth's surface, $\alpha = 0.3$ is the *albedo* (the fraction of solar radiation reflected). We can solve this equation for the 'effective temperature' of Earth, $T_e = 255 \text{ K}$. The surface is much warmer than this, as it is heated by the greenhouse effect.

The general principle of the *greenhouse effect* is this: the atmosphere decreases in temperature with height, so we can consider the atmosphere to be a layer that is colder than the surface ($T_{\text{atm}} < T_{\text{surf}}$). This will absorb a fraction of the upwards travelling radiation from the surface (the fraction is the absorptivity). Recall that emissivity equals absorptivity. The atmospheric layer emits both upwards and downwards, dependent on its temperature (from the Stefan–Boltzmann law) and its emissivity, with $F^\downarrow = F^\uparrow = \epsilon \sigma T_{\text{atm}}^4$. Given that $T_{\text{atm}} < T_{\text{surf}}$, less energy is then emitted to space than would have been with no absorbing layer. Energy that is radiated downwards heats the surface. The strength of the greenhouse effect not only depends on the amount of greenhouse gas, but also the temperature of the layer of the atmosphere from which emission occurs. The colder the emitting layer, the better the greenhouse effect.

For the discussion of the runaway greenhouse, some more technical details are necessary. A key quantity is the *optical depth*, τ , of a layer or the atmosphere. This depends on the number of molecules in the layer on the path the radiation takes, N , and some absorption coefficient for the relevant absorbing gas, k , thus $\tau = kN$. The amount of radiation that is transmitted through a layer depends directly on this optical depth: $F_1 = F_0 e^{-\tau}$ (*Beer's law*), where F_0 and F_1 are the incident and transmitted fluxes, respectively. Hence, the amount of absorption depends on the logarithm of the amount of absorbing gas. Further, most of the absorption will happen around within an order of magnitude of optical depth of 1 (unity): $e^{-0.1} = 0.90$, $e^{-1} = 0.37$, $e^{-10} = 0.000045$.

It is conventional to measure the optical depth from the top of the atmosphere, with $\tau = 0$ at the top of the atmosphere, so most radiation is then emitted to space from around an optical depth of unity. The strength of the greenhouse effect thus depends on the temperature structure of the atmosphere around this level.

We need to distinguish between *condensible* and *non-condensible greenhouse gasses*. Most of the greenhouse gases in Earth's atmosphere (carbon dioxide is the dominant non-condensible greenhouse gas) require a much colder temperature than is ever experienced in order to condense: we refer to these as non-condensibles. Water vapour, by contrast, does condense at ambient pressures and so is referred to as a condensible greenhouse gas. There is a large reservoir of the condensed phase in the oceans. The pressure of vapour that is in equilibrium (no net evaporation or condensation) is the *saturation vapour pressure*. This depends exponentially on the ambient temperature. Warming Earth's surface will cause there to be much more water vapour in the atmosphere. (It is only because of the temperature of Earth's atmosphere that particular gases are condensible or non-condensibles. On Mars, which is colder than Earth, carbon dioxide is condensible. On Titan, a moon of Saturn, which is colder still, water behaves more like rock does on Earth, and methane is a condensible gas.)

The structure of the atmosphere is important. We generally consider an atmosphere with two vertical regions. The lower region (the *troposphere*) is characterized by large-scale vertical (convective) motions. Given the dominance of convection, the vertical temperature profile is set by the behaviour of rising air, following a *moist adiabatic lapse rate* (air cools as it rises, but part of this cooling is offset by latent heat release as water condenses from the rising air). In the upper region (the *stratosphere*), convection is negligible, and the temperature structure is determined solely by the balance of absorbed and emitted radiation; it is said to be in *radiative equilibrium*. The boundary between these layers is the *tropopause*.

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