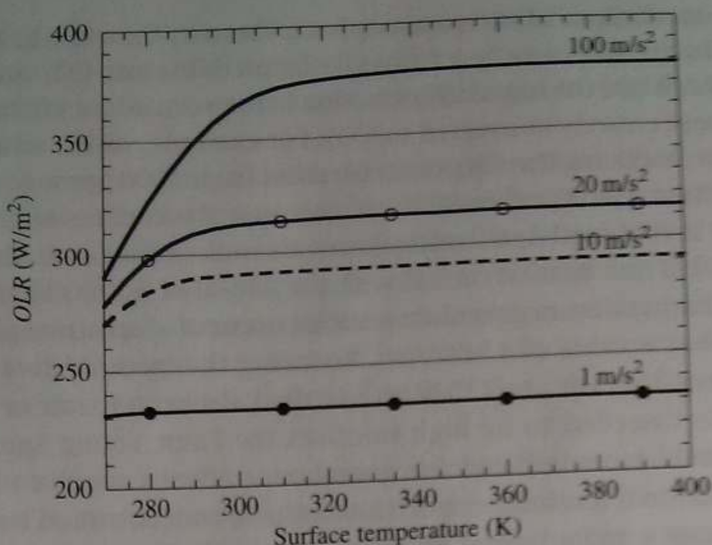
extremes of an all-methane or all-CO_2 atmosphere. The only case in which one can get a substantial reduction in greenhouse effect by oxidizing methane into CO_2 is when the initial CO_2 concentration is very high, the initial CH_4 fraction is between about 10% and 90% of the total, and the CH_4 is almost entirely converted to CO_2. For example, with a total carbon concentration of 10 000 ppmv, reducing the CH_4 concentration from 1000 ppmv to 1 ppmv reduces the greenhouse effect from 104 W/m^2 to 80 W/m^2. Because the curve is so flat, starting from an atmosphere which is 80% methane works almost as well: in that case the greenhouse effect is reduced from 101 W/m^2 to 80 W/m^2. If we have a total of 100 000 ppmv of carbon in the atmosphere, then the maximum greenhouse effect occurs for an atmosphere which is about 25% methane, and has a value of 151 W/m^2. Reducing the methane to 1 ppmv brings down the greenhouse effect 36 W/m^2, to 115 W/m^2. In the Paleoproterozoic or Archean, when the net greenhouse effect needed to be high to offset the Faint Young Sun, it is possible that a methane crash could have reduced the greenhouse effect enough to initiate a Snowball, but it is essential that in a methane crash, the methane concentration be brought almost all the way down to zero; a reduction of methane from 50% of the atmosphere to 10% of the atmosphere would not do much to the greenhouse effect. By the time of the Neoproterozoic, when the solar luminosity is higher and less total greenhouse effect is needed to maintain open water conditions, it is far less likely that a methane catastrophe could have initiated glaciation. Some further remarks on atmospheric transitions that could initiate a Snowball will be given in Chapter 8.

4.6 ANOTHER LOOK AT THE RUNAWAY GREENHOUSE

We are now equipped to revisit the runaway greenhouse phenomenon, this time using the absorption spectrum of actual gases in place of the idealized gray gas employed in Section 4.3.3. The setup of the problem is essentially the same as in the gray gas case. We consider a condensable greenhouse gas, optionally mixed with a background gas which is transparent to infrared and non-condensing. A surface temperature T_g is specified, and the corresponding moist adiabat is computed. The temperature and the greenhouse gas concentration profiles provide the information necessary to compute the *OLR*, in the present instance using the homebrew exponential sums radiation model in place of the gray gas *OLR* integral. As before, the *OLR* is plotted as a function of T_g for the saturated atmosphere, and the Kombayashi–Ingersoll limit is given by the asymptotic value of *OLR* at large surface temperature.

We'll begin with water vapor. Figure 4.37 shows the results for a pure water vapor atmosphere, computed for various values of the surfaced gravity. The overall behavior is very similar to the gray gas result shown in Fig. 4.3: the *OLR* attains a limiting value as temperature is increased, and the limit – defining the absorbed solar radiation above which the planet goes into a runaway state – becomes higher as the surface gravity is increased, and for precisely the same reasons as invoked in the gray gas case. The result, however, is now much easier to apply to actual planets, since with the real gas calculations we have the real numbers in hand for water vapor, and not for some mythical gas characterized by a single absorption coefficient. Several specific applications will be given shortly.

As in the gray gas case the limiting *OLR* increases as surface gravity is increased. It would be useful if the real gas result could be represented in terms of an equivalent gray gas, but first one must generalize Eq. (4.39) to incorporate the increase of absorption coefficient with pressure. This is done formally in Problem 4.5, but the qualitative derivation runs as



easy to runaway on a small planet. See also Fig. 4.3

Figure 4.37 *OLR* vs. surface temperature for a saturated pure water vapor atmosphere. The numbers on the curve indicate the planet's surface gravity. The calculation was done with the homebrew exponential sums radiation code, incorporating both the 1000 cm^{-1} and 2200 cm^{-1} continua, but neglecting temperature scaling of absorption outside the continua. Twenty terms were used in the exponential sums, and wavenumbers out to 5000 cm^{-1} were included; the atmosphere was considered transparent to higher wavenumbers.

follows. We need to determine the pressure p_1 where the optical thickness to the top of the atmosphere is unity, and then evaluate the temperature at that point, along the one-component saturated adiabat. For linear pressure broadening or the continuum the optical thickness requirement implies $\frac{1}{2}\kappa_0 p_1^2/p_0 g = 1$, where p_0 is the reference pressure to which the absorption is referred (generally 100 mb for the data given in our survey of gaseous absorption properties). Substituting the resulting p_1 into the expression for $T(p)$ we infer an expression of the form

$$OLR_\infty = A'\sigma T(p_1)^4 = A'\frac{\sigma(L/R)^4}{\left[\ln\left(p^*/\sqrt{2p_0 g/\kappa_0}\right)\right]^4}$$

where p^* is defined as before and A' is an order unity constant which depends on L/R. An examination of the g dependence of the calculated Kombayashi–Ingersoll limit in Fig. 4.37 shows that over the range $1\,\mathrm{m/s^2} \le g \le 100\,\mathrm{m/s^2}$, the numerically computed dependence can be fit almost exactly with this formula if we take $A' = 0.7344$ and $\kappa_0 = 0.055$ (assuming $p_0 = 10^4$ Pa). Though the pressure dependence of absorption causes the limiting *OLR* to vary more slowly with g than was the case for constant κ, the limit in the real gas case otherwise behaves very much like an equivalent gray gas with $\kappa_0 = 0.055$. This is a surprising result, given the complexity of the real gas absorption spectrum. The fact that the equivalent absorption is similar to that characterizing the 2500 cm^{-1} continuum suggests that the limiting *OLR* is being controlled primarily by this continuum. Thus, the behavior of this continuum is crucial to the runaway greenhouse phenomena (see Problem 4.25). It cannot be ruled out that other continua may affect the *OLR* as temperature is increased to very high values. For example, the total blackbody radiation at wavenumbers greater than 5000 cm^{-1} is only 1.14 W/m^2 at 500 K, so it matters little what the absorption properties are in that part of the spectrum at 500 K or cooler. However, when the temperature is raised to 600 K the shortwave emission is $15\,\mathrm{W/m^2}$ and so the shortwave absorption begins to matter; by the time T reaches 700 K the shortwave blackbody emission is $106\,\mathrm{W/m^2}$ and the shortwave emission properties are potentially important. On the other hand, at such temperatures there is so much water vapor in the atmosphere that a very feeble absorption would be sufficient to eliminate the contribution to the *OLR*.

Exercise 4.12 Verify the shortwave blackbody emission numbers given in the preceding paragraph by using numerical quadrature applied to the Planck function.

Exercise 4.13 Compute the Kombayashi–Ingersoll limit for water vapor on Mars, which has $g = 3.71\,\mathrm{m/s^2}$. Compute the limit for Titan, which has $g = 1.35\,\mathrm{m/s^2}$, and Europa which has $g = 1.31\,\mathrm{m/s^2}$.

Now let's generalize the calculation, and introduce a non-condensing background gas which is transparent in the infrared; we use N_2 in this example, though the results are practically the same if we use any other diatomic molecule. The background gas affects the Kombayashi–Ingersoll limit in two ways. First, the pressure broadening increases absorption, which should lower the limit. Second, the background gas shifts the lapse rate toward the dry adiabat, which is much steeper than the single-component saturated adiabat. The increase in lapse rate in principle could enhance the greenhouse effect, but given the condensable nature of water vapor, it actually reduces the greenhouse effect, because the low temperatures aloft sharply reduce the amount of water vapor there. If the background gas were itself a greenhouse gas, this effect might play out rather differently. At sufficiently high temperatures, water vapor will dominate the background gas and so the limiting *OLR* at high temperature will approach the pure water vapor limit shown in Fig. 4.37. However, for intermediate temperatures, the background gas can modify the shape of the *OLR* curve.

Results for various amounts of N_2 are shown in Fig. 4.38. As expected, at large temperatures the limiting *OLR* asymptotes to the value for a pure water vapor atmosphere. This can be seen especially clearly for the case with only 100 mb of N_2 in the atmosphere; with more N_2, one has to go to higher temperatures before the water vapor completely dominates the *OLR*, but the trend is clear. A very important qualitative difference from the pure water vapor case is that the *OLR* curve for a binary mixture shows a distinct maximum at intermediate temperatures. This maximum arises because the foreign broadening of water vapor absorption features is relatively weak, while the presence of a non-condensing background gas steepens the lapse rate and reduces the amount of water vapor aloft. The hump in the curve means that the surface temperature exhibits *multiple equilibria* for a given absorbed solar radiation. For example, with 100 mb of N_2, if the absorbed solar radiation is $320\,\mathrm{W/m^2}$ there is a cool equilibrium with $T_g = 288\,\mathrm{K}$ and a hot equilibrium with $T_g = 360\,\mathrm{K}$. The latter is an unstable equilibrium; displacing the temperature in the cool direction will cause water vapor to condense and *OLR* to increase, cooling the climate further until the system falls into the cool equilibrium. Conversely, displacing the temperature slightly to the warm side of the hot equilibrium will cause the climate to go into a runaway state. For these atmospheric parameters, the planet is in a metastable runaway state. The climate will persist in the cooler non-runaway state unless some transient event warms the planet enough to kick it over into the runaway regime. It is only when the absorbed solar radiation is increased to the maximum *OLR* at the peak ($328\,\mathrm{W/m^2}$) that a runaway becomes *inevitable*. For future use, we'll note that a calculation with $g = 10\,\mathrm{m/s^2}$ and 1

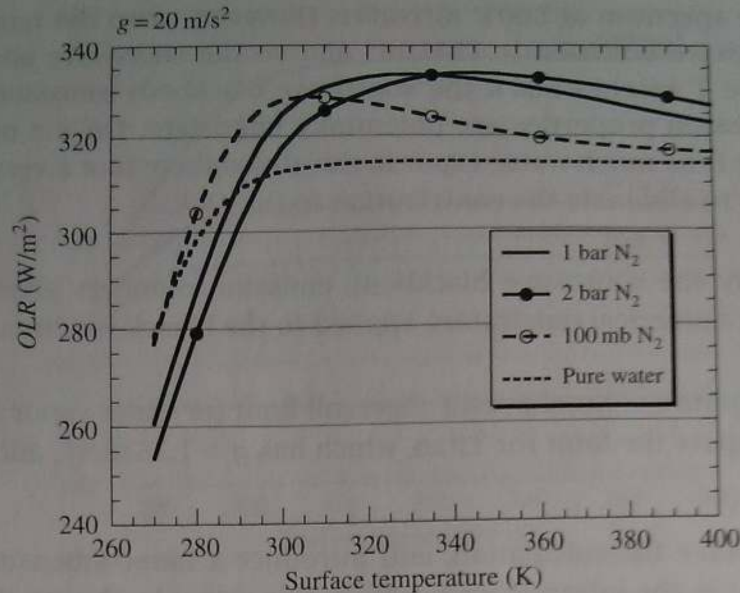


Figure 4.38 As for Fig. 4.37 but for a mixture of water vapor in N_2 on the saturated moist adiabat. Calculations were carried out with a surface gravity of $20\,m/s^2$, for the indicated values of N_2 partial pressure at the ground.

bar of N_2 has a peak *OLR* of $310\,W/m^2$ at $T_g = 325\,K$, while a case with the same gravity and 3 bars of N_2 likewise has a peak at $310\,W/m^2$ but the position of the peak is shifted to 360 K. The corresponding parameters for the slightly lower surface gravity of Venus differ little from these numbers. Both cases asymptote to the *OLR* for pure water vapor when the temperature is made much larger than the temperature at which the peak *OLR* occurs.

- *Runaway greenhouse on Earth*: With present absorbed solar radiation (adjusted for the cloud effects) of $265\,W/m^2$, the Earth at present is comfortably below the Kombayashi-Ingersoll limit for a planet of Earth's gravity. According to Eq. (1.1), as the solar luminosity continues to increase, the Earth will pass the $291\,W/m^2$ threshold where a runaway becomes possible in about 700 million years. In 1.7 billion years, it will pass the $310\,W/m^2$ threshold where a runaway becomes inevitable for an atmosphere with 1 bar of N_2 and no greenhouse gases other than water vapor.
- *Venus*: The present high albedo of Venus is due to sulfuric acid clouds that would almost certainly be absent in a less dry atmosphere. If we assume an Earthlike albedo of 30% then very early in the history of the Solar System, the absorbed solar radiation of Venus would be $327\,W/m^2$. This is just barely in excess of the mandatory runaway threshold of $310\,W/m^2$ for a planet of Venus' surface gravity, assuming a bar or two of N_2 in the atmosphere and no greenhouse gases other than water vapor. It is thus possible that neglected effects (clouds, subsaturation, a higher albedo surface) could allow Venus to exist for a while in a hot, steamy but non-runaway state with a liquid ocean. The high water vapor content of the upper atmosphere would still allow an enhanced rate of photodissociation and escape of water to space. If Venus indeed started life with an ocean, however, it is plausible that it eventually succumbed to a runaway state, since with the present solar constant the absorbed solar radiation without sulfuric acid clouds would be $457\,W/m^2$, well in excess of the runaway threshold.

- *Gliese 581c*: We can now improve our earlier estimates of the conditions on the extrasolar planet Gliese 581c, which has an absorbed solar radiation of 583 W/m^2 assuming a rocky surface. This flux is well above the threshold of 334 W/m^2 for a mandatory runaway for a planet with twice Earth's surface gravity, even allowing for 2 bar of N_2 in the atmosphere. Thus, if Gliese 581c ever had an ocean it is likely to have gone into a runaway state; if the composition of the planet included a substantial amount of carbonate in the interior, subsequent outgassing is likely to have turned it into a planet rather like Venus. It still remains, however, to assess the implications of the increased atmospheric absorption due to the greater proportion of infrared output of the M-dwarf host star.
- *Evaporation of icy moons in Earthlike orbit*: It has been suggested that icy moons like Europa or Titan could become habitable if the host planet were in an orbit implying Earthlike solar radiation. The low Kombayashi–Ingersoll limit for bodies with low surface gravity puts a severe constraint on this possibility, however. With the albedo of ice, such bodies could exist as Snowballs in an Earthlike orbit, but if the surface ever thawed, or failed to freeze in the first place, the absorbed solar radiation corresponding to an albedo of 20% would be 274 W/m^2 – well above the runaway threshold of 232 W/m^2 for a body with surface gravity of 1 m/s^2. Small icy moons in Earthlike orbits are thus likely to evaporate away, unless they are locked in a Snowball state.
- *Lifetime of a post-impact steam atmosphere*: Suppose that in the Late Early Bombardment stage, enough asteroids and comets hit the Earth to evaporate 10 bars worth of the ocean and give the Earth a 10 bar atmosphere consisting of essentially pure water vapor (and a surface temperature in excess of 440 K, according to Clausius–Clapeyron). How long would it take for the steam atmosphere to rain out and the temperature to recover to normal? To do this problem, we assume that the atmosphere remains saturated as it cools, and loses heat at the maximum rate given by the Kombayashi–Ingersoll limit for Earth; we also need to subtract the absorbed solar radiation from the heat loss. For Early Earth conditions, the net heat loss is about 100 W/m^2. On the other hand, the latent heat per square meter of the Earth's surface in a steam atmosphere with surface pressure p_s is Lp_s/g, or $2.5 \cdot 10^{11}$ J/m^2 for the stipulated atmosphere. To remove this amount of energy at a rate of 100 W/m^2 would take $2.5 \cdot 10^9$ s, or 80 years. The rainfall rate during this time would be warm but gentle: 3.5 (kg/m^2)/day, or a mere 3.5 mm/day based on the water density of 1000 kg/m^3. This is the average rainfall rate constrained by the rate of radiative cooling, but it is likely that at places the local rainfall rate could be orders of magnitude greater, owing to the lifting and condensation in storms and other large scale atmospheric circulations.
- *Freeze-out time of a magma ocean*: In Chapter 1 we introduced the problem of the freeze-out time of a magma ocean on the Early Earth, and estimated the time assuming a transparent atmosphere in Problem 3.24. How long does it take for the magma ocean to freeze-out if the planet is sufficiently water-rich that the atmosphere consists of essentially pure water vapor in saturation? The time is estimated in the same way as in Problem 3.24 except that the rate of heat loss is again taken to be the difference between the Kombayashi–Ingersoll limit – giving the maximum *OLR* – and the rate of absorption of solar energy. For the Early Earth this would be about 100 W/m^2, which is far smaller than the transparent atmosphere case, where the energy loss is nearly 100 000 W/m^2 based on σT^4 for the 2000 K temperature of molten magma. As a result, the freeze-out time (using the same assumptions as in Problem 3.24) increases to 3.5 million years.

The latter two estimates follow the line of thinking introduced by Norman Sleep of Stanford University, and again illustrate the principle that Big Ideas come from simple models.

Exercise 4.14 Estimate the lifetime of a post-impact pure water vapor atmosphere on Mars assuming that the planet absorbs 90 W/m^2 of solar radiation, per unit surface area. Estimate the precipitation rate, in mm of liquid water per day.

The above results presume that the saturated, condensable greenhouse gas is the only greenhouse gas present in the atmosphere. What happens if the atmosphere also contains a non-condensable greenhouse gas, whose total mass remains fixed as the surface temperature increases? For Earthlike or Venuslike conditions, for example, one would typically need to consider atmospheres consisting of a mixture of condensable water vapor in saturation, non-condensing CO_2, and perhaps a transparent non-condensing background gas such as N_2. Can the addition of CO_2 in this situation trigger a runaway greenhouse when water vapor alone could not support a runaway? We will not pursue detailed radiative calculations of this sort, but some simple qualitative reasoning, summarized in the sketch in Fig. 4.39, suffices to map out the general behavior. The essential insight is that, at sufficiently high temperatures, the atmosphere is completely dominated by the condensable component, whose mass increases exponentially with temperature. Hence, the Kombayashi–Ingersoll limit will be unaffected by the addition of the non-condensable greenhouse gas. However, as more and more non-condensable greenhouse gas is added to the atmosphere, one must go to ever-higher temperatures before the limiting *OLR* is approached. At lower temperatures, the addition of a large amount of non-condensing greenhouse gas brings down the *OLR* below the Kombayashi–Ingersoll limit. Whether or not this triggers a runaway depends on the details of the situation. If the *OLR* curve without the non-condensable greenhouse gas is essentially monotonic in temperature, as in the one-component cases in Fig. 4.37, then the addition of the non-condensable greenhouse gas warms the planet, but does not trigger a

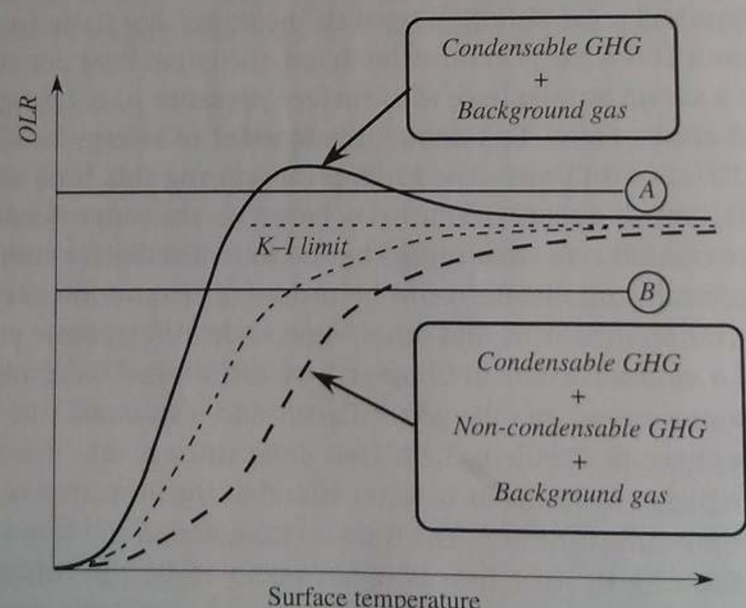


Figure 4.39 Qualitative influence of a non-condensable greenhouse gas (GHG) on the shape of the *OLR* curves. The upper curve gives the *OLR* for an atmosphere consisting of a mixture of a saturated condensable greenhouse gas with a non-condensing transparent background gas, as in the N_2/H_2O case shown in Fig. 4.38, while the lower curve illustrates how the behavior would change if a large amount of non-condensable greenhouse gas were added. The intermediate curve gives the *OLR* for a one-component saturated greenhouse gas atmosphere as in Fig. 4.37.

runaway if the absorbed solar radiation is below the Kombayashi-Ingersoll limit. However, in a case like Fig. 4.38, in which the *OLR* curve overshoots the limit and has a maximum, the addition of the non-condensable can eliminate the hump in the curve, eliminating the stable non-runaway state and forcing the system into a runaway. In the sketch, this situation is illustrated by the absorbed solar radiation line labeled "A." In that case, the addition of the non-condensing gas can indeed force the system into a runaway state. On the other hand, if the absorbed solar radiation is below the Kombayashi-Ingersoll limit, as in the line labeled "B," then the addition of the non-condensable greenhouse gas warms the equilibrium but does not trigger a runaway.

The concepts of runaway greenhouse and the Kombayashi-Ingersoll limit generalize to gases other than water vapor. For example, consider a planet with a reservoir of condensed CO_2 at the surface, which may take the form of a CO_2 glacier or a CO_2 ocean, according to the temperature of the planet. Specifically, if the surface temperature is above the triple point of 216.5 K the condensable reservoir takes the form of a CO_2 ocean; otherwise it takes the form of a dry-ice glacier. If the atmosphere is in equilibrium with the surface reservoir and has no other gases in it besides the CO_2 which evaporates from the surface, then one can use the one-component adiabat with the homebrew radiation code to compute an *OLR* curve for the saturated CO_2 atmosphere which is analogous to the water vapor result shown in Fig. 4.37. Results, for various surface gravity, are shown in Fig. 4.40. The general behavior is very similar to that we saw for water vapor, but the whole system operates at a lower temperature and the *OLR* reaches its limiting value at a much lower temperature than was the case for water vapor.

The CO_2 runaway imposes some interesting constraints on the form in which CO_2 could exist on Mars, both present and past. For Martian surface gravity, the Kombayashi-Ingersoll limit for CO_2 is a bit over $63\,W/m^2$. In consequence, when the absorbed solar radiation exceeds this value, a permanent reservoir of condensed CO_2 cannot exist at the surface of

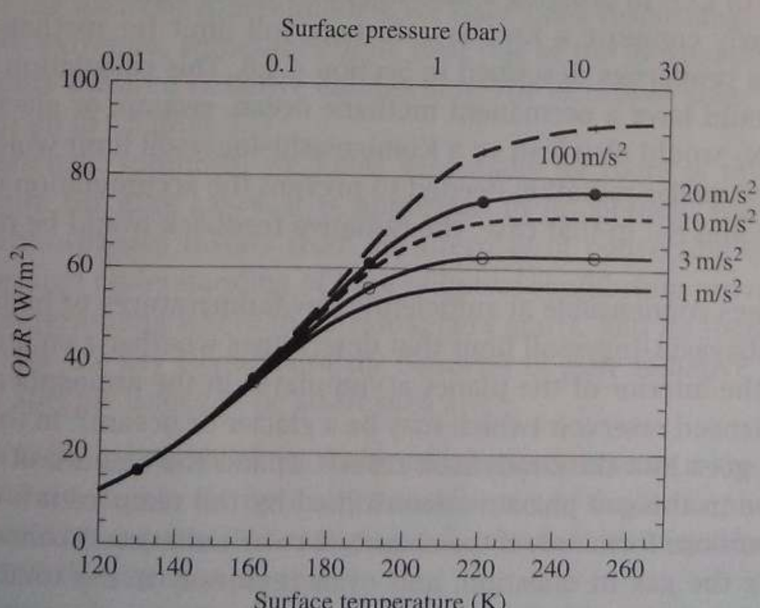


Figure 4.40 *OLR* vs. surface temperature for a saturated pure CO_2 atmosphere. Calculations were performed with the values of surface gravity indicated on each curve. The scale at the top gives the surface pressure corresponding to the temperature on the lower scale.

the planet; it will sublimate or evaporate into the atmosphere, and continue to warm the planet until all the condensed reservoir has been converted to the gas phase. At present, the globally averaged solar absorption is about $110\,W/m^2$, so the planet is well above the runaway threshold for CO_2. From this we can conclude that Mars cannot at present have an appreciable permanent reservoir of condensed CO_2 which can exchange with the atmosphere. Note, however, that this does not preclude the temporary buildup of CO_2 snow at the surface. Such deposits can and do form near the winter poles, but sublimate back into the atmosphere as spring approaches. This situation can be thought of as arising from the fact that the *local* absorbed solar radiation near the winter pole is below the Kombayashi–Ingersoll limit for CO_2. The local reasoning applies because the thin atmosphere of present Mars cannot effectively transport heat from the summer hemisphere.

Even without the albedo due to a thick CO_2 atmosphere, Early Mars would have an absorbed solar radiation of only $77\,W/m^2$. This is still somewhat above the Kombayashi–Ingersoll limit for a pure CO_2 atmosphere, but allowing for the scattering effects of the atmosphere and perhaps also the influence of nitrogen in the atmosphere, Early Mars could well have sustained permanent CO_2 glaciers, given a sufficient supply of CO_2. Because the planet is so near a threshold, a more detailed calculation – probably involving consideration of horizontal atmospheric heat transports – would be needed to resolve the issue.

In fact, the Kombayashi–Ingersoll limit for CO_2 is a critical factor in the determination of the outer limit of the habitable zone. It shows why you cannot make a planet in an arbitrary far orbit habitable simply by pumping enough CO_2 into the atmosphere until the greenhouse effect provides sufficient warming – if the absorbed stellar flux is below the threshold, extra CO_2 goes into the condensed reservoir instead of adding to the greenhouse warming. Early Mars is near the brink, which is why small exotic effects such as the radiative effects of CO_2 ice clouds can make the difference to habitability. Gliese 581c is in the same radiative regime. A planet in a much dimmer orbit, however, would need to get its habitability from a less condensable greenhouse gas; it is unclear whether any likely atmospheric constituent can take the place of CO_2 in bringing a distant planet to the liquid-water threshold.

One can similarly compute a Kombayashi–Ingersoll limit for methane, using the continuum absorption properties described in Section 4.4.8. This calculation would determine whether a body could have a permanent methane ocean, swamp, or glacier at the surface. Condensation of N_2 would also lead to a Kombayashi–Ingersoll limit which determines the threshold absorbed stellar radiation needed to prevent the accumulation of a surface reservoir of N_2 ice or liquid N_2. In that case, the radiative feedback would be provided by the N_2 self-continuum.

Any gas becomes condensable at sufficiently low temperatures or high pressures, and it is in fact the Kombayashi–Ingersoll limit that determines whether a volatile greenhouse gas outgassing from the interior of the planet accumulates in the atmosphere, or accumulates as a massive condensed reservoir (which may be a glacier or ocean).[9] In the latter case, additional outgassing goes into the condensed reservoir, and the amount of volatile remaining in the atmosphere in the gas phase is determined by the temperature of the planet. The condensed reservoir can form only if the absorbed solar radiation is below the Kombayashi–Ingersoll limit for the gas in question, and even then only if the total mass of volatiles available is sufficient to bring the atmosphere to a state of saturation. As an example of the latter constraint, let's suppose that Mars were in a more distant orbit, where the absorbed

[9] There are other places an outgassed atmosphere can go; water can go into hydration of minerals, and CO_2 can be bound up as carbonate rocks.

solar radiation were only 40 W/m^2. Then, according to Fig. 4.40, the equilibrium surface temperature would be 165 K in saturation, and the corresponding surface pressure would be 5600 Pa. In order to reach this surface pressure for a planet with acceleration of gravity g it is necessary to outgas $5600/g$, or 1509 kg of CO_2 per square meter of the planet's surface. On Earth, outgassed water vapor accumulates in an ocean because the Earth is below the Kombayashi–Ingersoll limit for water vapor. With present solar luminosity Venus (without clouds) is well above the limit, so any outgassed water vapor would accumulate in the atmosphere (apart from the leakage to space). For CO_2, Earth, Mars, and Venus are all above the Kombayashi–Ingersoll limit, so outgassed CO_2 accumulates in the atmosphere (apart from whatever gets bound up in mineral form). Even if you took away the water that allows CO_2 to be bound up as carbonate, Earth would not develop a CO_2 ocean; it would become a hot Venus-like planet instead, with a dense CO_2 atmosphere.

In parting, we must mention two serious limitations of our discussion of the runaway greenhouse phenomenon. First, in computing the Kombayashi–Ingersoll limit, it was assumed that the atmosphere was saturated with the condensable greenhouse gas. However, real atmospheres can be substantially undersaturated, though the dynamics determining the degree of undersaturation is intricate and difficult to capture in simplified models. Undersaturation is likely to raise the threshold solar radiation needed to trigger a runaway state. The second limitation is that the calculations were carried out for clear sky conditions. Clouds exert a cooling influence through their shortwave albedo, and a warming influence through their effect on *OLR*, and the balance is again hard to determine by means of any idealized calculation. Whether clouds have an inhibitory effect on the runaway greenhouse is one of the many remaining Big Questions.

4.7 PURE RADIATIVE EQUILIBRIUM FOR REAL GAS ATMOSPHERES

Pure radiative equilibrium amounts to an all-stratosphere model of an atmosphere, and is a counterpoint to the all-troposphere models we have been discussing. Real atmospheres sit between the two extremes, sometimes quite near one of the idealizations. In this section we will focus on pure infrared radiative equilibrium. The effects of solar absorption in real gases will be taken up in Chapter 5.

From simple analytic solutions, we know essentially all there is to know about pure radiative equilibrium for gray gases. It is important to understand these things because the structure of atmospheres results from an interplay of convection and pure radiative equilibrium. A thorough understanding of pure radiative equilibrium provides the necessary underpinning for determining where the stratosphere starts, and its thermal structure. We will now examine how the key elements of the behavior of pure radiative equilibrium differ for real gases. The specific issues to be addressed are:

- For gray gases in radiative equilibrium, the minimum temperature is the skin temperature based on *OLR*, and is found in the optically thinnest part of the atmosphere in the absence of atmospheric solar absorption. For real gases, can the radiative equilibrium temperature be much lower than the skin temperature?
- For a gray gas atmosphere with a given vertical distribution of absorbers, the radiative equilibrium temperature profile is uniquely determined once the *OLR* is specified. Specifically, one can determine the temperature profile of the radiative-equilibrium stratosphere without needing to know anything about the tropospheric temperature structure. To what extent is this also true for real gases?