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Comparative estimates of climatic consequences of Martian dust storms and of possible nuclear war

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ABSTRACT

We present a simple analytical model which yields estimates of the temperature of the surface T_s and mean atmospheric temperature T_a of a planet. The model considers radiative energy balance at the top of the atmosphere and at the underlying surface, and absorption and scattering within the atmosphere, for globally averaged conditions. The model gives reasonable estimates for (a) T_s and T_a for both clean and dusty Martian atmospheres, (b) and for clean and dusty (asteroid impact) atmospheres of Earth, and (c) the change in T_s for the Earth's atmosphere when the CO_2 content is doubled. When applied to the "war aerosol loaded" terrestrial atmosphere, the model produces results similar to those obtained by Turco, Toon, Ackerman, Pollack and Sagan (TTAPS) and others. However, the analytical nature of the present model exposes certain features which are not revealed by more numerically complicated studies, and this is a matter of particular importance when analysing the consequences of the large uncertainties in smoke injection rates into the atmosphere. Thus the model shows that the sharpest and main changes in T_s occur when the optical depth in the visible τ_s is of order unity or less, reaches the minimum at $\tau_s \approx 3$, and then, as τ_s increases further, T_s slowly recovers to the effective temperature T_e determined by the balance at the top of the atmosphere. This means the decrease in T_s associated with very heavy loading would be followed by a further decrease in T_s when the aerosol begins to settle out. The other important result is the atmospheric temperature T_a over ocean may be considerably higher than T_a over land. Qualitative arguments indicate how aerosol loading might change general circulation patterns and suppress the hydrological cycle.

1. Introduction

The problem of possible effects of nuclear war on the atmosphere and climate is now attracting much attention, following the publication of a paper by Crutzen and Birks (1982). The present authors became interested in the problem at the beginning of 1983 and first reported their estimates of the temperature regime of an aerosol-loaded atmosphere, together with a qualitative discussion of the changes in precipitation and dynamics of the atmosphere, at the All-Union Conference Against Nuclear War Threat for Disarmament and Peace, Moscow, May 1983 (for a condensed version of the report see Golitsyn (1983) and Oboukhov and Golitsyn (1983)).

The first version of the present paper, which described our model and summarized the results given at Figs. 1 and 2, was presented at the Conference on "The World after Nuclear War" October 31–November 1, 1983, Washington, D.C. For various reasons the paper was not sent to this journal until April 1984. In the interim period, general circulation model simulations have been published (Alexandrov and Stenchikov, 1984; Covey et al., 1984). Turco et al. (1983) (hereafter referred to as TTAPS) use the radiative–convective type of climate model, with detailed radiative transfer calculations. The other authors produce three-dimensional time evolution patterns of changes in temperature and other meteorological variables.

Any model is incomplete and may be easily criticized, as can be seen in the already long history of modelling the climate and its changes in general. Models that have been put forward range from "zero-dimensional" models that give only the average surface temperature of the whole globe, to the fully interactive dynamical models of the atmosphere and ocean which give three-dimensional fields of the variables considered. Each has its merits and shortcomings.

We present here an extremely simple analytical model which gives both the average atmospheric temperature, T_a and that of the underlying surface T_s. The model is based on considerations of radiative energy balance at the top of the atmosphere and at the underlying surface. It has already given reasonable estimates for the Martian temperature régime under normal conditions (Ginsburg and Feigelson, 1971) and during global dust storms (Ginsburg, 1973). The model has been generalized to the case of many atmospheric layers (Ginsburg and Safray, 1977) and applied to Venus, for which it produces a realistic vertical temperature profile. However, to investigate the main physical effects, one uses the simplest approximation possible that can be treated analytically, and this yields results which are easy to follow.

We begin with a description of the model (Section 2) and the derivation of simple expressions for the dependence of T_a and T_s on the optical properties of the atmosphere, presented with a discussion of assumptions involved and concomitant limitations of the model. In the same section, we apply the model to the case of doubling the CO_2 content of the atmosphere, which increases its IR opacity. Next, in Section 3, we consider the optical properties of various aerosols (mineral dust, products of forest and urban fires) and estimate absorption and transmission of aerosol-loaded atmospheres. Then (Section 4) we discuss the likely dependence of changes in the thermal régimes of the atmosphere and surface on the optical thickness of the atmosphere, checking our results against what is known about clear terrestrial and Martian atmospheres. Then, using new data for Martian dust storms, we follow the changes in T_a and T_s for the case of a dusty Martian atmosphere and find good agreement with direct observations. For the Earth's atmosphere when filled with dust raised by an event such as an asteroid impact (which may have occurred about 65 × 10^6 years ago), our model reproduces reasonably well the results of more complex model studies by Pollack et al. (1983) who used basically the same approach as TTAPS. Consideration of the fire aerosol case also agrees well with results of TTAPS. Our results presenting $T_a(\tau)$ and $T_a(\tau)$ reveal the effects of uncertainty in estimates of the aerosol concentration and its optical properties etc., which is an advantage over a numerical model, which requires a new set of new computations for each assumed combination of aerosol characteristics. A novel result is that the main changes in T_s and T_a occur when $\tau_s \lesssim 1$, where τ_s is the optical depth for the solar radiation; T_s and T_a are fairly insensitive to τ_s when $\tau_s \gtrsim 2$ or 3. The other result is that the atmosphere with smoke over oceans may be some 20 K warmer than over land because the thermal radiation from the ocean surface would remain virtually unchanged due to great oceanic thermal inertia.

Finally, in Section 5, we discuss the changes in the hydrological cycle and atmospheric dynamics that would accompany the modified thermal régimes of the atmosphere and surface. The arguments presented could facilitate the analysis of such effects in more comprehensive models.

2. The radiative model of the temperature régime of the atmosphere and surface

The intensity of the globally-averaged flux of the solar radiation incident upon an atmosphere is

$$S = 0.25\, I_0(1 - A), \tag{2.1}$$

where I_0 is the "solar constant" which is 1.37 kW m^{-2} at the Earth's mean distance from the Sun (1 a.u.), and A is the mean planetary albedo, equal to 0.3 for the Earth and 0.24 for Mars (Pollack, 1979). From the equality of the mean fluxes of the incoming solar and outgoing thermal radiation, one can define an effective temperature

$$T_e \equiv [I_0(1 - A)/4\sigma]^{1/4}, \tag{2.2}$$

This gives 255 K for the Earth and 211 K for Mars whose mean distance from the Sun is 1.52 a.u.

The thermal balance of a planet as a whole is determined by the radiation balance at the top of the atmosphere: $S(\infty) = F(\infty)$, i.e., by the equality between the net solar radiation flux $S = S\downarrow - S\uparrow$ and the net thermal radiation flux $F\uparrow - F\downarrow$. The arrows correspond to downwelling and upwelling fluxes. The value $S(\infty)$ is determined by (2.1). At the top of the atmosphere

$$F(\infty) = B(T_s)\, D(0, \infty) + \int_0^{\infty} B(z) \frac{dD(z, \infty)}{dz}\, dz, \tag{2.3}$$

where $B(z) = \sigma T^4(z)$, T_s the temperature of the surface, $D(z_1, z_2)$ the integral transmission function for the thermal radiation of the atmospheric layer (z_1, z_2) and $F\downarrow(\infty) = 0$.

At the underlying surface, the effective flux of the solar radiation is $S(0) = 0.25 I_0(1 - A - \alpha) = \alpha_s I$, where $I = 0.25 I_0$, α is the fraction of the insolation absorbed by the atmosphere, α_s is the same at the surface, and $A + \alpha + \alpha_s = 1$. The effective flux of the thermal radiation at the surface is

$$F(0) = F\uparrow(0) - F\downarrow(0) = B_s - \int_\infty^0 B(z) \frac{\mathrm{d}D(0,z)}{\mathrm{d}z} \, \mathrm{d}z, \tag{2.4}$$

where the last term describes the counter-radiation of the atmosphere.

Using the mean value theorem, we take B out of the integral sign in (2.3) and (2.4). Then we obtain effective values $B_a\uparrow$ and $B_a\downarrow$ and corresponding radiative temperatures $T_s\uparrow$ and $T_a\downarrow$ for the outgoing radiation $F\uparrow(\infty)$ and counter-radiation $F\downarrow(\infty)$. Using the values $B_a\uparrow$ and $B_a\downarrow$, we have the conditions of radiation balance at the upper and lower boundaries of the atmosphere in the form

$$B_s D + B_a\uparrow(1 - D) = I(1 - A), \tag{2.5}$$

$$B_s - B_a\downarrow(1 - D) = I(1 - A - \alpha), \tag{2.6}$$

two equations in three variable, B_s, $B_a\uparrow$ and $B_a\downarrow$.

For normal Earth conditions, one can estimate the values of $D(0, \infty)$, $B_a\downarrow$ and $B_a\uparrow$ by ground-based and satellite observations. The averaged temperature of the Earth's surface (northern hemisphere), T_s, is 288 K (Oort and Rasmusson, 1971). $F\uparrow(\infty) = S = 240$ W m^{-2} from the balance requirements. From data of Bolle (1982), the contribution of the radiation from the surface into $F\uparrow(\infty)$ is 23%, i.e. $B_s D = 55$ W m^{-2}. Because $B_s = \sigma T_s^4 = 390$ W m^{-2}, then $D = 0.14$. A very similar value $D = 0.15$ has been obtained by Ginsburg (1982) for the standard model for the mid-latitude atmosphere with $T_s = 288$ K and fractional cloudiness 0.5. For $D = 0.14$ and $T_s = 288$ K from (2.5), we get $T_a\uparrow = 248$ K. From (2.6) at $\alpha = 0.25$ ($\alpha_s = 0.45$) after Bolle (1982), we obtain $T_a\downarrow = 250$ K. At the same time, for $\alpha = 0.2$ ($\alpha_s = 0.5$) after NAS (1975), we have $T_a\downarrow = 250$ K. This is the range of uncertainty of our knowledge of the normal atmosphere. Note that $\frac{1}{2}(T_a\uparrow + T_a\downarrow)$ is close to the mass-weighted temperature of the atmosphere of the northern hemisphere, $T_a = 256$ K (see Oort and Rasmusson, 1971) and to the effective temperature, $T_e = 255$ K.

Because $T_a\uparrow$ is close to $T_a\downarrow$, we assume $B_0\uparrow = B_a\downarrow$, i.e. $T_a\downarrow = T_a\uparrow = T_a$. This would be exact for an isothermal atmosphere, but in our case it is a defect of the simplification we have used (to avoid having to consider a multi-layered atmosphere), which can produce errors of the order of several degrees in the sought values of T_s and T_a. However in an atmosphere with large amounts of smoke and dust, it is expected that the temperature profile would be closer to isothermal than under normal conditions (cf. Turco et al., 1983). An example is the dusty Martian atmosphere, observations of which indicate that approximately isothermal conditions obtain during global dust storms (c.f. Zurek, 1982; Gierasch, 1974). Therefore, the approximation $T_a\uparrow = T_a\downarrow$ should work better under the perturbed conditions than under non-perturbed conditions. In this approximation, eqs. (2.5) and (2.6) give

$$T_s = T_{e0}[(1 - A + \alpha_s)/(1 + D)]^{1/4} \tag{2.7}$$

$$T_a = T_{e0}[(1 - A - \alpha_s D)/(1 - D^2)]^{1/4}, \tag{2.8}$$

where $T_{e0} = (I_0/4\sigma)^{1/4}$ and is equal to 279 K for the Earth if $A = 0$.

Thus

$$T_a = \{0.5[T_s^4 + \alpha T_{e0}^4/(1 - D)]\}^{1/4}$$
$$= [(T_s^4 - \alpha_s T_{e0}^4)/(1 - D)]^{1/4}, \tag{2.9}$$

which is convenient once T_s is known.

Consider now some limiting cases. When all the solar radiation is absorbed in the atmosphere, then $\alpha \to (1 - A)$ and, from Kirchhoff's law, $D \to 0$. Then (2.7) and (2.8) give $T_s = T_e = T_a$, i.e., the temperatures of the atmosphere and the surface tend to the effective temperature T_e. The same result has already been noted in much more detailed numerical studies by Pollack et al. (1983). When the atmosphere is transparent to solar radiation, we have $\alpha \to 0$, $D \to 1$ and $T_s \to T_e$, $T_a \to 2^{-1/4} T_e$, however, as in the simplest estimates of the stratospheric temperature (Goody, 1964).

In intermediate cases, depending on the ratio between absorptions for solar and thermal radiation, "greenhouse" ($T_s > T_e$) and "anti-greenhouse" ($T_s < T_e$) effects are possible. The first effect takes place when the atmosphere absorbs more thermal radiation from the surface than solar radiation; the second is in the reverse case.

A simple test of this radiation model is an estimate of the T_s sensitivity to doubling of the CO_2 in the Earth's atmosphere, $\delta = \Delta T_s(2 \times CO_2)$. As is well-known (NAS, 1982) the change in the flux of the outgoing thermal radiation at the top of the troposphere is $\Delta F\uparrow = -4$ W m^{-2}. In our model,

$$F\uparrow = B_s D + B_a(1 - D). \tag{2.10}$$

Assuming B_s and B_a to be the same, we relate ΔF and ΔD from here. For $\Delta F = -4$ W m^{-2} and the actual values $T_s = 288$ K and $T_a = T_a\uparrow = 248$ K, we get $\Delta D = 0.023$ and then, from (2.7) we obtain $\delta = 1.4$ K. The change is caused only by an increase of the CO_2 concentration without taking into account various feedback mechanisms—in particular, the main one, the increase of atmospheric absolute humidity with the temperature increase. We recall that the characteristic values of δ for all models, radiative-convective ones and GCMs for fixed absolute humidity, are in the range 1 to 1.5 K and for constant relative humidity 2 to 3 K (see e.g. Manabe and Wetherald, 1967, 1980).

Another limitation of our model for the normal Earth is that for $D = 0.14$ and $\alpha = 0.2$ (see above discussion after, e.g. (2.6)) our eqs. (2.7) and (2.8) produce $T_s = 282$ K and $T_a = 250$ K, which are some 6 K below actual values. To obtain $T_s = 288$ K, one must decrease significantly either D or α, or both, which is not in good agreement with observations (see NAS, 1975; Bolle, 1982). But, as we have mentioned already, the model should work better for perturbed conditions when vertical temperature profiles are closer to isothermal.

For Mars, at an average distance from the Sun of 1.52 a.u. $I = 148$ W m^{-2} and $T_e = 211$ K if $A = 0.24$ (see Pollack, 1979). Eqs. (2.8) to (2.10) produce for $\alpha \simeq 0$ (normal conditions) $T_s = 218$ K, $T_a = 183$ K and $D = 0.76$. The simplest estimate of the effective IR-transmission for the Martian atmosphere assumes that the atmosphere is completely opaque in the CO_2 absorption band, 13–18 μm, and is transparent outside the band. Then D is equal to that part of the black body thermal radiation outside the band. For $T_s = 218$ K, we obtain $D = 0.77$, which is practically the same as $D = 0.76$ above. Therefore our model for normal terrestrial and for Martian conditions is not too far from reality, and there are reasons to expect that with the increase of the density of absorbing aerosols, it will perform better.

3. Optical properties of the aerosol atmosphere

Now we shall discuss in a crude way the relationships of the optical depths in the thermal IR, τ, and in the solar or visible spectral ranges, τ_s, with the values of D and α entering (2.7) and (2.8). The vertical optical depth $\tau = k_t m_t$, where k_t is the effective absorption coefficient and m_t the total mass of all IR-absorbing atmospheric constituents. Then

$$D = \exp(-r\tau) = \exp(-rk_t m_t), \tag{3.1}$$

where $r = 1.66$ is the dimensionless radiation diffusivity coefficient (Goody, 1964).

For solar radiation, $\tau_s = \tau_\sigma + \tau_\alpha = k_\sigma m_\sigma + k_\alpha m_\alpha$, where k_σ and k_α are the scattering and absorption coefficients, and m_σ and m_α the mass column densities of scattering and absorbing atmospheric constituents, respectively. The transmission of the direct solar radiation is

$$P_d = \exp(-\tau_s/\mu), \tag{3.2}$$

where $\mu = \cos\theta$ and θ is the solar zenith angle. For the transmission of the total, direct + scattered, solar radiation, P_Σ, simple analytical expressions are justified only for special conditions. We express P_Σ similarly to (3.1):

$$P_\Sigma = \exp(-r_s \tau_s). \tag{3.3}$$

The coefficient r_s depends on the ratio of absorbing and scattering properties of the atmosphere and surface, such as the albedo of the underlying surface A_s, the single scattering albedo $\tilde{\omega} = k_\sigma(k_\alpha + k_\alpha)^{-1}$, the scattering phase function, and the solar zenith angle arc cos μ. If the atmosphere absorbs the solar radiations, the value of P_Σ is determined mainly by the absorption optical depth, $\tau_\alpha = (1 - \tilde{\omega})\tau_s$. Scattering lengthens the effective photon path within the atmosphere. The soot particles and products of oil and gas combustion have $\tilde{\omega} = 0.5$ (see Tarasova and Feigelson, 1981; Crutzen and Birks, 1982) and products from wildfires have $\tilde{\omega} = 0.9$ (Crutzen et al., 1984). For such types of absorbing aerosols, using computation by Tarasova and Feigelson, one can estimate values of r_s. For $\tilde{\omega} = 0.9$, $\mu = 0.5$, one gets $r = (1.5$ to $2)(1 - \tilde{\omega})/\mu \approx 0.3$ to 0.4, and for $\mu = 0.2$ we have $r_s = (2$ to $3)(1 - \tilde{\omega})/\mu = 1$ to 1.5. Therefore, with the increase of θ due to scattering, the absorption increases from a factor 1.5 to 2 at

$\mu = \cos\theta = 0.5$, to 2 to 3 at $\mu = 0.2$. For $\tilde{\omega} = 0.5$ at $\mu = 0.5$ and 0.2, the coefficients $r_s = (1$ to $1.2)(1 - \tilde{\omega})/\mu$, and, practically, $P_\Sigma = \exp(-\tau_0/\mu)$ according to results of the same computations by Tarasova and Feigelson, i.e., with the increase of the absorption, the scattering is less significant. As a result, for a very strongly absorbing aerosol, $\tilde{\omega} = 0.5$, one may assume $P_\Sigma = \exp[-(1 - \tilde{\omega})\tau_s/\mu]$, and for an aerosol with $\tilde{\omega} \simeq 0.9$, we use $P_\Sigma = \exp[-2(1 - \tilde{\omega})\tau_s/\mu]$.

The fraction of the solar radiation absorbed by the surface is equal to $\alpha_s = P_\Sigma(1 - A_s)$. The albedo of the underlying surface, A_s, for the Earth and Mars is in the range 0.05–0.2, except for relatively small polar caps. While calculating $\alpha_s(\tau)$, we neglect any possible changes in A_s.

For two types of aerosol which we shall consider later, the mineral dust and smoke, we assume different ratios between optical depths τ_s (solar) and τ (thermal IR). According to estimates by Ginsburg (1973), during a mature Martian dust storm the IR transmission is $D = 0.5$, and the transmission for the solar radiation $P_\Sigma = 0.15$. For these conditions, one may estimate the IR optical depth τ to be about 0.25 of the depth τ_s for solar radiation. More recent detailed studies (e.g. Pollack et al. 1979) give for different stages of the dust storm evolution estimates of τ_s/τ. The values range from 1.2 to 5. On the basis of data summarized by Zurek (1982), one may assume $\tau_s/\tau \approx 3$ to 4. For later use we adopt $\tau_s = 4\tau$ when we consider the dust storms and results of an asteroid impact with Earth. Following Chylek et al. (1981) we shall use $\tau_s = 10\tau$ for combustion products as did Crutzen and Birks (1982).

In a more accurate approach, one should take into account the dependence of the system albedo and atmospheric transmission on the aerosol optical properties and the amount by solving the full radiation transfer equation. The increase of the aerosol amount can change planetary albedo considerably. For a large optical depth, the albedo A tends to the albedo of an aerosol layer with $\tau_s \gg 1$. Filling up the atmosphere with mineral dust or with sulphuric volcanic aerosol leads to an increase of the albedo. On the other hand, a strongly absorbing aerosol, like smoke from forest and urban fires, decreases the albedo.

Various approximate methods of the radiation transfer theory can be used to estimate the resulting effect. For example, Crutzen et al. (1984), in their calculation of the albedo and transmission of the smoke-filled atmosphere used expressions obtained by Sagan and Pollack (1967) and applied them to Venusian clouds. Without going into details here, we note that these expressions overestimate the albedo for large masses of absorbing aerosol. Using more accurate numerical and analytical calculations by Sagan and Pollack (1967), we computed asymptotic values of A when $\tau \to \infty$. For $\tilde{\omega}_3 = 0.9$, 0.7 and 0.5, we obtained $A = 0.2$, 0.1 and 0.05, respectively. Details of the computational methods can be found in Feigelson (1984). Intermediate cases may be approximated by the following expression

$$A(\tau) = A(\infty) + [A(\infty) - A(0)] \exp(-r_s \tau_s); \tag{3.4}$$

for a smoke mixture of urban and forest fires ($\tilde{\omega} = 0.7$), eq. (3.4) gives $A(\tau) = 0.1 + 0.2 \exp(-r_s\tau_s)$.

4. Changes in the thermal régime

Now the changes in the thermal régime for various cases with increases in aerosol amount can be estimated through eqs. (2.7)–(2.9). The radiation characteristics α_s and D for the normal atmosphere are multiplied correspondingly by factors $\exp(-r_s\,\Delta\tau_s)$ and $\exp(-r\Delta\tau)$, where $\Delta\tau_s$ and $\Delta\tau$ are changes of the optical depths for solar and thermal radiation and quantities r_s and $r = 1.66$ have been discussed in Section 3.

For Mars, and also for the Earth's land area, we suppose that the atmosphere and underlying surface are in a temperature equilibrium determined by their optical properties. This is justified because of the small heat capacity of soils. In contrast, the temperature of the ocean will change little, due to its large heat capacity. A crude estimate may be obtained as follows. The cooling will affect only the upper mixed layer (as in the winter convection of the upper ocean), about 70 m deep as averaged over the world ocean (Manabe and Wetherald, 1980). The total heat capacity of such a layer is about 3×10^8 J m^{-2} K^{-1}. An averaged heat flux from the ocean to atmosphere is of the order 100 W m^{-2} (or less). Then, for a timespan of half a year, it will give up 1.5×10^9 J m^{-2}, which produces a 5 K cooling. The maximum cooling in Pollack et al. (1983) and Turco et al. (1983) is about 3 K. Note that the characteristic time scale for the Earth's atmosphere to react to perturbations in its radiative régime is of order a week (Golitsyn, 1964; Goody, 1964; Ginsburg, 1980; Gryanik, 1982), while for the surface it is much shorter. Therefore the land surface temperature T_s would adjust according to (2.5)–(2.6) to the atmospheric temperature T_a. The latter over the ocean would tend to adjust to the ocean surface temperature according to (2.9). This justifies somewhat our quasi-stationary consideration (at least for mid-continental and mid-oceanic regions), which disregards transient phenomena in time and space.

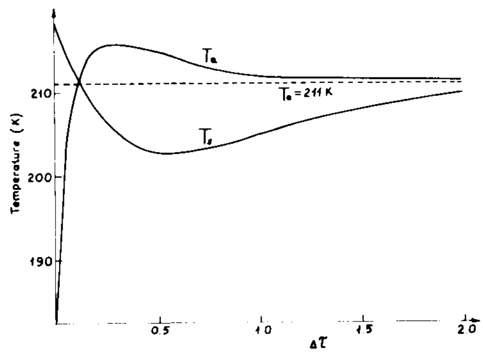

Fig. 1. The dependence of Martian surface, T_s, and atmospheric, T_a, temperatures on the IR optical depth $\tau = 0.25\tau_s$, τ_s being the optical depth for solar radiation. T_e is the effective temperature.

Fig. 1 presents results of calculations of T_s and T_a as functions of the IR optical depth, τ, for $\tau_s = 4\tau$. For a mature dust storm $\tau_s \simeq 2$ to 4, i.e. $\tau = 0.5$ to 1. Then $\Delta T_s = -(10$ to $15)$ K, which agrees quite well with Viking lander measurements (Ryan and Henry, 1979). At the same time, $\Delta T_a = +(27$ to $33)$ K, which again is of the order of changes in the Martian dusty atmosphere observed by the radio-occultation method (Moroz, 1978). The same observations have shown that the vertical temperature profiles are in fact nearly isothermal. One can also see a distinct anti-greenhouse effect: $T_s < T_e < T_a$ for $\tau_s > 0.8$ ($\tau > 0.2$).

Fig. 2 gives the results for the Earth. Curves 1 and 2 show changes in T_s and T_a over land for the case of mineral dust ($\tau_s = 4\tau$ and $A = 0.3 =$ const.). One sees that for $\tau > 0.6$, ($\tau_s > 2.4$), a weak anti-greenhouse effect ($T_s < T_e$) is present, but the maximum cooling at $\tau \approx 1$ does not exceed 1 K relative to T_e. For further increase of τ, both T_s and T_a change little and tend to their asymptotic value $T_e = 255$ K. The cooling of the surface is somewhat less than in the much more detailed model by Pollack et al. (1983), presumably because we have neglected any increase in albedo due to further scattering of solar radiation by the dust, but it is still sufficient to have caused an ecological catastrophe 65 million years ago.

The other curves on Fig. 2 are for fires when $\tau_s = 10\ \tau$. Curves 4 and 5 show the dependence of

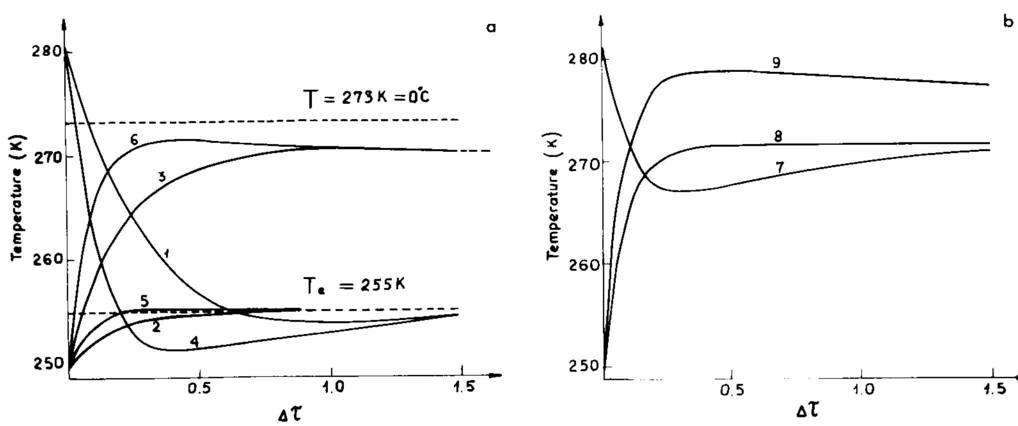

Fig. 2. The dependence of the surface and atmospheric temperature of the Earth in various cases: (a) 1, 2: T_s, T_a for the surface and atmosphere above land for mineral dust; 3 for T_a over ocean ($A = 0.3$); 4, 5 for T_s, T_a for aerosols of fire origin, $A = 0.3$; 6, the same over oceans; (b) 7, 8, the same for $A(\tau) \rightarrow 0.1$ over land; 9, T_a the same over ocean.

T_s and T_a on τ for $A = 0.3$, and curves 7 and 8 the same for $A(\tau) \to 0.1$, as in (3.4). In all cases, the maximum changes are near $\tau_s \approx 3$ or $\tau \approx 0.3$. We want to stress especially that the sharpest changes are for $\tau_s \lesssim 1$, including the fall of T_s below the freezing point of water. Curves 7 and 8 tend to the equilibrium temperature 271.5 K as $A \to 0.1$ according to (3.4), where we neglect any increase in albedo due to enhanced scattering of light by the dust.

Curves 3, 6 and 9 are related to the atmosphere over the ocean with $T_s = 282$ K, correspondingly, for mineral dust and combustion products, with $A = 0.3$ and $A \to 0.1$. In the limit of large τ, τ_s when $\alpha \to 1 - A$, $D \to 0$ eq. (2.10) gives

$$\lim_{\tau_s \to \infty} T_a = [0.5(T_s^4 + T_e^4)]^{1/4}. \tag{4.1}$$

It is important to note that the temperature difference between the atmosphere over land and over ocean may reach some 20 K. Comparison of the curves 1 and 4 for T_s and 2 and 5 for T_a in cases of dust and smoke illustrates the influence of the thermal IR opacity: the greater its value, the slower the rate at which curves approach their asymptotes.

5. Possible dynamic and other changes in the atmosphere

Significant changes of the thermal régimes in the atmosphere and of the surface would be accompanied by changes in the atmospheric dynamics. Consider first a homogeneous Mars, which lacks oceans. The heating of the atmosphere together with the cooling of the surface increases the atmospheric static stability. This leads to damping of the baroclinic instability of the atmosphere or, at least to slowing down of the development of baroclinic perturbations—cyclones (see e.g. Holton, 1972). Viking lander observations on the surface of Mars (Ryan and Henry, 1979) show that in a clear winter atmosphere, the passage of cyclones is quite regular, but when a dust storm arrives the cyclones cease. That means that one Hadley cell of the general circulation encompasses the whole winter hemisphere. Similar effects were noted by Alexandrov and Stenchikov (1984) and Covey et al. (1984) in their GCM modelling of the atmosphere loaded by smoke aerosol; without specifically mentioning cyclones and presenting weather maps (apparently due to the lack of space), both groups of authors describe the setting of a single Hadley cell.

Since the Washington conference Boubnov and Golitsyn (1985) carried out laboratory experiments in a differentially heated rotating annulus in the presence of an imposed vertical temperature gradient δT. A qualitative theory of regimes of such a flow was also developed as a generalization of the theory by Lorenz (1962), describing regimes without imposed vertical temperature gradient. Hide (1977) discusses the relationship of such laboratory experiments to atmospheric general circulation through the theoretical developments that have been developed, such as that by Lorenz (1962). In short, our theoretical analysis and experiments demonstrate that with the increase of δT, the non-axisymmetric régime changes in such a way that we may observe a decrease in the number of vortices or their complete elimination. Inspection of the changes of the non-dimensional parameters governing the flow régimes shows that this would be the case for the dusty Martian atmosphere, and thus may explain the formation of a single Hadley cell in the numerical experiments cited above.

In practice of course, there would be complications caused by the inhomogeneity of atmospheric and surface temperatures over land and ocean. In this connection, we note that curves 6 and 5 on Fig. 2 show that the temperature T_a over ocean is some 20 K warmer than over land. The warmer oceans and cooler continents would drive a monsoonal circulation of the winter type, thus introducing a dry season of the kind commonly found in tropical countries. Precipitation would be expected mainly over oceans, not over continents.

Because the aerosol-filled atmosphere over oceans would also be warmed, and even more so than over continents because of the virtually unchanged ocean thermal radiation, the atmospheric static stability over the oceans should also increase, and as a consequence, convection over oceans would be shallower than is the case today. On the other hand, convective instability is expected to develop near the upper boundary of the light-absorbing clouds which would be heated, and this would carry the smoke still higher in the atmosphere. Very crude estimates (Ginsburg et al., 1984) show that this mechanism may carry the smoke up to the tropopause (11 km) in a matter of about 2 weeks. The net effect would be a decrease in aerosol scavenging by precipitation.

In general, an increase of the static stability over both land and ocean will slow down the heat and humidity exchange between underlying surface and atmosphere, thus decreasing the rate of replenishment of atmospheric water vapour. The mere increase of the atmospheric temperature will, other things being equal, decrease the relative humidity of the atmosphere, making condensation more difficult. In the extreme case when the temperature rises right from the surface, condensation will be impossible (since the relative humidity cannot exceed 100%). It is known (Rodgers, 1976) that in normal conditions, the marine atmosphere is less polluted with aerosols than the atmosphere over land, so the droplets in oceanic rains are larger than in continental rains. In the smoke-filled atmosphere, one may expect smaller droplets everywhere, which would also hamper the scavenging. In general, there are several reasons to expect that in an atmosphere heavily loaded with absorbing aerosol, the whole hydrological cycle may be strongly attenuated.

This most qualitative discussion indicates several potential positive feedback mechanisms in an aerosol-loaded atmosphere. Of course, much detailed work is needed on many microphysical and mesoscale processes, in order to elucidate quantitatively many microphysical and mesoscale processes, such as patchiness of smoke clouds and circulation over oceanic coastal regions.

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