A Possible Factor Regulating the Amount of the Atmosphere on the Earth and Mars

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The attempt of the present note is to give some consideration on a possible mechanism which can regulate the total amount of the atmosphere on a planet that is transparent for visible and infrared radiation. It has been suspected that the major component gas of the present atmosphere of the earth has been supplied from the earth's interior for a long geologic time (Rubey 1951). It has been also suspected that molecules of the atmospheric gas can escape from the top of the atmosphere (exosphere) into the planetary space by either of thermal or non-thermal evaporation if appropriate conditions are satisfied (MacDonald 1963). Were this rate of escape of the gas independent on the rate of supply, the total amount of the atmosphere would increase or decrease monotonously. On the contrary, were the rate of escape to increase with the total amount of the atmosphere, the amount would have a tendency to approach an equilibrium value at which the rate of escape counterbalances the rate of supply.

The role of infrared radiators of the present atmosphere such as H_2O , O_3 , CO_2 , OH and O in the heat budget of the atmosphere is important, but their concentrations are very scarce as compared with the concentrations of nitrogen and oxygen molecules which are transparent for infrared radiation. In order to eliminate the role of infrared radiator and to emphasize the role of nitrogen atmosphere, a hypothetical atmosphere is assumed here. The atmosphere which is transparent for both of visible and infrared radiations and very opaque for ultraviolet radiation is assumed to lie on the ground. This may be one of the simplest model of our nitrogen atmosphere.

Under the assumption, the ultraviolet radia-

tion I_{UV} from the sun is completely absorbed in the layer (thickness 0 mb to P_* mb) at the top of the atmosphere, while the visible light I_V from the sun reaches the ground without being absorbed. The infrared radiation I_{IR} from the ground goes out to space without being absorbed at all. Accordingly, there is no heat sink in the atmosphere. The layer at the top of the atmosphere is only one heat source in the atmosphere, where a part of energy of the absorbed ultraviolet radiation becomes heat. The efficiency of the conversion of ultraviolet radiation into heat is ϵ . This heat should transfer downward to the ground by heat conduction or eddy diffusion in a steady state. Equations of heat balance under these circumstances are as follows;

$$I_{IR} = I_V + F \quad (1)$$

$$I_{IR} = \sigma T_s^4 \quad (2)$$

$$F = K \frac{\partial T}{\partial z} \quad (3)$$

$$F = \epsilon I_{UV} \quad (4)$$

where F is the downward heat flux from the top of atmosphere to the ground, σ is Stefan-Boltzmann constant, T_s is the temperature of the ground, K is the coefficient of heat conduction (or eddy diffusion), T is the temperature of the air and z is the height. In this atmosphere, the temperature of the air increases monotonously from the bottom to the top. The coefficient K would be smaller when the temperature gradient is positively steeper and the atmosphere is in consequence more stable statically so far as K represents the effect of eddy transport. On the other hand, it would be larger when the temperature of the air is higher so far as K represents the effect

of molecular heat conduction. For the sake of simplicity, K is assumed to be constant throughout the atmosphere irrespective of temperatures of the air. In this case, the vertical temperature gradient (lapse rate) should be constant, therefore the atmosphere is polytropic. From the pressure-temperature relation of the polytropic atmosphere, one obtains

$$\frac{T_*}{T_s} = \left(\frac{P_s}{P_*} \right)^\alpha, \quad \alpha = \frac{R}{Mg} \frac{\epsilon I_{UV}}{K}, \quad (5)$$

where P_s is the pressure at the ground representing the amount of the atmosphere, T_* is the temperature of the top of the atmosphere (more exactly temperature at the level of P_*), R is the universal gas constant 8.32×10^7 erg deg $^{-1}$, M is the molecular weight of the atmospheric composition, and g is the gravitational acceleration.

At the condition of the present earth, numerical values which are adopted in eq. (5) are follows; $M=29$, $g=980$ gal, $T_s=300^\circ K$,

and $I_{UV}=2.5$ erg $\cdot$ cm $^{-2}$ $\cdot$ sec $^{-1}$ ($\lambda < 1400 \text{ \AA}$) on the average of all latitudes and longitudes (Nawrocki and Papa 1961). ϵ is 0.4 (Harris and Priester 1962). The order of the magnitude of P_* may be 10^{-7} mb if it represents the bottom of the exosphere above the level of which the temperature is constant with height, but it may be 10^{-5} mb if it is estimated from the fact that the absorption coefficient of ultraviolet radiation ($\lambda < 1400 \text{ \AA}$) of the atmosphere is of the order of 10^2 cm $^{-1}$ (Watanabe 1958, Dalgarno and Parkinson 1960, Goody 1964). In the present calculation, P_* is assumed to be 10^{-6} mb as an intermediate. K is obtained tentatively as $\epsilon I_{UV}/(\partial T/\partial z)$ using the value of the temperature gradient between 100 km and 500 km level of the actual terrestrial atmosphere, because this part of the terrestrial atmosphere is considered to resemble best to the model atmosphere under consideration. As $\partial T/\partial z = 5 \times 10^{-5}$ deg $\cdot$ cm $^{-1}$ (MacDonald 1963), K is obtained as 2×10^4 erg cm $^{-1}$ sec $^{-1}$ deg $^{-1}$. Then, the exponent α in (5) is 0.15 for the condition of the earth.

Table 1. Rate of escape as a function of the amount of the atmosphere under conditions of the earth

Amount of atmosphere P_s (mb)	Temperature of the top of the atmosphere T_* ($^\circ K$)	Rate of thermal escape of N -atom E (atmos $\cdot$ cm $^{-2}$ $\cdot$ sec $^{-1}$)	Time required for loss of N_2 atmosphere of 1000 mb (year)
1	2382	3.26×10^{-4}	4.32×10^{21}
10	3366	41.8	3.36×10^{16}
100	4755	1.68×10^5	8.38×10^{12}
1000	6717	5.64×10^7	2.50×10^{10}
5510	8675	1.41×10^9	1.00×10^9
10000	9486	3.36×10^9	4.19×10^8

Numerical values calculated from (5) are tabulated in the first and the second columns of Table 1. These columns show the temperature of the top of the atmosphere as a function of the amount of the atmosphere. The amount of the atmosphere is expressed in pressure at the bottom. In the third column of Table 1 the rate of thermal escape of nitrogen atoms at the temperature T_* is tabulated. The rate of the escape E is calculated from the formula (MacDonald 1963),

$$E = n_* \left(\frac{g_*}{2\pi} \right)^{\frac{1}{2}} r \left(\frac{RT_*}{Mg_*} \right)^{-\frac{1}{2}} \exp \left(-\frac{rMg_*}{RT_*} \right), \quad (6)$$

where n_* is the number density of particles at the escape level 10^8 particles $\cdot$ cm $^{-3}$, $g_*=843$ gal at this level (500 km above sea level), r is the distance from the center of the earth to the escape level 6.87×10^8 cm (6370+500 km) and M is assumed to be the atomic weight of nitrogen 14. As the rate of escape of nitrogen atoms is considered to be very much larger than that of nitrogen molecules, this assumption is adopted. This means that nitrogen molecules are dissociated into atoms at the top of the atmosphere. From the table, it can be seen that the rate of thermal escape of nitrogen is practically negligible at the

amount of the atmosphere below the surface pressure 1000 mb, while at the surface pressure 10000 mb, the rate of escape attains 3.36×10^9 atoms $\cdot$ cm $^{-2}$ $\cdot$ sec $^{-1}$. At this rate of escape, nitrogen atmosphere of 1000 mb will be lost for the period 4.19×10^8 years. Though it is very difficult to estimate the rate of supply of nitrogen gas into the atmosphere, if it is assumed to be 1000 mb per 10^9 years, the rate of escape at $T_*=8675^\circ K$ can counterbalance it. This temperature corresponds to the amount of the atmosphere $P_s=5500$ mb. So, the surface pressure 5.5 atm may be the equilibrium amount toward which the amount of the nitrogen atmosphere would approach. If the effect of infrared radiators in the atmosphere were taken into consideration, the equilibrium value would be larger than 5.5 atm. On the contrary, if the effect of non-thermal escape under conditions of coupling with geomagnetic field were taken into consideration and if it were as much as the rate of thermal escape, the equilibrium value would be smaller than 5.5 atm. It has been known that hydrogen and helium gases escape most easily to space among all gases, but their effect on change of the total pressure of the atmosphere is considered to be very small.

Next, the same procedure adopted above is applied to the model atmosphere under conditions of Mars. Numerical values used are $g=0.38 \times 980=372$ gal, $I_{UV}=0.431 \times 2.5=1.08$ erg $\cdot$ cm $^{-2}$ $\cdot$ sec $^{-1}$ (the factor 0.431 is determined from the inverse square of the ratio of distances from the planets to the sun), $\epsilon=0.4$, $K=2 \times 10^4$ erg $\cdot$ cm $^{-1}$ $\cdot$ sec $^{-2}$ $\cdot$ deg $^{-1}$, $r=3.392 \times 10^8$ cm (equatorial radius), $T_s=200^\circ K$ and $P_*=0.38 \times 10^{-6}$ mb. Results are tabulated in Table 2. It can be seen that the rate of thermal escape is 1.41×10^9 atom $\cdot$ cm $^{-2}$ $\cdot$ sec $^{-1}$ at the surface pressure $P_s=19.7$ mb. This rate corresponds to loss of nitrogen atmosphere of 1000 mb (at $g=980$ gal) for the period of 10^9 years. If the effect of infrared radiators in the atmosphere were taken into consideration, the equilibrium value would be larger than 19.7 mb. The surface pressure of Martian atmosphere was estimated to be 90 mb by Dollfus (1961), 25 ± 15 mb by Kaplan *et al* (1964), and 4.1 to 7.0 mb by Kliore *et al* (1965) from the analysis of data obtained by the space probe Mariner IV. It should be noted that the equilibrium amounts of the model atmosphere under conditions of the earth and Mars have the same order of the magnitude with the amount of the actual atmospheres of these planets.

Table 2. Rate of escape as a function of the amount of atmosphere under conditions of Mars.

Amount of atmosphere P_s (mb)	Temperature of the top of the atmosphere T_* ($^\circ K$)	Rate of thermal escape of N -atom E (atoms $\cdot$ cm $^{-2}$ $\cdot$ sec $^{-1}$)	Time required for loss of N_2 atmosphere of 1000 mb (year)
1	2468	1.97×10^6	7.14×10^{11}
10	3652	4.28×10^8	3.30×10^9
19.7	4100	1.41×10^9	1.00×10^9
100	5400	1.51×10^{10}	9.33×10^7
1000	7988	1.59×10^{11}	8.86×10^6

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地球及び火星の大気の量を制御する可能な因子について

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