EVOLUTION OF SEAFLOOR SPREADING RATE BASED ON ^{40}Ar DEGASSING HISTORY

Eiichi Tajika[1] and Takafumi Matsui

Department of Earth and Planetary Physics, University of Tokyo

Abstract A new degassing model of ^{40}Ar coupled with thermal evolution of the mantle is constructed to constrain the temporal variation of seafloor spreading rate. In this model, we take into account the effects of elemental partition and solubility during melt generation and bubble formation, and changes in both seafloor spreading rate and melt generation depth in the mantle. It is suggested that the seafloor spreading rate would have been almost the same as that of today over the history of the Earth in order to explain the present amount of ^{40}Ar in the atmosphere. This result may also imply the mild degassing history of volatiles from the mantle.

Introduction

Temporal variation of seafloor spreading rate over the history of the Earth has profound influences on the global geochemical cycle through volatile degassing from and regassing into the mantle. In other words, it affects the evolutions of the atmosphere and oceans, and thus the terrestrial environment [e.g., McGovern and Schubert, 1989; Tajika and Matsui, 1992, 1993]. It is, however, difficult to reconstruct such a temporal variation of seafloor spreading rate from any geological evidence. This is because the ancient seafloor itself has subducted into the mantle. Because the seafloor spreading rate plays a dominant role in transfer of internal heat [e.g., Turcotte and Schubert, 1982], we may obtain the evolution of seafloor spreading rate by studying the global geochemical cycle coupled with the thermal evolution of the mantle.

In this paper, we try to infer the temporal variation of seafloor spreading rate, using a new degassing model of ^{40}Ar coupled with thermal history of the mantle in which temporal variation of degassing rate is assumed to depend on changes in seafloor spreading rate and melt generation depth in the mantle.

Models

We use ^{40}Ar as a tracer for the mantle degassing. ^{40}Ar is considered to be the best tracer for the degassing history of the mantle among various volatile components. This is because (1) ^{40}Ar is a decay product of potassium, and so very little amount of ^{40}Ar existed at the time of formation of the Earth. This means that ^{40}Ar may not have experienced impact degassing which is considered to have occurred during accretion of the Earth [e.g., Matsui and Abe, 1986]. (2) The existence of ^{40}Ar in the present atmosphere suggests the occurrence of subsequent continuous degassing over the history of the Earth [e.g., Turekian, 1964; Hamano and Ozima, 1978]. And (3) ^{40}Ar probably has not experienced any regassing into the mantle, which means that we can only consider the degassing process for the case of ^{40}Ar. In these respects, ^{40}Ar is considered to be an adequate indicator to constrain the degassing history of the Earth after its accretion.

^{40}Ar is produced by potassium decay as ^{40}K+e^{-} →^{40}Ar, where the decay constant is λ_e=5.85×10^{-11} yr^{-1}. ^{40}Ar produced in the mantle degasses to the atmosphere mainly through the mid-ocean ridge volcanism accompanied by the formation of oceanic crusts. Since potassium is an incompatible element,

mantle potassium concentrates into melt being associated with melt generation in the mantle below the mid-ocean ridge, and eventually solidifies as the oceanic crusts. Thus, potassium concentrates into the oceanic crust from the mantle concurrently with ^{40}Ar degassing. Potassium in the oceanic crust also decay to ^{40}Ar which would degas to the atmosphere probably by diffusion through the oceanic crust or by arc volcanisms at the subduction zone. Potassium remaining in the oceanic crust would finally return to the mantle.

The continental crust is considered to have formed by remelting of the oceanic crust with water at the subduction zone [e.g., Campbell and Taylor, 1983]. If this were the case, the parts of potassium in the subducting oceanic crusts might be transported to the continental crust at the time of its formation. ^{40}Ar produced by potassium in the continental crust has degassed to the atmosphere with decay constant $K^{Ar}_{D,cc}$ = 3.71×10^{-10} yr^{-1} which is estimated from a relation between K–Ar mineral ages and Rb–Sr whole rock ages for the same crustal rock [Hamano and Ozima, 1978].

We can calculate transportation of K and Ar between various reservoirs using a simple box model composed of three potassium reservoirs (mantle, oceanic crust, continental crust) and of four argon reservoirs (mantle, oceanic crust, continental crust, atmosphere). The mass balance equations for argon and potassium are as follows:

$$\frac{d}{dt}(^{40}K)_{mant} = -\lambda_T(^{40}K)_{mant}-K^K_D(^{40}K)_{mant}+(1-A_K)tr^{-1}(^{40}K)_{oc} \tag{1}$$

$$\frac{d}{dt}(^{40}K)_{oc} = -\lambda_T(^{40}K)_{oc} + K^K_D(^{40}K)_{mant} - tr^{-1}(^{40}K)_{oc} \tag{2}$$

$$\frac{d}{dt}(^{40}K)_{cc} = -\lambda_T(^{40}K)_{cc} + A_K tr^{-1}(^{40}K)_{oc} \tag{3}$$

$$\frac{d}{dt}(^{40}Ar)_{mant} = \lambda_e(^{40}K)_{mant} - K^{Ar}_D(^{40}Ar)_{mant} \tag{4}$$

$$\frac{d}{dt}(^{40}Ar)_{oc} = \lambda_e(^{40}K)_{oc} - tr^{-1}(^{40}Ar)_{oc} \tag{5}$$

$$\frac{d}{dt}(^{40}Ar)_{cc} = \lambda_e(^{40}K)_{cc} - K^{Ar}_{D,cc}(^{40}Ar)_{cc} \tag{6}$$

$$\frac{d}{dt}(^{40}Ar)_{atm} = K^{Ar}_D(^{40}Ar)_{mant}+tr^{-1}(^{40}Ar)_{oc}+K^{Ar}_{D,cc}(^{40}Ar)_{cc} \tag{7}$$

where tr is the residence time of seafloor (=area of seafloor/ seafloor spreading rate), K^i_D ia the coefficient of degassing rate, and λ_T is the total ^{40}K decay constant (= 5.305×10^{-10}yr^{-1}). Subscripts represent as follows: mant = mantle, oc = oceanic crust, cc = continental crust, and atm = atmosphere, respectively. A_K is the accretion ratio of potassium, which is a free parameter in this model. The fraction A_K of potassium in the subducting oceanic crust is assumed to be transported to the continents with continental growth. A_K is iteratively determined so that the amount of potassium in the continental crust at present may be equal to the observed amount. We considered various models of continental growth. However, the continental growth model did not affect the numerical results seriously (Table 2), and then we use the model with a constant growth rate as a standard.

The degassing process of Ar is modeled as follows: ^{40}Ar in a certain volume V_D of the mantle degas to the atmosphere when oceanic plates are formed at the mid-ocean ridge. This volume V_D probably depends on the seafloor spreading rate Sr and the melt generation depth d_m in the mantle. Assuming a first order rate process (i.e., assuming a homogeneous mantle), the degassing rate F_D of ^{40}Ar from the mantle is expressed by

[1]Now at Geological Institute, University of Tokyo

$$F_D = K_D^{Ar} \cdot ({}^{40}\text{Ar})_{mant} \qquad (8)$$

where K_D^{Ar} is the coefficient of degassing rate of argon.

In the previous studies of degassing history [e.g., Turekian, 1964; Hamano and Ozima, 1978], the coefficient of degassing rate was given as an adjustable parameter for the degassing history and called "degassing constant". However, in this study, we define the coefficient of degassing rate as follows [e.g., Tajika and Matsui, 1992]:

$$K_D^{Ar}(t) \equiv f_{Ar} \cdot V_D(t)/V_M = f_{Ar} \cdot Sr(t) \cdot d_m(t)/V_M \qquad (9)$$

where V_M is the volume of the mantle, $V_D(t)$ is the volume of mantle material from which degassing occurs in an unit time, $Sr(t)$ is the seafloor spreading rate, and $d_m(t)$ is the melt generation depth in the mantle. Here we assume the degassing volume $V_D = Sr \times d_m$, which is time-dependent in this case. f_{Ar} is the degassing fraction of argon which represents the mass ratio of argon partitioning into the gas and melt phases to the total amount originally included in the degassing volume [Tajika and Matsui, 1992]. The effect of processes such as melt generation, partition of volatile elements to the melt, and bubble formation under the mid-ocean ridge are included in this factor. f_{Ar} can be estimated from observation and physical quantities such as partition coefficient, solubility in the silicate melt and so on. [Tajika and Matsui, 1992]. For concentration of potassium into melt, we use f_K and $({}^{40}\text{K})_{mant}$ instead of f_{Ar} and $({}^{40}\text{Ar})_{mant}$. The values of f_{Ar} and f_K are estimated to be 0.93 and 0.99, respectively [Tajika, 1992]. It is noted that the coefficient of degassing rate in this model has a physical meaning (not an adjustable parameter) and varies with time (not constant) since the seafloor spreading rate and the melt generation depth change with the mantle cooling.

The melt generation depth is defined by the depth where the ascending mantle material intersects the mantle solidus, thus extensive melting and melt segregation occur at this depth. We assume that mantle volatile components are released effectively with the melt segregation at this depth. Temporal variation of the melt generation depth can be calculated from that of the average mantle temperature and the mantle solidus profile [Tajika and Matsui, 1992]. On the other hand, temporal variation of the seafloor spreading rate is related to the mantle heat flow. Both the mantle heat flow and the average mantle temperature are obtained by solving the thermal evolution of the mantle [Tajika and Matsui, 1992].

We use the parameterized convection model to calculate the whole mantle convection [e.g., Schubert et al., 1980; Christensen, 1985; McGovern and Schubert, 1989; Tajika and Matsui, 1992]. In this model, we trace the temporal variations of average mantle temperature T_m and heat flow q from the mantle which is parameterized in terms of the Rayleigh number Ra. The equation of conservation of energy is given by

$$\rho C_p(R_m^3 - R_c^3)\frac{dT_m}{dt} = -3R_m^2 q + (R_m^3 - R_c^3)Q \qquad (10)$$

where ρ is the density, C_p is the specific heat at constant pressure, and R_m and R_c are the outer and inner radii of mantle, respectively. Q is the energy production rate by decay of radiogenic heat sources in the mantle. Mantle heat flow q is parameterized in terms of the Rayleigh number Ra as, $q \propto k(T_m - T_s)/(R_m - R_c)Ra^\beta$, where k is the thermal conductivity and T_s is the surface temperature. The factor β is a Rayleigh-Nusselt exponent, usually taken as 0.3 [e.g., Schubert et al., 1980]. Because $q \propto \sqrt{Sr}$ [e.g., Turcotte and Schubert, 1982], we can estimate Sr from q by using a relation of $Sr \propto q^2$ [e.g., Christensen, 1985; McGovern and Schubert, 1989; Tajika and Matsui, 1992]. In this way, we can obtain the average mantle temperature and the seafloor spreading rate by solving the above equations (we call this "model A").

Forsyth and Uyeda (1975) suggested that plate motion may be driven by the forces acting on the downgoing slab and that drag force on the bottom of oceanic plates may be less important. If this were the case, the oceanic lithosphere would couple weakly with the underlying flow because of lower viscosity of asthenosphere [Forsyth and Uyeda, 1975]. This is consistent with the numerical result of thermal history of the mantle with variable viscosity [Christensen, 1985]. According to Christensen (1985), the Rayleigh-Nusselt exponent β should be nearly zero under the condition of the mantle convection, and in such a case seafloor spreading rate becomes to be constant in spite of drastic change in convective vigor. Therefore we consider another model in which β is set to be zero ("the model B"). In this model, the seafloor spreading rate is constant at the present value over the history of the Earth.

We need to know the total amount of potassium to calculate the amount of ${}^{40}\text{Ar}$ produced by the potassium decay. Because it is difficult to estimate potassium content in the present mantle, we use its upper estimate of 400 ppm and lower estimate of 100 ppm derived from the discussion of surface heat flow [Hamano and Ozima, 1978]. Recent geochemical estimates on the abundance of mantle potassium [e.g., Jochum et al., 1983], however, agree well with the average of the two (= 250ppm). We therefore regard 250ppm as the most probable value in this study. The initial average mantle temperature T_m^0 is simply assumed to be in the range of 2000–3000K.

Numerical Results

Figure 1 shows the temporal variations of the seafloor spreading rate, the melt generation depth in the mantle, and
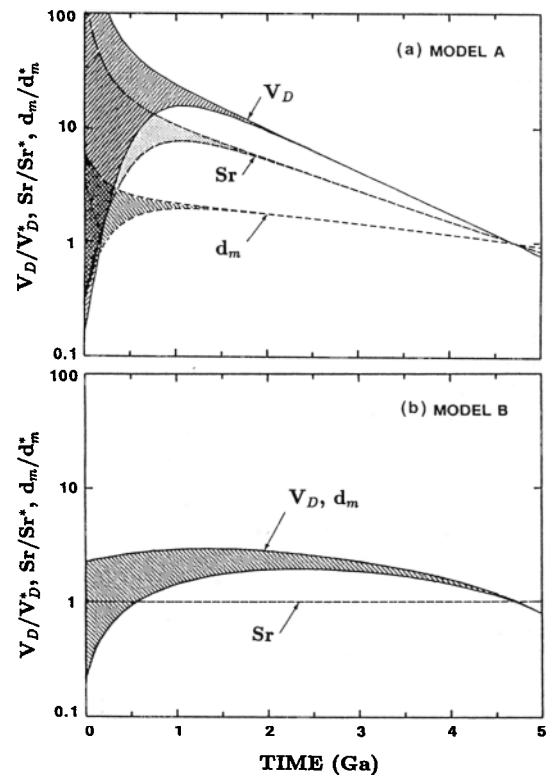

Fig. 1. Temporal variation of the seafloor spreading rate Sr, the melt generation depth in the mantle d_m, and the mantle degassing volume V_D (from which the degassing is assumed to occur) for (a) the models A and (b) the model B. Each value is normalized by the present value ($Sr^* = 2.84 \times 10^{12}\text{m}^2/\text{Ma}$, $d_m^* = 40 \times 10^3\text{m}$, and $V_D^* = 11.4 \times 10^{16}\text{m}^3/\text{Ma}$). Initial mantle temperature is assumed to be 2000–3000 K for the model A and 2000–2500 K for the model B (expressed as hatched area).

the mantle degassing volume from which the degassing is assumed to occur. These variations are closely linked with thermal evolution of the mantle. In the model A, average mantle temperature, mantle viscosity, and mantle heat flow are shown to be greatly changed throughout the history of the Earth. These features are qualitatively similar to the results of previous studies [e.g., Schubert,et al., 1980; McGovern and Schubert, 1989; Tajika and Matsui, 1992]. As a consequence, the seafloor spreading rate in Archean is estimated to be 4–16 times larger than that of today and the melt generation depth at that time to be about 70–100 km (i.e., 2–3 times larger than that of today), resulting in a large mantle degassing volume about 8–45 times greater than that of today in the model A (Figure 1a). This strongly suggests the vigorous degassing during the Hadean and Archean periods, and thus the total amount of ^{40}Ar degassed to the atmosphere would be larger. In the model B, however, degassing from the mantle would be less effective compared to the model A (Figure 1b) because the mantle degassing volume varies only with melt generation depth in this case. The numerical results on ^{40}Ar degassing history are summarized in Table 1.

Figure 2 shows the total amount of ^{40}Ar degassed to the atmosphere as a consequence of each corresponding degassing history (the error bars represent uncertainties of the mantle potassium content, 100–400ppm). In the model A, compared to the observed value, ^{40}Ar has degassed too much to the atmosphere for any initial average mantle temperature (Figure 2a). As shown in Table 1, the degassing fraction X, which represents the ratio of the amount of ^{40}Ar degassed to the atmosphere to the total amount of ^{40}Ar produced by the potassium decay, is about 0.7. These results reflect more effective degassing history of the model A. On the other hand, in the model B, the amount of ^{40}Ar degassed to the atmosphere shows a good agreement with the present observed value (Figure 2b). The degassing fraction X is about 0.35 for this model (Table 1). This means a mild degassing history of the model B. Figure 2c

Table 1. ^{40}Ar degassing history for the models A and B (for the mantle potassium content 250 ppm). T_m^0 is the initial average mantle temperature. The superscript * represents the present value and the overline represents the time average over the history of the Earth. X is the ratio of the amount of ^{40}Ar degassed to the atmosphere to the total amount of ^{40}Ar produced by potassium decay.

model	$T_m^0(K)$	$\overline{Sr}/Sr^*$	$\overline{d_m}/d_m^*$	$\overline{V_D}/V_D^*$	$^{40}Ar^\dagger$	X
A	3000	8.69	1.93	30.75	13.30	0.70
	2500	5.48	1.73	11.12	13.30	0.70
	2000	3.72	1.52	6.44	12.90	0.67
B	2500	1.0	2.33	2.33	8.42	0.44
	2000		1.50	1.50	6.61	0.35
	2500	0.5	2.04	1.02	4.96	0.26
	2000		1.25	0.62	3.53	0.18
	2500	2.0	2.93	5.87	13.30	0.70
	2000		1.89	3.78	11.60	0.61

$\dagger$ amount of ^{40}Ar degassed to the atmosphere in 10^{16} kg at t=4.6 Ga (cf. the observed value is 6.6×10^{16} kg).

and 2d show the results for the cases with different constant seafloor spreading rate (larger and smaller than the present one) for comparison. In the case higher than the present value, the amount of ^{40}Ar degassed to the atmosphere is much larger (Figure 2c), whereas in the case lower than the present value, the amount is much smaller (Figure 2d). This result strongly suggests that the seafloor spreading rate may not have changed greatly from the present value over the Earth's history. It is also suggested that volatile degassing rate from the mantle may have been as same as that of today, with possible variations of at most 2–3 times larger than that of today.

Discussion

The seafloor spreading rate is suggested to have been almost constant during the Earth's history to explain the present amount of ^{40}Ar in the atmosphere. In this respect, the parameterized convection model with $\beta = 0$ is more reasonable than the usual model with $\beta = 0.3$. This would be, in turn, consistent with the idea that the oceanic lithosphere would couple weakly with the underlying asthenosphere because of lower viscosity of the latter. The plausibility of constant seafloor spreaing rate were discussed by Christensen (1985) based on theoretical considerations and geological evidence.

The numerical results, however, include some sort of ambiguities because of simple assumptions for the model and also uncertainties in physical parameters. Therefore, we performed sensitivity analysis with respect to the standard model (model B: T_m^0 =2000 K, $(K)^*_{mant}$ =250ppm) which gives a proper amount of ^{40}Ar degassing to the atmosphere. The results are summarized in Table 2. As shown in Table 2, the ambiguities caused by uncertainties of the model parameters and assumptions may not drastically change the results. Even if we take into account the effects of hot spot volcanism and mantle plumes on the ^{40}Ar degassing, we do not need to change our conclusions. This is simply because much larger amount of ^{40}Ar must have degassed in this case.

The assumptions of homogeneous mantle, constant partition coefficient under the variable conditions from the melt generation depth to the surface, and the definition of the mantle degassing volume would provide another sources for uncertainties. Because the extent of mantle heterogeneity and the change in partition coefficients have not been known, there remain ambiguities due to these assumptions. Although the mantle degassing volume is defined as $Sr \times d_m$ in this study, the studies of flow patterns of ascending mantle materials and
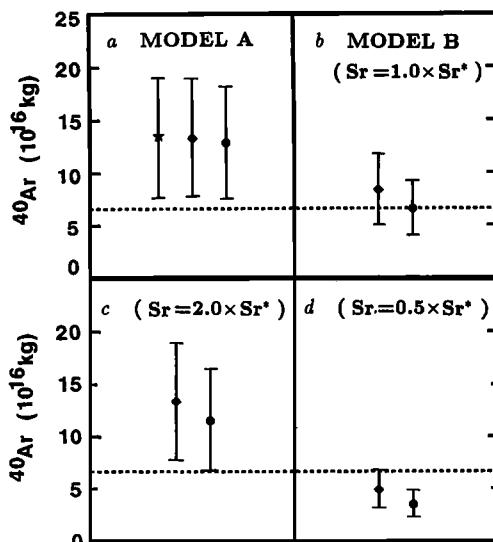

Fig. 2. Total amount of ^{40}Ar degassed to the atmosphere for the models A and B as the consequence of each degassing history (figures b, c, d correspond to the models with constant seafloor spreading rate once, twice, and half of the present value, respectively). The present potassium abundance in the mantle is assumed to be 100–400 ppm (range of change in the results due to potassium abundances is represented by error bars), and the initial mantle temperature is assumed to be 2000 K (circles), 2500 K (diamonds), and 3000 K (stars), respectively. The broken line represents the observed amount of ^{40}Ar in the present atmosphere (6.6×10^{16} kg).

Table 2. Summary of sensitivity analysis

parameters/asumptions	uncertainties	errors[†]
accretion ratio A_K	0~1.0	−5.0% ~ 6.2%
continental degassing rate $K_{D,cc}^{Ar}$	±50%	±3.6%
mantle solidus profile	±50%	±30%
continental growth model	(no continental growth)	−5.0%
	(growth after 2.5Ga BP)	−3.3%
degassing history	(no degassing for the first 1.0Ga)	−6.0%[‡]

[†] Difference between the result of degassed ^{40}Ar for the standard case (model B: T_m^0 =2000K, $(K)_{mant}^*$ =250ppm) and that for sensitivity analysis (values are shown relative to the standard result in %).
[‡] Relative to the result of model A.

melts below the spreading center might suggest more complex geometry of mantle degassing volume [e.g., Spiegelman and McKenzie, 1987]. Here we modify the definition of the mantle degassing volume by introducing the geometrical factor ξ as $V_D \equiv Sr \cdot d_m/\xi$. In order to explain the observed amount of ^{40}Ar by the model A, ξ ~20 is required. This seems to be too large for considering modification due to the flow pattern. Although some of the problems still remain unknown, the model B seems much more reasonable than the model A.

Sleep (1979) has also attempted to constrain the seafloor spreading rate and the amount of mantle potassium by using a simple analytic formulation of the ^{40}Ar degassing and thermal evolution models. According to his result, the average seafloor spreading rate during the last 3 billion years is no more than twice the present value. The difference between his and our results would be partly because he did not take into account the effect of temporal variation of melt generation depth on the degassing rate. As shown in Table 1, the faster constant seafloor spreading rate model (twice the present value) results in deeper average melt generation depth (about 2–3 times the present one), which means the average mantle degassing volume is about 4–6 times the present one. In such a case, as described earlier (Fig. 2c), much more ^{40}Ar degasses to the atmosphere. This suggests that more realistic treatment of temporal variation of melt generation depth is required to constrain the seafloor spreading rate. It is noted that the temporal variation of melt generation depth is an inevitable result as far as the mantle has been cooling over the history of the Earth.

References

Campbell I. H. and S. R. Taylor, No water, no granites—no oceans, no continents, *Geophys. Res. Lett., 10,* 1061-1064, 1983.

Christensen U. R., Thermal evolution models for the Earth, *J. Geophys. Res., 90,* 2995-3007, 1985.

Forsyth D., and S. Uyeda, On the relative importance of the driving forces on plate motion, *Geophys. J. R. Astron. Soc., 43,* 163-200, 1975.

Hamano Y., and M. Ozima, Earth-atmosphere evolution model based on Ar isotopic data, In *Terrestrial Rare Gases,* edited by E. C. Alexander Jr. and M. Ozima, pp.155-171, Japan Scientific Society Press, 1978.

Jochum K. P., A. W. Hofmann, H. M. Seufert, W. M. and White, K, U and Th in mid-ocean ridge basalt glasses and heat production, K/U and K/Rb in the mantle, *Nature, 306,* 431-436, 1983.

Matsui T., and Y. Abe, Evolution of an impact-induced atmosphere and magma ocean on the accreting Earth, *Nature, 319,* 303-305, 1986.

McGovern P. J., and G. Schubert, Thermal evolution of the Earth: effects of volatile exchange between atmosphere and interior, *Earth Planet. Sci. Lett., 96,* 27-37, 1989.

Schubert G., D. Stevenson, and P. Cassen, Whole planet cooling and the radiogenic heat source contents of the Earth and Moon, *J. Geophys. Res., 85,* 2531-2538, 1980.

Sleep N. H., Thermal history and degassing of the Earth: Some simple calculations, *J. Geology, 87,* 671-686, 1979.

Spiegelman M., and D. McKenzie, Simple 2-D models for melt extraction ar mid-ocean ridges and island arcs, *Earth Planet. Sci. Lett., 83,* 137-152, 1987.

Tajika, E., Evolution of the atmosphere and ocean of the Earth: global geochemical cycles of C, H, O, N, and S, and degassing history coupled with thermal history, pp. 416, Doctoral Thesis, University of Tokyo, Tokyo, March 1992.

Tajika, E., and T. Matsui, Evolution of terrestrial proto-CO_2-atmosphere coupled with thermal history of the Earth, *Earth Planet. Sci. Let., 113,* 251-266, 1992.

Tajika, E., and T. Matsui, Degassing history and carbon cycle of the Earth: From an impact-induced steam atmosphere to the present atmosphere, *Lithos* (submitted), 1993.

Turcotte D. R., and G. Schubert, *Geodynamics,* 450pp, Wiley, New York.

Turekian, K. K., Degassing of argon and helium from the Earth, in The Origin and Evolution of Atmosphere and Oceans, edited by P. J. Brancazio and A. G. W. Cameron, pp.74-82, Wiley, 1964.

T. Matsui, Department of Earth and Planetary Physics, Faculty of Science, University of Tokyo, Tokyo 113, Japan.
E. Tajika, Geological Institute, Faculty of Science, University of Tokyo, Tokyo 113, Japan.