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Magma generation on Mars constrained from an ^{40}Ar degassing model

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Abstract. The degassing history of ^{40}Ar on Mars is reconstructed from a model based on the Martian volcanic record. Accumulation history of ^{40}Ar is influenced by the volcanic eruption rate at each Martian epoch in addition to the production rate of ^{40}Ar due to potassium decay in the mantle. In order to degas the present amount of ^{40}Ar to the Martian atmosphere, the volume fraction of erupted magma to the original mantle materials averaged over history should be about 0.006–0.018. The integrated volume of mantle materials that produced the observed volcanic materials and atmospheric ^{40}Ar is about $(4\text{–}11)\times10^{18}$ m^3, hence the extent of mantle differentiation after the formation of Mars is only 0.03–0.08. A significant amount of ^{40}Ar (about 73% of the present atmospheric content) has apparently been degassed during the formation of the major volcanic regions on Mars, such as Tharsis and Elysium. This means that a significant portion of ^{40}Ar in the Martian atmosphere would have been degassed by relatively recent volcanic activity, whereas large fractions of other volatile components, such as H_2O and CO_2, might have been degassed by an early catastrophic degassing. The possibility of ancient plate tectonic activity on Mars, which was proposed recently to explain the origin of the Martian crustal dichotomy, is also examined by using the ^{40}Ar degassing model. Although the possibility of Martian plate tectonics may not be ruled out from the model, we can constrain an upper limit for the duration of plate tectonics. The duration should be much shorter than 6–350 Ma (probably $\ll$100 Ma), which is rather short compared with the ranges of the northern lowlands ages (600–750 Ma).

1. Introduction

Secondary degassing from planetary interiors are believed to form the present atmospheres of the terrestrial planets [*Brown*, 1952]. Degassing histories of the planets may be divided into two stages: an early degassing stage during or within a short period after the planetary formation (stage 1) and a continuous or episodic degassing stage throughout the planetary history (stage 2). Impact degassing [*Lange and Ahrens*, 1982] and/or outgassing during magma ocean cooling [*Sasaki*, 1994] are thought to be plausible processes for stage 1, and outgassing due to subsequent volcanic activity is a process for stage 2 [*Rubey*, 1951]. In this respect, the present atmospheres of terrestrial planets may have information about volcanic activities during planetary histories.

The atmospheric abundance and ratios of radiogenic noble gases, such as ^{4}He, ^{40}Ar, and ^{129}Xe, reveal valuable information about the histories of degassing and volcanism on the terrestrial planets [e.g., *Turekian*, 1964; *Fanale*, 1971; *Hamano and Ozima*, 1978; *Sleep*, 1979; *Allégre et al.*, 1987; *Scambos and Jakosky*, 1990; *Tajika and Matsui*, 1991, 1993; *Namiki and Solomon*, 1994, 1995; *Sasaki*, 1994; *Pepin*, 1994; *Sasaki and Tajika*, 1995; *Matsui and Tajika*, 1995]. In particular, ^{40}Ar is an appropriate tracer that may constrain the degassing and volcanic histories during stage 2. This is primarily because it has been produced from ^{40}K decay, hence (1) the initial amount of ^{40}Ar is very small (^{40}Ar/^{36}Ar$\sim 10^{-4}$ at 4.6Ga ago [*Cameron*, 1970]), and (2) ^{40}K has a half life ($\sim$1.25 Ga) comparable to the timescale of planetary evolution. In other words, ^{40}Ar may not experience the degassing in stage 1 but experienced the degassing due to volcanic activity in stage 2. Therefore volcanic activities on terrestrial planets, such as the average magma eruption rate and the average melt fraction, will be constrained by studying the ^{40}Ar degassing histories [*Tajika and Matsui*, 1991, 1993; *Namiki and Solomon*, 1994, 1995; *Sasaki*, 1994; *Sasaki and Tajika*, 1995; *Matsui and Tajika*, 1995].

Unlike the Earth and Venus, Mars has retained volcanic record spanning most of the planetary history [e.g., *Greeley*, 1987; *Tanaka et al.*, 1988, 1992; *Greeley and Schneid*, 1991]. Hot spot volcanism is thought to be responsible for the large volcanic edifices such as Tharsis and Elysium. From surface images of Mars, erupted vol-

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umes which occurred at each Martian epoch have been estimated [*Greeley*, 1987; *Greeley and Schneid*, 1991]. Combined with absolute ages determined from crater densities [e.g., *Tanaka*, 1986], which are highly dependent on the applied impact flux [*Neukum and Wise*, 1976; *Hartmann et al.*, 1981], magma eruption rate throughout Martian history can be estimated. Because the age and volume of volcanic materials can be estimated for each region and epoch [*Tanaka et al.*, 1988], the Martian degassing history can be reconstructed much more reasonably than those of Venus and even for the Earth, which have suffered intensive resurfacing. For example, *Greeley* [1987] and *Craddock and Greeley* [1995] estimated the amount and timing of release of nonradiogenic gases (such as water) based on the Martian volcanic record. Assuming the nonradiogenic volatiles to be released with the same timing and efficiency as ^{40}Ar, *Scambos and Jakosky* [1990] estimated an outgassing release factor for nonradiogenic volatiles on Mars to be 0.017–0.112 based on the ^{40}Ar release factor (which is estimated to be 0.012–0.040) and a correction coefficient derived from the history of Martian volcanism. In this paper, we show that information on Martian volcanism, such as a volume fraction of erupted magma to the original mantle materials and a total volume of mantle materials that related to Martian volcanism, may be obtained from the record of magma erupted volumes and the atmospheric abundance of ^{40}Ar.

On the other hand, *Sleep* [1994] proposed a hypothesis that the northern lowlands of Mars may have been produced by ancient plate tectonics. This hypothesis might provide an explanation for the origin of Martian crustal dichotomy; that is, the northern hemisphere of Mars consists mainly of topographically low young plains, whereas the southern hemisphere consists of old heavily cratered highlands, and the relative crustal thickness variation across the boundary is 21 km, which is estimated from an average elevation difference of 3 km and an assumption of isostatic compensation [*McGill and Dimitrou*, 1990]. Considering the physics of plate tectonics in the lower gravity of Mars, *Sleep* [1994] concluded that plate motion must have been relatively rapid (~80 mm/yr). If such were the case, ridge volcanism associated with crustal formation would have resulted in an effective volatile degassing of Martian mantle relative to hot spot volcanism and affected the ^{40}Ar degassing history greatly. In this respect, Martian plate tectonics may be constrained from the atmospheric abundance of ^{40}Ar. According to an investigation of geologic evidence on Mars [*Pruis and Tanaka*, 1995], there are many inconsistencies between plate tectonic predictions by *Sleep* [1994] and actual observations; hence Martian plate tectonics seems to be doubtful. However, as *Solomon* [1994] pointed out, because plate tectonics on another planetary body would provide a deeper understanding of the evolution of the Earth, the possibility of *Sleep*'s [1994] proposal deserves to be debated from every possible aspects. Because the study of the ^{40}Ar degassing history may provide a theoretical constraint on the Martian plate tectonics, we examine it as a special case study.

In this study, we reconstruct the ^{40}Ar degassing history of Mars by using an ^{40}Ar degassing model based on volcanic record on Mars. We estimate the volume fraction of erupted magma averaged over the Martian history, and the total volume of mantle materials which produced the observed erupted magma and atmospheric ^{40}Ar. We also estimate the contributions of recent volcanic formations such as Tharsis and Elysium to the ^{40}Ar degassing history. As the special case for ancient plate tectonics on Mars, an upper limit for the duration of presumed plate tectonics is also estimated from the constraints imposed by the ^{40}Ar degassing model.

2. Models

Argon 40 is produced by potassium decay as $^{40}\mathrm{K}+\mathrm{e}^- \rightarrow {}^{40}\mathrm{Ar}$, where the decay constant is λ_e=5.85×10^{-11} yr^{-1}. Because the initial amount of ^{40}Ar is negligibly small ($^{40}\mathrm{Ar}/^{36}\mathrm{Ar} \sim 10^{-4}$), accumulation of ^{40}Ar in the planetary atmospheres directly reflects the ^{40}Ar production and the planetary degassing history.

We calculate degassing of ^{40}Ar by using a simple two-component model. Mass balance equations for potassium and argon are as follows:

$$\frac{d}{dt}[^{40}K]_{\mathrm{man}} = -\lambda_T[^{40}K]_{\mathrm{man}} \tag{1}$$

$$\frac{d}{dt}[^{40}Ar]_{\mathrm{man}} = \lambda_e[^{40}K]_{\mathrm{man}} - K_D[^{40}Ar]_{\mathrm{man}} \tag{2}$$

$$\frac{d}{dt}[^{40}Ar]_{\mathrm{atm}} = K_D[^{40}Ar]_{\mathrm{man}} \tag{3}$$

where K_D is a coefficient of degassing rate, and λ_T is the total ^{40}K decay constant (= 5.305×10^{-10}yr^{-1}). The [^{40}Ar] denotes amount of ^{40}Ar, and subscripts atm and man represent atmosphere and mantle, respectively. It is noted that potassium is an incompatible element and is transported through melt migration. However, we do not take into account the potassium accumulation into the crust for simplicity. The validity of this assumption will be discussed later. We use a recent estimate of potassium abundance in the Martian mantle of 305 ppm based on the composition of SNC meteorites [*Longhi et al.*, 1992; *Wänke and Dreibus*, 1988].

Degassing of ^{40}Ar associated with hot spot volcanism is modeled as follows: a fraction of mantle plume ascends to the surface, then magma generation occurs owing to pressure release melting and ^{40}Ar in a volume of the mantle plume concentrates into the magma, which finally degasses to the atmosphere when the magma erupts to the surface. The degassing rate of ^{40}Ar is expressed as

$$F_D = K_D[^{40}\mathrm{Ar}]_{\mathrm{man}} = f_{\mathrm{Ar}} q_P \frac{[^{40}\mathrm{Ar}]_{\mathrm{man}}}{V_{\mathrm{man}}} \tag{4}$$

where q_P represents the ascending volume flux of mantle plume, and V_{man} is the volume of Martian mantle. The f_{Ar} value is the degassing fraction of argon which represents the mass ratio of argon partitioning into gas and liquid phases to the total amount that originally

Table 1. Magma Eruption Rate Due to Hot Spot Volcanism During the Martian History Estimated for Absolute Age Models of *Hartmann et al.* [1981] (HT) and *Neukum and Wise* [1976] (NW)

Epoch	Absolute Age, Ga		Volcanic Materials, 10^6 km^3	Eruption Rate, km^3/yr	
	HT	NW		HT	NW
Late Amazonian	0.25-0.00	0.70-0.00	2.11	0.008	0.003
Middle Amazonian	0.70-0.25	2.50-0.70	8.49	0.019	0.005
Early Amazonian	1.80-0.70	3.55-2.50	15.76	0.014	0.015
Late Hesperian	3.10-1.80	3.70-3.55	15.63	0.012	0.104
Early Hesperian	3.50-3.10	3.80-3.70	17.65	0.044	0.177
Late Noachian	3.85-3.50	4.30-3.80	7.77	0.022	0.016
Middle Noachian	3.92-3.85	4.50-4.30	1.39	0.020	0.007
Early Noachian	4.60-3.92	4.60-4.50	—	—	—

Data are taken from *Greeley and Schneid* [1991].

included in the mantle plume. The f_{Ar} is assumed to be ~ 1 because of incompatibility of argon [e.g., *Tajika and Matsui*, 1993]. Instead of the volume flux of mantle plume q_P, the volume flux of erupted magma or the magma eruption rate q_M ($= X \cdot q_P$, where X is the volume fraction of erupted magma to the original mantle plume) can be estimated from the observation. Then, the above equation can be rewritten as

$$F_D \sim \frac{q_M}{X} \frac{[^{40}\mathrm{Ar}]_{\mathrm{man}}}{V_{\mathrm{man}}} \qquad (5)$$

where the erupted fraction X is the only independent variable. It is noted that there may be subsurface emplacement of the magma (intrusive volcanism), which cannot be estimated from the surface image of Mars. Therefore the erupted fraction is defined here as a product of the melt fraction ξ and the fraction of extrusive magma to the total η; that is, $X = \eta \cdot \xi$ [*Sasaki*, 1994; *Sasaki and Tajika*, 1995]. Magma eruption rate $q_M(t)$ is estimated from the erupted volumes of volcanic material in each Martian epoch [*Greeley and Schneid*, 1991] with absolute ages determined by *Tanaka* [1986]. We use $q_M(t)$ for two cases corresponding to the two different impact flux models: a NW model, based on *Neukum and Wise* [1976] and a HT model, based on *Hartmann et al.* [1981]. They are shown in Table 1 and Figure 1. We integrated the ^{40}Ar degassing flux from t=0 to 4.6 Ga to obtain the atmospheric amount of ^{40}Ar at present time t_p.

$$[^{40}Ar]_{\mathrm{atm}} = \int_0^{t_p} F_D(t)dt \qquad (6)$$

We adjust a model parameter X so that the total amount of $[^{40}\mathrm{Ar}]_{\mathrm{atm}}$ may be equal to the observed amount. We can also estimate the integrated volume of mantle plumes V_{plume} or the total volume of mantle materials that produced the observed amounts of volcanic materials and atmospheric ^{40}Ar. V_{plume} is defined here as

$$V_{\mathrm{plume}} = \int_0^{t_p} q_P(t)dt = \int_0^{t_p} \frac{q_M(t)}{X}dt. \qquad (7)$$

Using the integrated volume of mantle plumes, we estimate the extent of mantle differentiation after the formation of Mars, which is defined as $\phi = V_{\mathrm{plume}}/V_{\mathrm{man}}$. We use mantle volume of 1.4×10^{20} m^3 corresponding to Martian core radius of 1762 km [*Schubert et al.*, 1992].

Plate tectonics during the earliest history of Mars [*Sleep*, 1994] is another proposed mechanism of volcanism that may have resulted in an effective degassing of ^{40}Ar for that period. Because, at least, some fractions of ^{40}Ar in the present atmosphere must be degassed by hot spot volcanism (such as Tharsis and Elysium), the duration of ^{40}Ar degassing due to mid-ocean ridge volcanism will be constrained from the model which takes into account the ^{40}Ar degassing by these two types of volcanisms. Because of this, we try to apply the ^{40}Ar degassing model to the special case where presumed plate tectonics proposed by *Sleep* [1994] occurred on Mars.

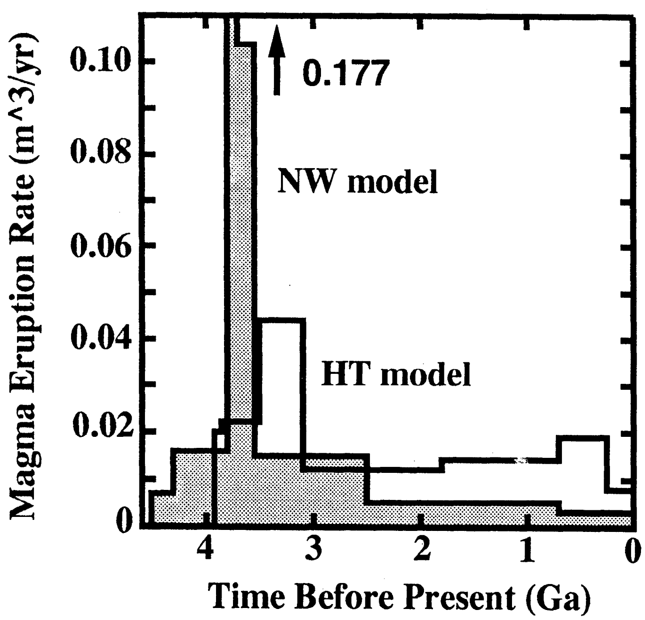

Figure 1. Magma eruption rates during the history of Mars estimated from erupted volume of volcanic materials in each Martian epoch [*Greeley and Schneid*, 1991] and absolute age models [*Tanaka*, 1986]. There are two cases corresponding to the proposed impact flux models; HT model [*Hartmann et al.*, 1981] and NW model [*Neukum and Wise*, 1976].

The volume flux of mantle materials that produces magma and degasses ^{40}Ar due to seafloor spreading depends on the melt fraction and the magma eruption rate. The magma eruption rate for mid-ocean ridge volcanism is called seafloor production rate, which usually includes the eruptions of both extrusive and intrusive magma. Therefore the melt fraction is the only parameter for the degassing model of mid-ocean ridge volcanism. The degassing rate of ^{40}Ar due to mid-ocean ridge volcanism is expressed as

$$F_{D,p} = K_{D,p}[^{40}\mathrm{Ar}]_{\mathrm{man}} \sim \frac{q_{M,p}}{\xi} \frac{[^{40}\mathrm{Ar}]_{\mathrm{man}}}{V_{\mathrm{man}}} \qquad (8)$$

where subscript p denotes plate tectonic activity. We consider three cases for the magma eruption rate at Martian ridges. In Case 1, we apply the Earth's seafloor production rate of 20 km^3/yr and somewhat larger melt fraction of 20% (which corresponds to higher mantle temperature of the early Mars) together with the surface area correction between the Earth and Mars:

$$F_{D,p} = \frac{0.3}{0.6}\left(\frac{R_M}{R_E}\right)^2 \left(\frac{20\ \mathrm{km}^3/\mathrm{yr}}{0.2}\right) \frac{[^{40}\mathrm{Ar}]_{\mathrm{man}}}{V_{\mathrm{man}}} \qquad (9)$$

where R_E and R_M represent radii of the Earth and Mars, respectively. It is noted that the area of oceanic plate is restricted to the northern lowlands of Mars [*Sleep*, 1994], corresponding to ~30% of Martian surface, whereas the area of the terrestrial seafloor is about 60% of the Earth's surface. In case 2, we assume the Earth's seafloor production rate but use the same melt fraction as that for the hot spot magmatism. Therefore

$$F_{D,p} = \frac{0.3}{0.6}\left(\frac{R_M}{R_E}\right)^2 \left(\frac{20\ \mathrm{km}^3/\mathrm{yr}}{\xi}\right) \frac{[^{40}\mathrm{Ar}]_{\mathrm{man}}}{V_{\mathrm{man}}}. \qquad (10)$$

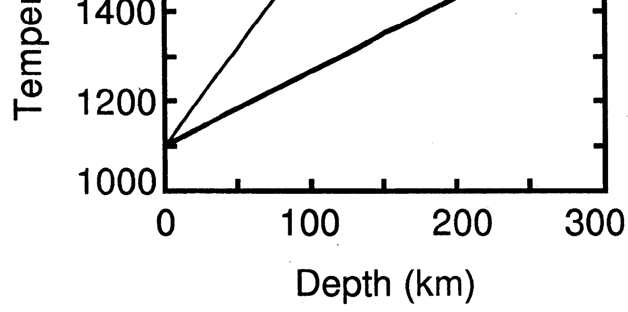

Figure 2. Mantle solidus curves for Mars (solid curve). For comparison, solidus curve for the Earth is also shown (thin solid curve). It is noted that, for ascending mantle materials of the same potential temperature, partial melting occurs much deeper for Mars than those for the Earth.

The seafloor production rate is obtained by the product of seafloor spreading rate and thickness of oceanic crust. In contrast to the previous two simple cases, in case 3, we use a model which combines Martian seafloor spreading rate estimated by *Sleep* [1994] and the crustal thickness estimated from the model of melt generation processes beneath the terrestrial mid-ocean ridges [*McKenzie*, 1984; *McKenzie and Bickle*, 1988]. In this respect, case 3 is regarded as the most suitable model for discussing the specific scenario of Martian plate tectonics proposed by *Sleep* [1994]. The plate velocity on Mars is estimated to be 8 cm/yr, which is obtained by considering the physics of seafloor spreading under the lower gravity of Mars [*Sleep*, 1994]. If the length of the Martian ridge is equal to the length across the northern lowlands (~8000 km [*Sleep*, 1994]), the seafloor spreading rate is estimated to be ~0.64 km^2/yr. As a result,

$$F_{D,p} = \frac{0.64\ \mathrm{km}^2/\mathrm{yr} \times d_c(T_p)}{\xi(T_p)} \frac{[^{40}\mathrm{Ar}]_{\mathrm{man}}}{V_{\mathrm{man}}} \qquad (11)$$

where d_c represents crustal thickness (equals generated melt thickness) and T_p is mantle potential temperature. The potential temperature is defined as the temperature of a fluid mass compressed or expanded adiabatically to some constant reference pressure (the surface pressure in this case). We estimate crustal thickness and melt fraction from the mantle potential temperature by using the model of magma generation beneath the mid-ocean ridges on the Earth proposed by *McKenzie* [1984]. For a given mantle potential temperature, this model predicts the melt generation depth d_m, where ascending mantle material intersects the mantle solidus curve, and the integrated thickness of generated melt, d_c (see *McKenzie* [1984] for more details). The melt fraction is defined here as $\xi \equiv d_c/d_m$. To apply this model to magma generation on Mars, we use the solidus curve of peridotites estimated for the terrestrial mantle [*McKenzie and Bickle*, 1988] to the Martian mantle by considering the lower Martian gravity (Figure 2). The crustal thickness and the melt fraction as a function of mantle potential temperature for Mars estimated from this model are shown in Figure 3. In the original paper of *McKenzie and Bickle* [1988], the solidus curve of mantle peridotites is determined by fitting the experimental data under the pressure conditions of <7 GPa. Therefore this model can be applied to the magma generation in the Martian mantle under the same pressure condition. Because we estimate that the ascending mantle material of $T_p \sim 1700°$C intersects the solidus curve at 7 GPa in the Martian mantle, this model will be applicable under the condition of T_p <1700°C, which corresponds to the condition for the melt fraction of < 0.23 (see Figure 3). This condition will be satisfied throughout our calculations.

We consider ^{40}Ar degassing due to plate tectonic activity in addition to that due to hot spot volcanism, although plate tectonics is assumed to operate during $t_0 < t < t_0 + \tau$. That is,

$$[^{40}Ar]_{\mathrm{atm}} = \int_0^{t_p} F_D dt + \int_{t_0}^{t_0+\tau} F_{D,p} dt. \qquad (12)$$

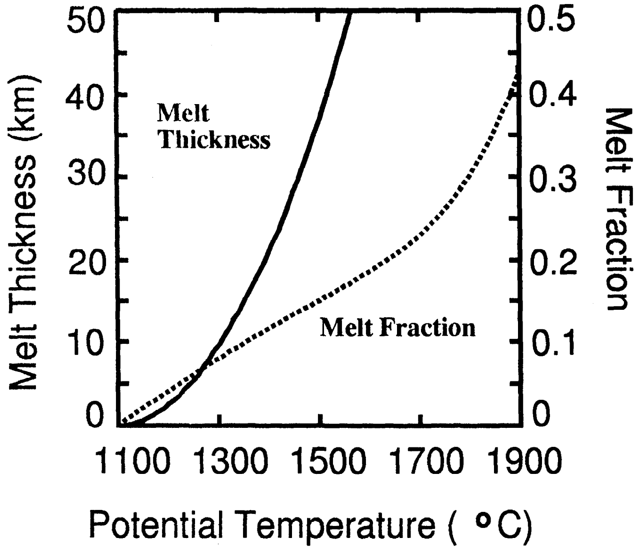

Figure 3. Crustal thickness (equals generated melt thickness, in kilometers) and melt fraction as a function of mantle potential temperature for Mars.

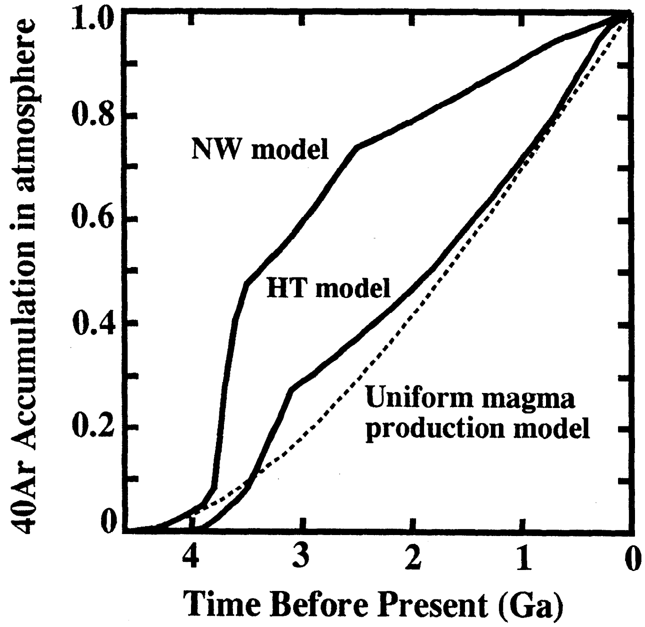

Figure 4. History of ^{40}Ar accumulation in the atmosphere by mantle degassing. Solid curves correspond to the two different impact flux models (NW and HT models), and dotted line represents the case for degassing by an uniform volcanic eruption rate.

According to *Sleep* [1994], seafloor spreading produced thin lowland crust during Late Noachian and Early Hesperian time. We assume that plate tectonics started at the beginning of Late Noachian, that is, t_0=0.75 Ga (HT model) or 0.30 Ga (NW model). The effect of t_0 on the results will be discussed later. The total amount of $[^{40}\mathrm{Ar}]_{\mathrm{atm}}$ integrated to the present time should be equal to the observed amount also in this case. It is noted that an adjustable parameter is melt fraction ξ in cases 1 and 2, whereas it is the mantle potential temperature T_p, and $\xi(T_p)$ also affects hot spot magmatism in case 3.

3. Numerical Results

3.1. Argon 40 Degassing History and Martian Volcanism

We show the standard case of the ^{40}Ar degassing history where ^{40}Ar is degassed only by the hot spot volcanism. Temporal variations of ^{40}Ar accumulated to the Martian atmosphere are shown in Figure 4. For comparison, Figure 4 also shows a temporal variation of ^{40}Ar in the case of an uniform volcanic eruption rate (q_M =68.80×10^6 km^3/4.6×10^{9}yr =0.015 km^3/yr). As shown in Figure 4, the accumulation of ^{40}Ar is obviously influenced by the volcanic eruption rate at each Martian epoch (compare to the case for an uniform volcanic eruption). However, because ^{40}Ar is produced by potassium decay with time, the accumulation of ^{40}Ar to the atmosphere will be small during the early period and then become large during the later period, comparing with the case for the degassing of nonradiogenic gases.

In order to degas the present amount of ^{40}Ar to the Martian atmosphere, the erupted fraction of magma to the original mantle materials averaged over the history of Mars, X, should be 0.018 (HT model) or 0.013 (NW model). Then, the total volume of mantle plumes V_{plume} that produced the observed volcanic materials and the atmospheric ^{40}Ar is estimated to be 3.8 × 10^{18} m^3 (HT model) or 5.4 × 10^{18} m^3 (NW model). The mantle differentiation factor after the formation of Mars is thus estimated to be $\phi = V_{\mathrm{plume}}/V_{\mathrm{man}} \sim 0.027$ (HT model) or 0.038 (NW model). This means that only 3–4% of the mantle materials have experienced partial melting and added volcanic materials to the proto-crust over the history of Mars. These values are listed in Table 2 with the case where the constraining amount of ^{40}Ar in the atmosphere is twice the observed value (which will be discussed later).

After the Late Hesperian epoch, volcanism is dominated by the formation of major volcanic regions such as Tharsis and Elysium [e.g., *Tanaka et al.*, 1988]. In order to estimate the contribution of the formation of these major volcanic regions to ^{40}Ar degassing on Mars, we integrate the amounts of ^{40}Ar degassed to the atmosphere after the Late Hesperian time (3.1 Ga B.P. by

Table 2. The Erupted Fraction X, the Integrated Volume of Mantle Plumes V_{plume}, and the Extent of Mantle Differentiation After the Formation of Mars, ϕ

	$[^{40}\mathrm{Ar}]^{\mathrm{obs}}_{\mathrm{atm}}$		$2\times[^{40}\mathrm{Ar}]^{\mathrm{obs}}_{\mathrm{atm}}$	
	HT	NW	HT	NW
X	0.018	0.013	0.009	0.006
$V_{\mathrm{plume}}(\times 10^{18}\ \mathrm{m}^3)$	3.81	5.37	7.67	10.83
ϕ	0.027	0.038	0.054	0.077

Estimates for the cases where the total amount of degassed ^{40}Ar is once and twice the observed value.

HT model and 3.7 Ga B.P. by NW model) to the present time (Figure 5). Contribution to the ^{40}Ar abundance by volcanic activity after the Late Hesperian is estimated to be ~73% in either model. This suggests that a significant amount of ^{40}Ar in the present Martian atmosphere was degassed by these younger volcanic events, whereas large fractions of other nonradiogenic volatile components such as H_2O and CO_2 might have been degassed by an early catastrophic degassing in stage I due to the planetesimal accretion and a possible magma ocean cooling [e.g., *Lange and Ahrens*, 1982; *Abe and Matsui*, 1985; *Matsui and Abe*, 1986; *Sasaki*, 1994; *Sasaki and Tajika*, 1995].

3.2. Case for Martian Plate Tectonics

We try to estimate a theoretical upper limit for the duration of possible plate tectonic activity on Mars. We consider the ^{40}Ar degassing by mid-ocean ridge volcanism ($t_0 < t < t_0 + \tau$) in addition to that by hot spot volcanism ($0 < t < t_p$). It is noted that the fraction of extrusive magma to the total η is a parameter for the model of hot spot volcanism, although it does

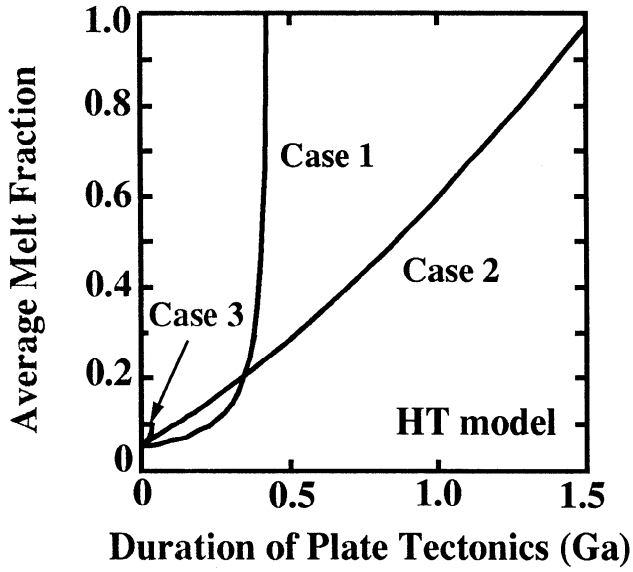

Figure 6. Average melt fraction in order to degas the present amount of ^{40}Ar as a function of the duration of plate tectonics for cases 1, 2, and 3 (HT model). Note that there is a critical duration of plate tectonics for case 3, above which a solution does not exit.

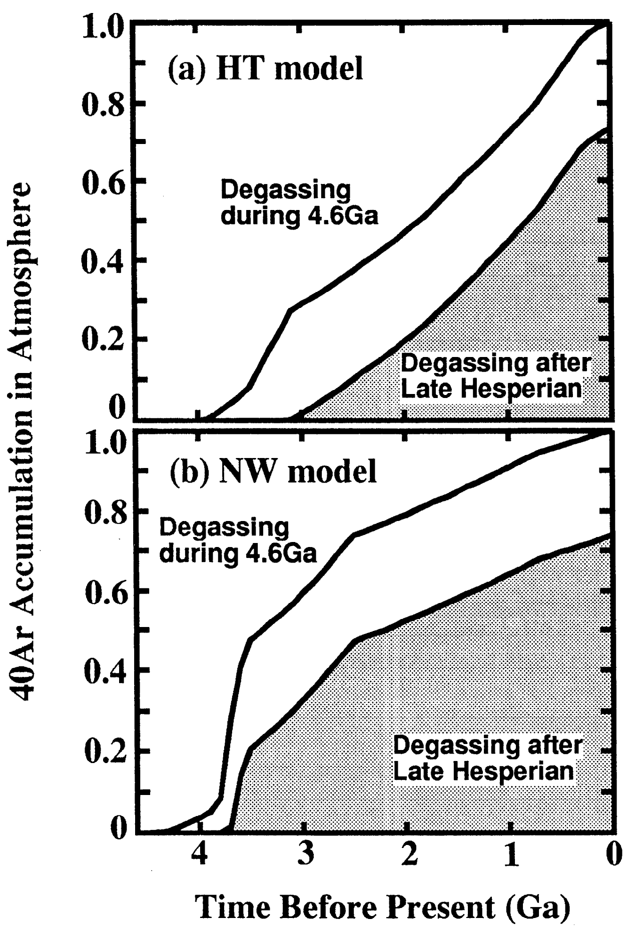

Figure 5. Contribution of degassing due to volcanic activities after Late Hesperian epoch to the ^{40}Ar accumulation for (a) HT model and (b) NW model. About 73% of ^{40}Ar in the present atmosphere would have been degassed by these younger volcanic activities in either case.

not appear in the model for mid-ocean ridge volcanism. Therefore when we assume a specific value of η, we can obtain the relation between the duration of plate tectonics τ and the melt fraction ξ. According to *Crisp* [1984], ratios of intrusive to extrusive magma volumes are typically 5 (which correspond to η=1/6=0.17) for oceanic localities and 10 (η=1/11=0.09) for continental localities on the Earth. On the other hand, we roughly estimate the value of η for Tharsis region on Mars to be ~ 1/3 by using the estimated values of the average elevation of 8–10 km and the excess crustal thickness of 25–30 km for Tharsis region [*Banerdt et al.*, 1992] (here we assume the average elevation to be an extrusive portion and the excess crustal thickness to be extrusive + intrusive portions). Although we do not know which value is appropriate for Mars, we set η=1/3 as a standard case in order to estimate an upper limit for the duration of plate tectonics. Lower values (η=1/6, 1/11) should result in much shorter duration as will be discussed in the next section. The results of the melt fraction as a function of the duration of plate tectonics for cases 1, 2, and 3 are shown in Figure 6. Variations of ξ (or T_p) will change contributions of hot spot volcanism and mid-ocean ridge volcanism to ^{40}Ar degassing (Figure 7), since the degassed amount of ^{40}Ar is fixed.

In case 1 (the case where the corrected Earth's seafloor spreading rate and the fixed melt fraction of 0.2 are assumed for the mid-ocean ridge volcanism), degassing due to hot spot volcanism decreases with an increase in the melt fraction (see equation (5)), while magma eruption rate due to plate tectonic activity remains constant (see equation (9)). As shown in Figure 6, the melt fraction increases gradually then rapidly with the duration of plate tectonics. As a result, the contribution of ^{40}Ar degassing due to hot spot volcanism decreases and that

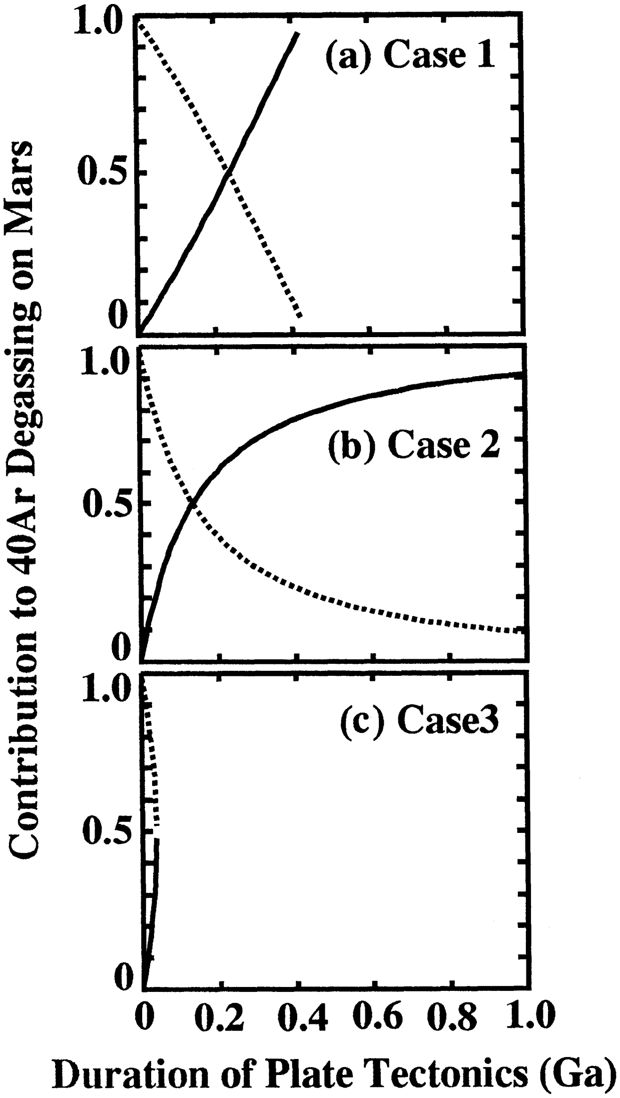

Figure 7. Contributions of degassing due to hot spot volcanism (dotted lines) and due to mid-ocean ridge volcanism (solid lines) to the ^{40}Ar degassing on Mars for cases 1, 2, and 3 (HT model).

due to mid-ocean ridge volcanism increases (Figure 7a). Duration of plate tectonics for a probable value of melt fraction would be as long as 300 for $\xi \sim 0.05$, 380 Ma for $\xi \sim 0.1$, and 410 Ma for $\xi \sim 0.2$ (HT model; see Figure 6 and Table 3). Therefore, if plate tectonics had operated on Mars, it would have lasted about <400 Ma (HT model) or <600 Ma (NW model) at most.

In case 2 (the case where the corrected Earth's seafloor spreading rate is assumed and the same parameter ξ is used for two types of volcanisms), both degassing by hot spot volcanism and that by mid-ocean ridge volcanism should decrease with an increase in the melt fraction (see equations (5) and (10)). Because the melt fraction also increases with the plate tectonics duration (Figure 6), the total degassing rate of ^{40}Ar should decrease, resulting in the duration of plate tectonics to be much longer than the case 1 (Figure 7b). It is, however, noted that τ for $0 < \xi < 0.2$ is shorter than the case 1 (Figure 6, Table 3). As a result, presumed plate tectonics would have lasted about <400 Ma (HT model) or <600 Ma (NW model) also in this case.

In case 3, we use much more reasonable estimate for the seafloor spreading rate and the melt thickness. In this case, degassing due to hot spot volcanism should decrease with an increase in mantle potential temperature through melt fraction. It is because a large melt fraction at high potential temperature corresponds to small volume of the original mantle plumes that can degas small amount of ^{40}Ar, since we fix the erupted magma volume and the fraction of extrusive magma in the model. However, degassing due to plate tectonic activity increases with mantle potential temperature as will be shown in the next section. As a result, the duration of plate tectonics tends to be short compared with cases 1 and 2 (Figure 7c). In this case, it is noted that there is a critical value for the duration of plate tectonics, τ_{cr} =35 Ma (HT model) or 103 Ma (NW model), as shown in Figures 6 and 7 and Table 2. Above the critical value, no solutions exist. The reason will be explained in the next section. As the melt fraction increases with an increase in the duration of plate tectonics (Figure 6), the critical value of the melt fraction, ξ_{cr}, is also regarded as an upper limit for average melt fraction for a given η (Table 3). Among three cases we examined, case 3 gives the shortest duration of plate tectonic activity on Mars.

4. Discussion

4.1. Implication for Volcanism and Differentiation on Mars

We assumed that all ^{40}Ar degassed from the mantle accumulated to the atmosphere. In other words, we did not take into account the escape of ^{40}Ar to space. For example, atmospheric erosion due to cometary impacts

Table 3. Upper Limit for the Duration of Martian Plate Tectonic Activity

Age Model	Case 1			Case 2			Case 3	
	$\tau_{0.05}$*	$\tau_{0.1}$†	$\tau_{0.2}$‡	$\tau_{0.05}$*	$\tau_{0.1}$†	$\tau_{0.2}$‡	τ_{cr}	ξ_{cr}
HT	300	375	407	82	200	407	35	0.103
NW	521	586	615	180	354	616	103	0.075

*At $\xi \sim 0.05$.
†At $\xi \sim 0.1$.
‡At $\xi \sim 0.2$.

could have resulted in removal of Martian atmosphere [*Walker*, 1986; *Melosh and Vickerly*, 1989; *Kuramoto and Matsui*, 1992; *Zahnle*, 1993]. However, xenological constraints, such as the relatively high absolute abundance of radiogenic ^{129}Xe, imply that Mars lost its nonradiogenic noble gases by impact perhaps before it was 100 Ma old [*Zahnle*, 1993]. Even if the influence of impact erosion continued, it might be limited to the earliest history of Mars because of a decrease in impact flux with time. If the effective impact erosion was limited before Late Noachian (when only a small amount of ^{40}Ar accumulated as shown in Figure 4), its influence in the ^{40}Ar accumulation history might be negligible. On the other hand, *Jakosky et al.* [1994] proposed that pickup ion sputtering by solar winds and photochemical escape should decrease atmospheric species. Using isotopic fractionation of ^{38}Ar/^{36}Ar, about half of ^{40}Ar may have been lost from the Martian atmosphere (B. M. Jakosky, personal communication, 1994). To examine this, we simulate the case where the total ^{40}Ar is twice as large as the present amount. In this case, the average erupted fraction of Martian magmatism over 4.6 Ga should be $X = 0.009$ (HT model) or 0.006 (NW model) as shown in Table 2. The total volume of mantle plumes and the mantle differentiation factor corresponding to this are $V_{\text{plume}} = 7.7 \times 10^{18}$ m^3 (HT model) or 1.1×10^{19} m^3 (NW model) and $\phi = V_{\text{plume}}/V_{\text{man}} \sim 0.054$ (HT model) or 0.077 (NW model), respectively (Table 2). In any cases, the estimated range of the erupted fraction X (=0.006–0.018) seems to be very small.

Next, we consider the contribution of the melt fraction ξ and the fraction of extrusive magma to the total η to the erupted fraction. As mentioned earlier, estimates of ratios of intrusive to extrusive magma volumes are typically 5 (which correspond to η=1/6=0.17) for oceanic localities and 10 (η=1/11=0.09) for continental localities on the Earth [*Crisp*, 1984]. *Greeley and Schneid* [1991] estimated the total volume of extrusive and intrusive magma generated on Mars by assuming the ratio of intrusive to extrusive of 8.5 (η=0.11). When we use X=0.018 (HT model), we can estimate the melt fraction ξ (=X/η) for η=0.09 to be 0.201, and for η=0.17 to be 0.106. Although there is no estimate for the melt fraction on Mars, *Matsui and Tajika* [1995] estimated temporal variations of mantle temperature, magma generation, melt fraction, and ^{40}Ar degassing by using the constraint of ^{40}Ar abundance and the thermal evolution model of Mars. They estimated the melt fraction averaged over the Martian history to be $\sim$0.10 [*Matsui and Tajika*, 1995]. Therefore the average melt fraction estimated for η $\sim$0.17 might be reasonable. Because it is unlikely that Martian large volcanoes, whose heights are 25-26 km, should hide nearly 10 times larger intrusive volumes beneath the surface, a larger value of η on Mars than that on the Earth will be expected [*Sasaki*, 1994; *Sasaki and Tajika*, 1995]. It is noted that degassing of ^{40}Ar from intrusive magma would not be as efficient as that from erupted magma. In this case, the parameter η should represent the fraction of extrusive magma to the net volume of generated magma that actually contributes to the ^{40}Ar degassing. Although the actual melt fraction would be similar to the value of X, the actual erupted fraction is smaller than X in this case.

We estimated V_{plume} as a volume of mantle materials required to degas the present amount of atmospheric ^{40}Ar from the Martian mantle by assuming that ^{40}Ar can be treated as a highly incompatible element (i.e., $f_{\text{Ar}} = 1$) and the Martian mantle is homogenous. It is noted that, because there is a possibility of inefficient degassing from intrusive magma as mentioned above, the values of V_{plume} and ϕ (which is derived from V_{plume}) can be regarded as, at least, lower estimates for them. *Taylor and McLennan* [1985] estimated the extent of mantle differentiation to be 30–50% for the Earth from Rb/Yb of continental crust and depleted mantle. In this respect, the estimated value of ϕ (=0.03–0.08) for Mars is very small. This is probably because Mars is a small planet, which results in a rapid cooling and lack of long-lived plate tectonics (even if plate tectonics proposed by *Sleep* [1994] could operate on Mars as discussed below). Although the extent of core and mantle differentiation of Mars is poorly known, isotope systematics of SNC meteorites suggest core formation essentially contemporaneously with the completion of accretion [e.g., *Schubert et al.*, 1992]. The differentiation of Martian mantle, which produced the proto-crust, was probably completed in the earliest history of Mars, and the volume of mantle material that related to the later differentiation would be very small, about $\geq$3–8% of the Martian mantle volume. Because of this, an assumption of no transport of K from the mantle to the crust in our model may be justified (only 3–8% of the total K in the mantle would be transported to the crust by Martian volcanism).

4.2. Constraints on Martian Plate Tectonics

We estimate the duration of Martian plate tectonics to be <400–600 Ma. In particular, the upper limit of τ obtained from case 3, which is regarded as the most appropriate model for Martian plate tectonics, should be smaller than 35–100 Ma. Because case 3 is the most suitable model for discussing the specific scenario of Martian plate tectonics proposed by *Sleep* [1994], we will concentrate our discussion on the parameter study of case 3 below.

Before performing the parameter study, we discuss the mathematical behavior of the model. In case 3, there is a critical value for the duration of plate tectonics τ_{cr}. Above the critical value, no solutions exist. The reason is clearly demonstrated in Figures 8 and 9. The amount of ^{40}Ar degassed by hot spot volcanism decreases with an increase in melt fraction, that is, $dF_D/d\xi < 0$ (see equation (5)). However, because $F_{D,p} \propto d_c/\xi = d_m$ (see equation (11)), $dF_{D,p}/d\xi \propto (dd_m/dT_p)/(d\xi/dT_p) > 0$ where $dd_m/dT_p > 0$ and $d\xi/dT_p > 0$ [e.g., *McKenzie*, 1984; *McKenzie and Bickle*, 1988] (see also Figure 3). Therefore the amount of ^{40}Ar degassed by mid-ocean ridge volcanism increases with melt fraction. Consequently, the total amount of degassed ^{40}Ar has a minimum as melt fraction increases (represented as C in Figure 8). Because solutions are

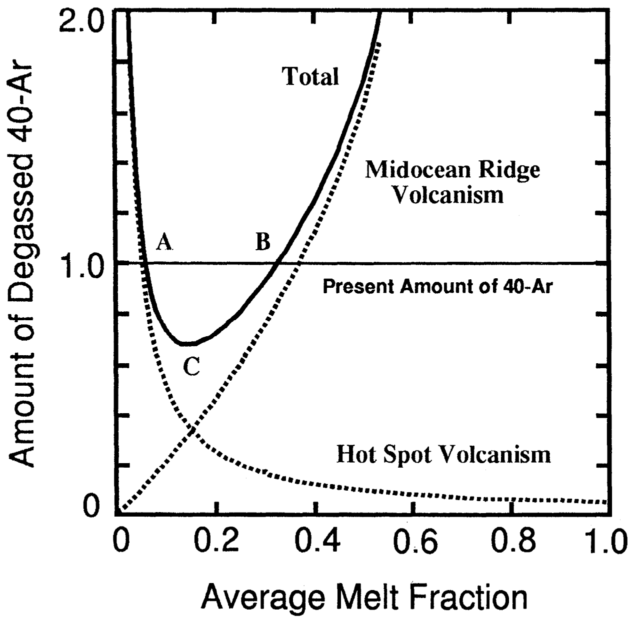

Figure 8. Mathematical behavior of solutions for case 3 with small τ (=16 Ma). Dotted lines are amounts of ^{40}Ar degassed due to hot spot volcanism and mid-ocean ridge volcanism. Solid line represents total amount of ^{40}Ar degassed by both two types of volcanic activities. There is a minimum (represented by C) of the total ^{40}Ar curve along with increasing melt fraction. Note that there are two solutions in this case, which are expressed as cross points (represented by A and B) between the total degassed ^{40}Ar curve and the horizontal line (the present amount of ^{40}Ar).

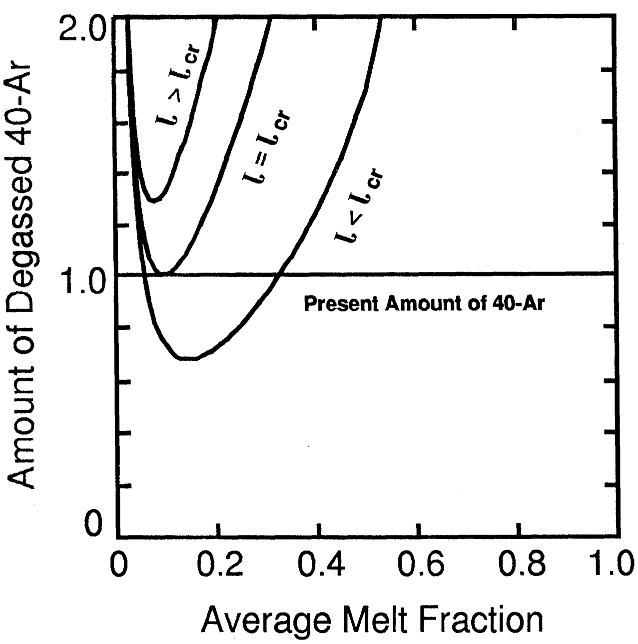

Figure 9. Mathematical behavior of solutions for case 3 with various τ (= 16, 35, 60 Ma). There is a critical value τ_{cr} =35 Ma above which a solution does not exit. Number of solutions n should vary with increasing τ: n =2 for $\tau < \tau_{cr}$, n =1 for $\tau = \tau_{cr}$, and n =0 for $\tau > \tau_{cr}$.

obtained by cross points between the total ^{40}Ar curve (solid curve) and the present amount of ^{40}Ar in the atmosphere (horizontal line) in Figure 8, there are two solutions when the minimum C is smaller than the present amount (solutions are represented by A and B in Figure 8). Solution A is the case where contribution of degassing due to hot spot volcanism is dominant, and solution B is the case where contribution of degassing due to mid-ocean ridge volcanism is dominant. Solution B is, however, considered to be physically unreasonable because average melt fraction is considered to be too large ($\sim$ 0.3) for Mars (this may be beyond the limit of application of the model). So, only the solution A is shown in Figures 6 and 7. Although two solutions are obtained for small τ, solution disappears for larger τ when the minimum C becomes larger than the present amount (Figure 9). The critical value τ_{cr}, above which a solution does not exist, is estimated to be 35 Ma (HT model) or 103 Ma (NW model), and corresponding critical melt fraction ξ_{cr} is 0.103 and 0.075, respectively (Table 3).

Then, we have examined the critical value of plate tectonics duration τ_{cr} for different parameters in case 3. The results are summarized in Table 4. If we consider an escape of ^{40}Ar to space (the case where the total degassed ^{40}Ar is twice as large as the present amount), the upper limit of τ is estimated to be 145 Ma (HT model) and 345 Ma (NW model). On the other hand, seafloor spreading rate on the Earth ranges 12–18 cm/yr for oceanic plates with subducting slab, and 2–6 cm/yr for most of plates without slab [*Forsyth and Uyeda*, 1975]. Although the presumed oceanic plate on Mars should have subducting slab, we estimate τ_{cr} in the cases where seafloor spreading rate is assumed to be 16 cm/yr (=2×8 cm/yr) and 4 cm/yr (=1/2×8 cm/yr). A faster assumed velocity would shorten the duration (τ_{cr}=17 Ma) and a slower velocity would prolong the duration (τ_{cr}=70 Ma) by suppressing ^{40}Ar degassing rate (HT model; see Table 4). Next, we change a parameter t_0 (the beginning of plate tectonics). When

Table 4. Parameter Studies for the Critical Values of τ and ξ

Case	HT Model		NW Model	
	τ_{cr}, Ma	ξ_{cr}	τ_{cr}, Ma	ξ_{cr}
Standard case	35	0.103	103	0.075
Constraining amount of degassed $[^{40}Ar]_{atm}$				
×2	145	0.053	345	0.037
Assumed plate velocity				
×2	17	0.104	55	0.075
×1/2	70	0.103	188	0.075
Beginning of plate tectonics				
Early Hesperian	26	0.103	49	0.074
0 Ga	222	0.105	267	0.075
Extrusive magma to the total η				
1/4	25	0.136	78	0.099
1/6	16	0.198	51	0.144
1/8	11	0.255	37	0.187
1/11	6	0.332	25	0.248

we assume larger value of t_0, an upper limit of the duration of plate tectonics will be shorter because larger amounts of ^{40}Ar accumulates in the mantle, which results in larger degassing rate at mid-ocean ridges. For example, when we assume that plate tectonics begins at Early Hesperian (which corresponds to t_0=1.1 Ga (HT model) or 0.8 Ga (NW model) as shown in Table 1), the critical duration τ_{cr} is 26 Ma (HT model) or 49 Ma (NW model) as shown in Table 4. Similarly, although there is no geologic evidence, the theoretical upper limit for the duration of plate tectonics on Mars is obtained from the case of $t_0 = 0$, because $[^{40}\mathrm{Ar}]_{\mathrm{man}}$ is nearly zero at $t_0 = 0$. In this case, τ_{cr} is estimated to be 222 Ma (HT model) or 267 Ma (NW model) as shown in Table 4. All these values are still very short. On the other hand, we assumed that the fraction of extrusive magma to the total for hot spot volcanism, η, is 1/3 (the value estimated for Tharsis region) in order to obtain the upper limit for τ_{cr}. When we assume a smaller value of η (<1/3), we obtain much shorter τ as shown in Table 4. The reason is as follows. In order to maintain the solution obtained for the standard case, the value of X may not be changed greatly. Because $X = \eta\xi$, small η results in large ξ_{cr} as shown in Table 4. Large ξ_{cr} means high T_p which results in large d_c and d_m [e.g., *McKenzie*, 1984; *McKenzie and Bickle*, 1988](see also Figure 3). Therefore $dF_{D,p}/d\eta \propto (dd_m/d\eta) = (dd_m/dT_p) \times (d\xi/d\eta)/(d\xi/dT_p) < 0$, where $(dd_m/dT_p) > 0$, $(d\xi/dT_p) > 0$, and $(d\xi/d\eta) < 0$. As a results, the duration of plate tectonics, τ_{cr}, becomes short in order to degas the fixed amount of ^{40}Ar degassed to the atmosphere. Because η for the Earth is estimated to be 1/6–1/11 [*Crisp*, 1984], η is more likely to be <1/3 (although η could be larger than 1/11 on Mars as discussed earlier). Therefore the value of τ_{cr} estimated for the standard case ($\eta = 1/3$) can be regarded as the upper limit. The duration of plate tectonics proposed by *Sleep* [1994] will be therefore much shorter than the value obtained for the standard parameters.

The timescale of plate tectonic duration constrained by the ^{40}Ar degassing model is shorter than the estimated range of age of the northern lowlands on Mars, that is, from Late Noachian to Early Hesperian time [*Sleep*, 1994]. It corresponds to 3.85-3.10 Ga B.P. [*Hartmann et al.*, 1981] or 4.30-3.70 Ga B.P. [*Neukum and Wise*, 1976], so the range of northern lowlands ages is 600–750 Ma. It takes 100 Ma for full seafloor spreading rate of 8 cm/yr to resurface the northern lowland plains of about 8000 km wide. According to Sleep's reconstruction, the time required for the ridge to move from the breakup margin to the trench is 200 Ma [*Sleep*, 1994]. These values are still larger than, or comparable to the upper limit for τ (note that the duration of plate tectonics for smaller η is estimated to be $\ll$100 Ma as shown in Table 4). This suggests that the duration of Martian plate tectonics is too short to replace the crust of northern lowland plains more than two times or that the plate tectonics has never operated on Mars. When we consider many uncertainties in assumption and simplification of the model, the ^{40}Ar constraint may not rule out a possibility of ancient plate tectonics on Mars. However, because the upper limit of the duration of plate tectonics is very short compared with the ranges of the northern lowlands age, it seems to be difficult to support the idea of Martian plate tectonics, at least, the specific reconstruction proposed by *Sleep* [1994].

5. Conclusion

We have reconstructed the degassing history of ^{40}Ar on Mars based on the Martian volcanic record, although there is uncertainty in absolute age model to estimate the history of Martian volcanic activities. Accumulation history of ^{40}Ar is influenced by the volcanic eruption rate at each Martian epoch in addition to the production rate of ^{40}Ar due to potassium decay in the mantle. In order to degas the present amount of ^{40}Ar to the Martian atmosphere, the erupted fraction of Martian magmatism averaged over the history should be about 0.006–0.018. The integrated volume of mantle plumes is about $(4–11)\times10^{18}$ m^3, and the extent of mantle differentiation after the formation of Mars is 0.03–0.08. Significant amount of ^{40}Ar, about 73% of the present atmospheric content, would have been degassed by the volcanic activities in the major volcanic regions on Mars, such as Tharsis, and Elysium. We have also examined a possibility of ancient plate tectonic activity on Mars proposed by *Sleep* [1994]. From the models of ^{40}Ar degassing due both to hot spot volcanism and mid-ocean ridge volcanism, we constrained the upper limit of duration of plate tectonics to be no longer than several hundreds million years. In particular, from the most appropriate model of the specific scenario proposed by *Sleep* [1994], the upper limit of the duration of plate tectonics is much shorter than 6–200 Ma (HT model) or 25–350 Ma (NW model). When we consider many uncertainties in assumption and simplification of the model, the ^{40}Ar degassing constraint may not rule out a possibility of ancient plate tectonics on Mars. However, because the upper limit of the duration of plate tectonics is very short compared with the ranges of the northern lowlands ages, it seems to be difficult to support the idea of Martian plate tectonics proposed by *Sleep* [1994].

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