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The young age of Earth

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Abstract—Patterson (1956) established that the age of Earth is close to that of meteorites. Over the last 20 years, workers argued for younger age for core differentiation based on Pb-Pb model ages and tungsten isotopic data and for gas retention based on I-Xe modeling. However, disagreement is abundant, and the young age of Earth has not been widely accepted. In this work, I examine all radiogenic noble gases in the atmosphere and use a model-independent approach and total inversion to show that (1) the Xe-closure age of Earth is 109 ± 23 million years younger than the formation of meteorite Bjurbole (~4560 Ma) and (2) all radiogenic components of noble gases in the atmosphere can be quantitatively accounted for by production and degassing ~60% of the bulk silicate earth. The agreement between the ^{129}I-^{129}Xe clock and ^{244}Pu-^{238}U-^{136}Xe-^{134}Xe-^{132}Xe-^{131}Xe clock suggests that the volatility of iodine does not affect the ^{129}I-^{129}Xe clock. Earth's Xe-closure age is 4.45 ± 0.02 Ga, consistent with the model age of Pb and the ^{146}Sm-^{142}Nd, ^{147}Sm-^{143}Nd and ^{182}Hf-^{182}W systematics. On the basis of the consistency of these ages, 4.45 ± 0.02 Ga probably represents the time when the last Martian-sized planetesimal hit Earth and reinitialized the global clocks. *Copyright © 1998 Elsevier Science Ltd*

1. INTRODUCTION

The age of Earth is a question of great interest. Patterson (1956) established that the age of Earth is close to that of meteorites and is 4.55 Ga. However, the resolution of lead isotopes is not enough to distinguish possible small differences between the age of Earth and that of most meteorites. Furthermore, the formation of Earth (or any other planet) was a complicated process and must have taken a long time. The most straightforward definition of the age of Earth is probably the time when the mass of Earth reached 90% (or another specific fraction) of the present-day mass. Because this cannot be obtained at present, other events, such as core separation, closure of Earth to heavy noble gas loss (similar to closure age of a mineral to Ar loss), are used to date Earth. In such cases, the specific meaning of the age must be clarified.

Since the monumental work by Patterson (1956), meteorites have been found to have slightly variable ages with the oldest meteorites roughly 4.56–4.57 Ga (e.g., Swindle and Podosek, 1988; Allegre et al., 1995a). Workers have argued for a younger age of Earth based on several isotopic systems. Wetherill (1975) used the I-Pu-Xe system to estimate the age of Earth to be 113–127 million years younger than that of most meteorites. Staudacher and Allegre (1982) carried out a similar treatment of the I-Pu-Xe system and arrived at the conclusion that Earth was 50–70 million years younger than the meteorites. The difference between the result of Wetherill (1975) and that of Staudacher and Allegre (1982) is owing to differences in input parameters. Recent progress in geochemistry has improved some of the input data significantly (Hudson et al., 1989; Deruelle et al., 1992; McDonough and Sun, 1995). Allegre et al. (1995a) revisited the age of Earth using U-Pb system (using ^{207}Pb*/^{206}Pb* in MORB where the superscript * signifies radiogenic component) and ^{129}I-^{129}Xe method. They obtained results somewhat different from that of Staudacher and Allegre (1982) and concluded that the absolute age of Earth is about 100 million years younger than that of the meteorite Bjurbole, and the age of core differentiation about 110 million years younger than that of meteorites. Lee and Halliday (1995) used the Hf-W system to show that the age of core formation must be ≥50 million years younger than meteorite formation.

Although the point that the age of Earth (estimated by a variety of methods) is younger than that of meteorites has been made on and off for 20 years, some of the calculations are model dependent and internal consistency is not always insured. Disagreement is abundant and the younger age of Earth has not been widely accepted. In estimating the age of core differentiation using the U-Pb system, Allegre et al. (1995a) used the simple average ^{207}Pb*/^{206}Pb* in MORB but did not account for the large scatter in ^{207}Pb*/^{206}Pb* ratio (with equivalent age up to >4.6 Ga). The use of I-Pu-Xe method has also been questioned. Azbel and Tolstikhin (1993) argued that the ^{129}I-^{129}Xe clock is questionable because iodine is highly volatile. Meshik et al. (1995) suggested that there is no ^{129}I-generated Xe signature. Although the Hf-W system seems to provide a firm limit on the age of core formation (Lee and Halliday, 1995; Halliday et al., 1996), Jacobsen and Harper (1997) claimed that a mean time of core formation of only 2–15 million years is allowed by the Hf-W isotopic data. Another area of confusion arises from the meanings of such ages. For example, Allegre et al. (1995a) referred to the I-Xe age as the mean age of atmosphere outgassing in one place and the absolute age of Earth in another.

In this paper, I use a model independent approach to evaluate the mean Xe closure age of Earth. I emphasize model independence because avoiding complicated chemical geodynamics models hopefully will increase the transparency of the treatment. Hence degassing models (whether solubility-controlled, steady-state, or bulk degassing models) are not used. The rate, timing, and history of mantle degassing are not needed. Xenon isotopy in the mantle will not be considered because these are model dependent. In other words, the mean Xe-closure age is

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Table 1. Some xenonisotopic ratios and radiogenic xenon

	^{129}Xe	^{130}Xe	^{131}Xe	^{132}Xe	^{134}Xe	^{136}Xe
Contemporary atmospheric Xe	6.4958 ± 0.0058	≡1	5.2127 ± 0.0059	6.6068 ± 0.0053	2.5628 ± 0.0037	2.1763 ± 0.0022
Nonradiogenic atmospheric Xe	6.053 ± 0.029	≡1	5.1873 ± 0.071	6.518 ± 0.013	2.470 ± 0.013	2.075 ± 0.013
Amount of radiogenic Xe in air (10^9 mol)	277 ± 18	≡0	15.9 ± 4.4	55.6 ± 8.3	58.1 ± 8.1	63.2 ± 8.2

Data are from Pepin (1991).

an age for the whole earth (including the atmosphere) to become a closed-system, not the mean age for the mantle to become a closed-system. Unlike the mean degassing age, the mean Xe-closure age is independent of degassing models.

Input data are from most recent evaluations. A total inversion technique is used to determine a single Xe closure age using up-to-date data in the ^{129}I-^{129}Xe and ^{244}Pu-^{238}U-^{136}Xe-^{134}Xe-^{132}Xe-^{131}Xe system. The results show that all the data available at present are consistent with a single mean Xe closure age ~110 million years younger than that of Bjurbole. The consistency of all the input data from a variety of sources and methods means that independent new evaluation of input data is not necessary for this work. Furthermore, the high internal consistency between the ^{129}I-^{129}Xe clock and ^{244}Pu-^{238}U-^{136}Xe-^{134}Xe-^{132}Xe-^{131}Xe clock reinforces each other and suggests that the volatility of iodine does not affect the estimation of Xe closure age, contrary to the claim of Azbel and Tolstikhin (1993).

2. I-PU-U-XE SYSTEM

Xenon is a noble gas element with nine stable isotopes (mass number 124, 126, 128, 129, 130, 131, 132, 134, 136). Among these, ^{129}Xe, ^{131}Xe, ^{132}Xe, ^{134}Xe, and ^{136}Xe receive a fission contribution from ^{244}Pu (an extinct nuclide with a half life of 80 million years) and ^{238}U. The fission contributes mostly to ^{136}Xe and ^{134}Xe but only negligibly to ^{129}Xe. Radiogenic ^{129}Xe is produced by decay of ^{129}I, an extinct nuclide with a half life of 15.7 million years.

Variations in xenon isotopic ratios in Earth and meteorites are complex. The xenon isotopic ratios in the atmosphere are well-known (Table 1). Some of these ratios have been increased from the primordial ratios due to radiogenic growth. Nonradiogenic xenon isotopic ratios in the atmosphere have been obtained by Pepin (1991; see also Igarashi and Ozima, 1988) and are also shown in Table 1. The nonradiogenic xenon isotopic ratios of Pepin (1991) are somewhat model dependent. However, since the Pepin (1991) model did not use age as constraint, there is no circular argument. Using these ratios, the amount of radiogenic ^{129}Xe*, ^{131}Xe*, ^{132}Xe*, ^{134}Xe*, and ^{136}Xe* can be calculated (Table 1). The amount of each radiogenic xenon isotope in the atmosphere can be related to the parent concentration at the time when Earth became closed to Xe. For ^{129}Xe*, the equation is

$$^{129}\mathrm{Xe}^* = (^{129}\mathrm{I})_T F_{129},$$

where ^{129}Xe* is the amount of radiogenic ^{129}Xe in the atmosphere, $(^{129}\mathrm{I})_T$ is the amount of ^{129}I in the bulk silicate earth (hereafter referred to as BSE) at T (the mean Xe-closure age of Earth), and F_{129} is the fraction of ^{129}Xe* that is in the atmosphere over total ^{129}Xe* production in BSE. It is assumed that radiogenic xenon isotopes at T is negligible for obtaining the mean closure age. The meteorite Bjurbole is used as the reference point for age determination following I-Xe age determinations (e.g., Swindle and Podosek, 1988) and because consistent ^{129}I/^{127}I and ^{244}Pu/^{238}U initial ratios are obtained relative to this meteorite (Swindle and Podosek, 1988; Hudson et al., 1989). The age of Bjurbole (referred to as T_0 hereafter) is not precisely and unambiguously known but is very close to 4.56 Ga, almost certainly within 0.01 Ga (Hudson et al., 1989). The amount of $(^{129}\mathrm{I})_T$ in the above equation can be expressed as $(^{127}\mathrm{I})_T(^{129}\mathrm{I}/^{127}\mathrm{I})_T = (^{127}\mathrm{I})_{BSE}(^{129}\mathrm{I}/^{127}\mathrm{I})_{T0}\exp[-\lambda_{129}t]$ where $t = T_0 - T$ with $T_0 = 4.56$ Ga. Hence, the expression for ^{129}Xe* becomes

$$^{129}\mathrm{Xe}^* = (^{127}\mathrm{I})_{BSE}F_{129}(^{129}\mathrm{I}/^{127}\mathrm{I})_{T0}\exp[-\lambda_{129}t]$$

.

For simplicity, $(^{127}\mathrm{I})_{BSE}(^{129}\mathrm{I}/^{127}\mathrm{I})_{T0}$ will be referred to as $(^{129}\mathrm{I})_{T0}$ even though at T_0 the bulk silicate earth might not have formed yet. The above equation provides a constraint on the mean Xe-closure age (T) of Earth. Similarly, four equations can be obtained from constraints provided by the ^{244}Pu-^{238}U-^{136}Xe*-^{134}Xe*-^{132}Xe*-^{131}Xe* system (Table 2). Among the four equations, the equations involving ^{244}Pu-^{238}U-^{136}Xe* and ^{244}Pu-^{238}U-^{134}Xe* provides the main constraint and others do not provide strong additional constraint. Nevertheless, for consistency and symmetry, all the equations are included in constraining the Xe-closure age of Earth (Table 2). Best estimates with uncertainties (somewhat subjective) for all parameters in these equations are also listed in Table 2, except for the amount of radiogenic atmospheric xenon isotopes, which are listed in Table 1.

Table 2 also lists the fractions (F) of a radiogenic nuclide that is in the atmosphere out of the total radiogenic nuclide in the atmosphere plus crust plus mantle. These fractions are estimated by considering the amount of ^{40}Ar and ^{21}Ne* in the atmosphere and the total production in the BSE (e.g., Ozima and Podosek, 1983) as follows. All ^{40}Ar is essentially radiogenic from ^{40}K. The amount of ^{40}Ar in the contemporary atmosphere is 1.64×10^{18} mol (Ozima and Podosek, 1983). Using the new and slightly different concentration of K of 240 ± 50 ppm in BSE (McDonough and Sun, 1995; see also Allegre et al., 1995b; Zindler and Hart, 1986), total ^{40}Ar produced in the BSE over the entire history of Earth is $(3.3 \pm 0.7)10^{18}$ mol. Therefore, 1.64/3.3 = (50 ± 10)% of the total radiogenic ^{40}Ar in BSE is in the atmosphere ($F_{40} = 0.5 \pm 0.1$) (see also Ozima and Podosek, 1983; Sarda et al., 1985; Allegre et al., 1986/87; Zhang and Zindler, 1989). Using the same approach, ^{21}Ne* in the atmosphere and ^{21}Ne* production by ^{235}U, ^{238}U, and ^{232}Th in the BSE can be treated. ^{21}Ne* does not provide a strong constraint due to the large uncertainty in estimating ^{21}Ne* in the atmosphere, but is consistent with F = 0.5. Therefore, for radiogenic noble gas nuclides whose parents have long half lives (^{40}K, ^{235}U, ^{238}U, and ^{232}Th), F = 0.5 ± 0.1. Defining ^{136}Xe*^{2}, ^{134}Xe*^{2}, ^{132}Xe*^{2}, and ^{131}Xe*^{2} to be the radiogenic Xe from fission of ^{238}U (Table 2), $F_{136Xe*2}$, $F_{134Xe*2}$, $F_{132Xe*2}$, and $F_{131Xe*2}$ are hence 0.5 ± 0.1.

The fraction of a radiogenic gas that is in the atmosphere over total production in BSE (F) is different from (1) the mass fraction of degassed mantle over BSE or (2) the degree of degassing of the degassed mantle. The F value is simply the radiogenic gas in the atmosphere divided by total gas production in BSE and is model independent. The F value is not affected by (1) whether or not the degassing is controlled by solubility and (2) whether there was recycling of Xe back to the degassed mantle, as long as the degree of degassing from the degassing mantle is high. Only the concentrations in the degassed mantle depend on such details. Estimation of the mass fraction of degassed mantle over BSE using gas budget and total gas production must consider radiogenic gas in the continental crust and in the degassed mantle (Zhang and Zindler, 1989), in addition to the atmosphere. Estimation of the degree of degassing of the degassed

Table 2. Equations and parameters used for total inversion to obtain the Xe-closure age

The equations:

$$^{129}Xe^* = (^{127}I)_{BSE}F_{129}(^{129}I/^{127}I)_{T0} \exp[-\lambda_{129}t] \quad (1)$$

$$^{136}Xe^* = Y_{136}(^{244}Pu)_{T0}F_{136Xe^*1} \exp[-\lambda_{244}t] + X_{136}(^{238}U)_{T0}F_{136Xe^*2} \exp[-\lambda_{238}t] \quad (2)$$

$$^{134}Xe^* = R_{134}Y_{136}(^{244}Pu)_{T0}F_{134Xe^*1} \exp[-\lambda_{244}t] + Q_{134}X_{136}(^{238}U)_{t0}F_{134Xe^*2} \exp[-\lambda_{238}t] \quad (3)$$

$$^{132}xe^* = R_{132}Y_{136}(^{244}Pu)_{T0}F_{132Xe^*1} \exp[-\lambda_{244}t] + Q_{132}X_{136}(^{238}U)_{T0}F_{132Xe^*2} \exp[-\lambda_{238}t] \quad (4)$$

$$^{131}Xe^* = R_{131}Y_{136}(^{244}Pu)_{T0}F_{131Xe^*1} \exp[-\lambda_{244}t] + Q_{131}X_{136}(^{238}U)_{T0}F_{131Xe^*2} \exp[-\lambda_{238}t] \quad (5)$$

$$F_{129} - F_{40} = 0.10 \pm 0.05 \quad (6)$$

where $t = T_0$-T, $(^{244}Pu)_{T0}$ and $(^{238}U)_{T0}$ have similar meaning as $(129I)_{To}$, Y_{136} is the fraction of decayed ^{244}pu that goes to ^{136}Xe, X_{136} is the fraction of decayed ^{238}U that goes to ^{136}Xe, $F_{136_{Xe^*1}}$ is the fraction of ^{136}Xe* that is in the atmosphere over total ^{136}Xe* production in BSE by ^{244}Pu, and $F_{136_{Xe^*2}}$ is the fraction of ^{136}Xe* that is in the atmosphere over total ^{136}Xe* production in BSE by ^{238}U, R_{134} is the yield of ^{134}Xe from ^{244}Pu normalized to the yield of ^{136}Xe, and Q_{134} is the yield of ^{134}Xe from ^{238}U normalized to the yield of ^{136}Xe.

The parameters:

$\lambda_{129} = 44.15 \times 10^{-9}$ yr.$^{-1}$; $\lambda_{244} = 8.664 \times 10^{-9}$ yr.$^{-1}$; $\lambda_{238} = 0.155125 \times 10^{-9}$ yr.$^{-1}$

$X_{136} = (3.43 \pm 0.22) \times 10^{-8}$, $Y_{136} = (7.00 \pm 0.75) \times 10^{-5}$ (ref. 1)

$R_{134} = 0.939 \pm 0.008$, $Q_{134} = 0.832 \pm 0.012$ (ref. 1)

$R_{132} = 0.885 \pm 0.030$, $Q_{132} = 0.595 \pm 0.017$ (ref. 1)

$R_{131} = 0.246 \pm 0.020$, $Q_{131} = 0.076 \pm 0.003$ (ref. 1)

^{127}I concentration in BSE = $10 \overset{\times}{\div} 3$ ppb (ref. 2)

^{127}I in BSE = $(3.23 \overset{\div}{\times} 3) \times 10^{17}$ mol

$^{129}I/^{127}I$ at T_0 (Bjurbole) = $(1.10 \pm 0.03) \times 10^{-4}$ (ref. 3)

$(^{129}I)_{T0} = (3.55 \overset{\times}{\div} 3) \times 10^{13}$ mol

U concentration in present BSE = 20.3 ± 4.0 ppb (ref. 2)

$(^{238}U)_{T0} = (7.0 \pm 1.4) \times 10^{17}$ mol

$^{244}Pu/^{238}U$ at T_0 (Bjurbole) = 0.0068 ± 0.0010 (ref. 4)

$(^{244}Pu)_{T0} = (4.8 \pm 1.4) \times 10^{15}$ mol

$F_{136Xe^*2} = F_{134Xe^*2} = F_{132Xe^*2} = F_{131Xe^*2} = F_{40} = 0.5 \pm 0.1$

$F_{129} = F_{136Xe^*1} = F_{134eXe^*1} = F_{132Xe^*1} = F_{131Xe^*1} = 0.6 \pm 0.1$.

References: (1) Ozima and Podosek, 1983; (2) McDonough and Sun, 1995; (3) Swindle and Podosek, 1988; (4) Hudson et al., 1989 (and references therein).

mantle must exclude gas production in the undegassed mantle (Zhang and Zindler, 1989). Both of the latter two fractions are dependent on isotopic ratios in the mantle and somewhat model-dependent. The F value here is smaller than the mass fraction of the degassed mantle out of the total BSE because the degassed mantle still contains some ^{40}Ar due to continuous production and because some ^{40}Ar is in the continental crust. Hence, the fraction of ^{36}Ar that has been degassed to the atmosphere from the BSE is greater than F_{40} because ^{36}Ar is not continuously produced. The fraction of ^{36}Ar that has been degassed into the atmosphere from the BSE is similar to the mass fraction of the degassed mantle but different from the degree of degassing of the degassed mantle, which is >99% (Allegre et al., 1986/87; Zhang and Zindler, 1989).

For a radiogenic noble gas nuclide whose parent has a very short half life, the fraction that has been degassed to the atmosphere from the BSE is similar to that of a stable nonradiogenic nuclide and larger than that of ^{40}Ar. Based on a variety of considerations (e.g., Allegre et al., 1986/87; Zhang and Zindler, 1989), F_{129}, F_{136Xe^*1}, F_{134Xe^*1}, F_{132Xe^*1}, and F_{131Xe^*1}, are assumed to be 0.6 ± 0.1, where $^{136}Xe^{*1}$, $^{134}Xe^{*1}$, $^{132}Xe^{*1}$, and $^{131}Xe^{*1}$ are from fission of ^{244}Pu (Table 2). The choice of the exact F_{129}, F_{136Xe^*1}, F_{134Xe^*1}, F_{132Xe^*1}, and F_{131Xe^*1} values has only a small effect on the results presented below.

3. RESULTS

The calculated amounts of radiogenic ^{129}Xe and ^{136}Xe in the atmosphere using equations and parameters in Table 2 as a function of the Xe-closure age (T) of Earth are compared with the observed amounts in Fig. 1 . Figure 1a shows that with all independent estimates given in Tables 1 and 2, the Xe-retention age of Earth is 4.43 to 4.46 Ga, depending on which xenon isotope is considered. The small difference can be eliminated by adjusting the input parameters within the uncertainties. Figure 1b shows that a Xe-retention age of Earth that can reconcile available data for ^{129}Xe* and ^{136}Xe* is 4.45 Ga. Even though the Xe-retention age of Earth is determined using a more quantitative approach (below), Fig. 1b shows qualitative how trade-offs are made in the quantitative approach.

In order to use all the constraints from ^{129}Xe*, ^{136}Xe*, ^{134}Xe*, ^{132}Xe*, and ^{131}Xe* in a quantitative manner accounting for uncertainties in all input parameters, a total inversion technique (Tarantola and Valette, 1982; Zhang and Zindler, 1989) has been used to obtain the Xe-retention age of Earth using the input parameters in Tables 1 and 2. This age is found to be younger than Bjurbole by 109 ± 23 million years. Hence the Xe-retention age of Earth is 4.45 ± 0.02 Ga, consistent with the result of Wetherill (1975), and Allegre et al. (1995a), but different from the result of Staudacher and Allegre (1982). This age is independent of the input of the age estimate and almost independent of the initial estimate of F_{129}. For example, if a priori estimate of F_{129} is 0.7 ± 0.1 (i.e., if $F_{129} - F_{40} = 0.20+0.05$), t would be 114 ± 23 instead of 109 ± 23 million years. The total inversion results show that all the input parameters and the equations are consistent. That is, the output values of the parameters that satisfy all equations are within the initial specified bounds. The high internal consistency between the ^{129}I-^{129}Xe clock and ^{244}Pu-^{238}U-^{136}Xe-^{134}Xe-^{132}Xe-^{131}Xe clock (although input data are from a variety of sources and methods) shows that the volatility of iodine does not affect the ^{129}I-^{129}Xe clock, contrary to the claim of Azbel and Tolstikhin (1993). This result is understandable if iodine is closed before or at the same time of xenon closure. The consistency also shows that all radiogenic isotopes (including ^{40}Ar*, ^{21}Ne*, and all radiogenic xenon isotopes) in the atmosphere can be well accounted for by radiogenic growth and mantle degassing.

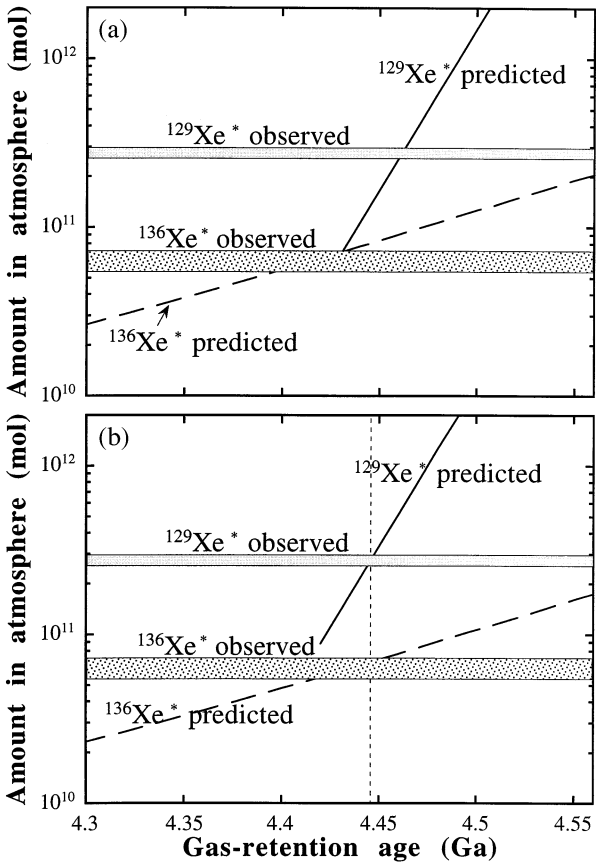

Fig. 1. Calculated amount of radiogenic ^{129}Xe* (solid line) and ^{136}Xe* (dashed line) in the atmosphere as a function of the Xe-closure age of Earth, compared with the observed amount of ^{129}Xe* and ^{136}Xe* in the atmosphere (shaded fields). The intersection between the calculated line and the observed field is a solution of the mean Xe-closure age. Uncertainties in calculated ^{129}Xe* and ^{136}Xe* are large and not shown. (a) All the input parameters are directly taken from Table 2. The Xe-closure age from ^{129}Xe* and ^{136}Xe* are different but the difference is not large. (b) The input parameters are adjusted within their uncertainties so that the closure age from both ^{129}Xe* and ^{136}Xe* are the same (vertical dashed line). This age is just one possible age (and slightly different from the age from total inversion) without quantitative considerations of all constraints and uncertainties.

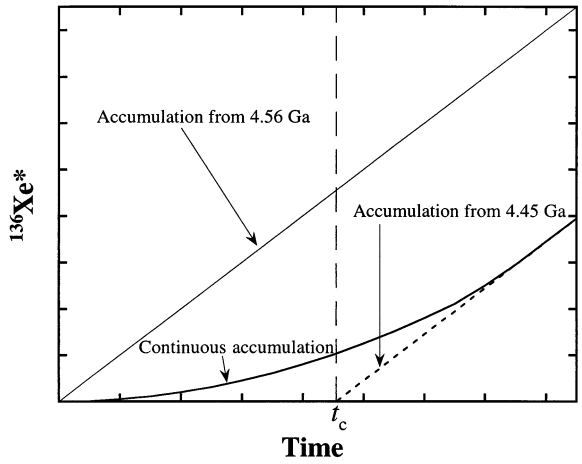

Fig. 2. A schematic drawing of the growth of radiogenic ^{136}Xe* in the atmosphere as a function of time and the concept of mean Xe-closure age. The time scale is not linear and is such that ^{136}Xe* accumulation in the atmosphere vs. this nonlinear time scale follows a straight line. The evolution curve marked continuous accumulation shows a case for continuous ^{136}Xe* accumulation and loss over the first 200 million years of Earth's history. In this case, the Xe-closure time is the extrapolated t_c. Atmospheric ^{136}Xe* might also have been completely lost at 4.45 Ga due to the impact of a Martian size planetesimal. Then the Xe-closure time records this event. One may also construct other models with episodic Xe loss to satisfy the data.

4. DISCUSSION

The meaning of the Xe-closure age of Earth inferred from I-Pu-U-Xe system is similar to that of the Ar-closure age of a mineral (Dodson, 1973). Figure 2 shows schematically the meaning of the age. For example, this mean Xe closure age can be accounted for by continuous accumulation and loss of atmospheric Xe over the first ~200 million years of Earth's history. Such a continuous loss could be due to either hydrodynamic escape or many giant impacts by planetesimals. This age may also represent the time when the last Martian-sized planetesimal impacted Earth, stripping Earth's protoatmosphere and rehomogenizing Earth. This last giant impact may or may not have resulted in the formation of the moon. In this scenario, the accretion of Earth roughly ended at 4.45 Ga, and there was only much slower mass addition (no giant event that could cause Xe loss) after this event. Besides these two end-member scenarios (continuous loss and one single event at 4.45 Ga) shown in Fig. 2, one may also construct more complicated Xe loss patterns that result in a closure age of 4.45 ± 0.02 Ga. For example, in an inhomogeneous accretion model, the Xe-retention age is consistent with the following scenario: ^{244}Pu-generated Xe was lost completely in the first 100 million years of proto-Earth's history. At ~4.45 Ga, about 5% volatile-rich mass with the present-day I-Xe ratio but with negligible radiogenic Xe was added to Earth, accounting for the present-day volatile budget. To choose between these choices requires other lines of evidence, as discussed below.

In obtaining the Xe-retention age, radiogenic xenon isotopes at T was assumed to be negligible. This assumption is reasonable in at least two scenarios: (1) all Xe (radiogenic and nonradiogenic) in the proto-Earth was lost at T (and hence all the present-day Xe came to Earth in a volatile-rich layer containing no significant radiogenic Xe), or (2) prior to T, Xe concentration in proto-Earth was very high (compared to the present-day Earth), hence contribution of radiogenic Xe did not significantly affect the xenon isotopic ratios. If the assumption is relaxed, the Xe-retention age would be younger.

The gas-retention age of Earth, younger by ~110 million years than Bjurbole, are consistent with several other age estimates for Earth. Allegre et al. (1995a), among others (e.g., Wetherill, 1975; Doe and Zartman, 1979; Staudacher and Allegre, 1982), showed that the Pb model and geochron age, interpreted as the core formation age of Earth, is 4.45 ± 0.03 Ga. Lee and Halliday (1995) and Halliday et al. (1996) studied the ^{182}Hf-^{182}W system and showed that the core formation age of Earth is <4.50 Ga. The interpretation of the ^{182}Hf-^{182}W data by Jacobsen and Harper (1996) leads to a mean age for core formation of Earth to be <4.55 Ga. Since the ^{182}Hf-^{182}W

system only provides a limit, all the model results of the ^{182}Hf-^{182}W system are consistent with the gas-retention age above, and I will not delve into the difference. Using the coupled ^{146}Sm -^{142}Nd and ^{147}Sm-^{143}Nd systems, Harper and Jacobsen (1992) and Jacobsen and Harper (1996) inferred an age for mantle differentiation of ~4.47 Ga (with an unconstrained uncertainty) although there are some controversies about the data (e.g., Sharma et al., 1996a; Jacobsen and Harper, 1996; Sharma et al., 1996b). Therefore, all the presently available data are consistent with a 4.45 ± 0.02 Ga age of Earth. Although there is uncertainty associated with every method of age estimation, and although some estimates are more model-dependent than others, the consistency of these estimations of the ages is significant. Hence, I suggest that the young age of Earth is now firmly established. Furthermore, the similarity between the Xe-closure age, the Pb model and geochron age, and the Sm-Nd age suggests a common cause that re-initialized all clocks (including Xe accumulation, core-mantle segregation, and mantle-crust segregation), probably the last Martian-size giant impact that stripped Earth's protoatmosphere and rehomogenized Earth (Wetherill, 1985).

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