Pennsylvania State University
The Graduate School

CONSTRAINTS ON EARTH’S THERMAL EVOLUTION FROM THE HEAVY NOBLE GAS CONTENT OF THE MANTLE

A Thesis in
Geosciences

by
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Abstract

The presence of liquid water on a planet over geological timescales is critical for its potential to harbor life. On the Earth, surface water is cycled from the atmosphere into the lithosphere, subducted into the mantle and degassed back to the atmosphere. However, some of the subducted water is retained in the mantle as determined by the thermal and metamorphic conditions of the subduction zone. Therefore, in order to understand the planet’s surface water over time, it is important to characterize Earth’s thermal subduction history. The Earth is currently in a cold subduction thermal regime, however, measurements and models suggest that hot subduction dominated in the past. Here we constrain when the switch from hot to cold subduction occurred in Earth history based on present day noble gas abundances of the mantle. Noble gases are retained in a subducting slab until certain conditions are reached, at which point they are lost from the slab. However, these gases experience an elemental fractionation effect in hot subduction zones but remain unfractionated in cold subduction zones. This affects the mantle noble gas input ratios and thus the mantle noble gas content as a whole, indicating that noble gases can be used as tracers to determine the planet’s thermal subduction history.

We model the cycling of noble gases between the surface and mantle coupled with the mantle’s thermal evolution and compare our results to modern mantle noble gas observations to answer this question. We consider the classic case of mantle convection where plate tectonics is active throughout Earth history and the heat flux scales with the Rayleigh number to the 1/3 power. We first find the mantle steady states for hot and cold subduction, then find the time frame for the mantle to reach steady state (it’s e-folding time). Finally, we run the full model to constrain the cold subduction onset time with parameters adjusted to push it as far back as possible. We find the modern MORB mantle noble gas ratio is closer to the hot steady state than the cold steady state, and is therefore not in equilibrium with the planet’s modern, cold subduction regime. The e-folding time of the mantle is approximately 1 Ga, indicating the switch from hot to cold subduction occurred in the recent geologic past. Our full model results indicate a transition time between 500 Ma to 1 Ga ago at the very earliest, during the Neoproterozoic.
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1 Introduction

A longstanding question in geosciences is how Earth has maintained a relatively stable ocean mass throughout its history, despite regassing of water at subduction zones and outgassing and volcanic arcs, hot spots, and mid-ocean ridges. Oceanic plates act as vectors for large quantities of water as they subduct, with water being stored in sediments (both within pore spaces and chemically bound in minerals), altered oceanic crust and serpentinized lithosphere (Parai & Mukhopadhyay, 2012). As pressures and temperatures increase within the subducting slab, hydrous minerals in the sediments, ocean crust, and lithospheric mantle break down, releasing water into the mantle wedge (Kerrick & Connolly, 2001; van Keken et al., 2011). In this way, a portion of the subducted water is returned to the surface, rather than being recycled into the deep mantle. However, the mantle itself is at least partially hydrated, suggesting that not all water in a slab is degassed back to the atmosphere. Estimates for mantle water capacity range from approximately one ocean in the upper mantle (Fei et al., 2017), up to three oceans in the transition zone (Nishi et al., 2014; Schmandt et al., 2014) and possibly four oceans in the lower mantle (Peslier et al., 2017). Thus, understanding the exchange of water between the surface and the mantle over geologic time is critical for understanding the long term habitability of the Earth and terrestrial exoplanets in their respective habitable zones.

To understand the history of surface water on the Earth, we must understand the history of deep water cycling on the planet. Current work suggests the hotter mantle of the early Earth was primarily degassing water from the mantle to the surface until at least 2 - 2.5 Ga (Rüpke et al., 2004) and any mantle regassing during this time was limited (Magni et al., 2014). This was expanded upon by Parai & Mukhopadhyay (2018), where they calculated the Earth was in a net degassing state up to 2 Ga and later switched to a net regassing state within 1 Ga. This indicates an overall change in how hydrated mantle downwellings and upwellings were during the early Earth in comparison to today. As mantle downwellings are the equivalent of subduction zones, it is clear the process of subduction is a critical component in the water cycle and therefore we investigate the metamorphic and thermal environments of subduction zones and how they have changed over Earth history. Our work contributes a new, independent constraint on subduction zone thermal regime over time to compliment these prior studies.

In this paper we use the noble gas abundances of the modern mantle to constrain when the average thermal regime of subduction on
Earth transitioned from hot to cold. Cold subduction is where relatively low temperatures are retained at depth in a subducting slab, whereas in hot subduction zones, significant amounts of mantle heat conducts into the slab at shallow depths (van Keken et al., 2011; Syracuse et al., 2008; Smye et al., 2017). As a result, cold subduction zones remain hydrated at great depths, whereas hot subduction zones dehydrate at comparatively shallow depths. The modern Earth is primarily in a state of cold subduction (Syracuse et al., 2008; van Keken et al., 2011), but Earth’s mantle was hotter in the past (Korenaga, 2006; Herzberg et al., 2010). This has consequential effects on the average subduction zone thermal regime. Hot subduction likely prevailed early in Earth’s history due to the higher mantle temperature (Magni et al., 2014); the early Earth likely had relatively dry mantle downwelling compared to today (Parai & Mukhopadhyay, 2018) indicating dehydrated subduction zones and therefore suggesting an overall hot subduction regime (van Keken et al., 2011). In addition, the lack of high pressure, low temperature metamorphic rocks until 1 Ga also indicates that subduction thermal regime on the early Earth was hot (Brown, 2014). Moreover, our results will show that the modern noble gas ratios of the mantle can only be explained if the Earth started in a state of hot subduction, and later transitioned to cold subduction as the mantle cooled (Sec. 3.1).

In this paper we investigate the timing of this transition from hot to cold subduction using the ratios of $^{84}\text{Kr}/^{36}\text{Ar}$ and $^{130}\text{Xe}/^{36}\text{Ar}$ in the mantle. We use these gases as they are non-radiogenic, their contents are well constrained in the atmosphere, crust and mantle and they undergo similar reactions to water in a subducting slab (as they are chemically inert and therefore only subject to physical processes). In hot subduction zones, noble gases are fractionated, where $^{36}\text{Ar}$ is preferentially lost from the subducting slab relative to $^{84}\text{Kr}$ and $^{130}\text{Xe}$, whereas in cold subduction zones, no fractionation occurs and the relative abundances of all three gas species will remain constant in the slab (Smye et al., 2017). Therefore, the relative inputs of these gases into the mantle, and thus the mantle ratios, depends on the subduction history of the Earth and can be modeled. It is important to note that while water hydrates the crust and possibly the lithospheric mantle, we will be focusing on these noble gases in the subducting crust only. This model will predict the ratios of these noble gases in the mantle at varying times of hot-cold subduction regime transition. These predictions will then be compared to the measured noble gas rations, taken from the Bravo Dome gas field in New Mexico, described by Holland & Ballentine (2006).
2 Methods

2.1 Noble Gas Cycling

We model the evolution of $^{36}$Ar, $^{84}$Kr and $^{130}$Xe content in the mantle over deep time in order to constrain the evolution of $^{84}$Kr/$^{36}$Ar and $^{130}$Xe/$^{36}$Ar ratios. To do this, we simulate the cycling of these gases between the mantle, atmosphere and crust. Noble gases in the oceanic crust are cycled into the mantle at subduction zones and returned to the atmosphere through volcanic outgassing at mid-ocean ridges. The rate of change of the mantle content of a particular noble gas is thus equal to the difference between the input rate at subduction zones and output rate by volcanic outgassing. Therefore the change in mantle content of $^{36}$Ar, $^{84}$Kr and $^{130}$Xe over time is as follows:

$$\frac{dC_M}{dt} = NC_c F_{sub} - C_M F_{ridge}. \quad (1)$$

where $N$ is the noble gas retention factor, described in section 2.3. $C_M$, and $C_c$ are the total noble gas contents, in moles, of the mantle and crust, respectively, $C_c F_{sub}$ is the rate at which gases enter the mantle at subduction zones and $C_M F_{ridge}$ is the rate at which gases leave the mantle through spreading ridges, where $F_{sub}$ and $F_{ridge}$ are defined as:

$$F_{sub} = \frac{v L_{trench}}{A_{sf}} \quad (2)$$

$$F_{ridge} = \frac{2v L_{ridge} d_m}{V_m}. \quad (3)$$

$L_{ridge}$ is the total length of all spreading ridges, $d_m$ is the average depth where partial melting begins below the ridges, $v$ is the global average plate velocity, $L_{trench}$ is the total length of all subduction zones, $V_m$ is the volume of the mantle and $A_{sf}$ is the total area of the seafloor.

We assume the loss of noble gases to space is negligible. This may not be an accurate assumption, as Xe may have coupled with hydrogen during hydrogen escape during the early Earth (Zahnle et al., 2019). However, this Xe loss ends when atmospheric oxygen rises, and our results are not affected by processes that occur in the first half of Earth history, as described in Section 4. Therefore, our assumption of a closed system where the total noble gas content of the Earth is conserved is reasonable. Thus

$$C_{tot} = C_A + C_c + C_M, \quad (4)$$
where $C_A$ is the noble gas contents of the atmosphere and ocean combined, and $C_{tot}$ is the total noble gas content of the mantle and surface reservoirs. We assume that the noble gas content of the crust is determined by the abundance in the atmosphere. To characterize the relationship we simply assume that the ratio of $C_c/C_A$ is fixed through time. We define a partition coefficient, $K$, based on the modern day noble gas contents of the crust and atmosphere:

$$K = C_{co}/C_{Ao}$$

where $C_{co}$ and $C_{Ao}$ are the modern crustal and atmospheric noble gas contents, respectively.

The crustal noble gas content is then given by:

$$C_c = C_A K$$

and the atmospheric noble gas content is calculated by combining Eq. 4 and 6:

$$C_A = (C_{tot} - C_M)/(1 + K).$$

Now, Eq. 6 and 7 allow us to calculate $C_A$ and $C_c$.

The modern day crustal contents are calculated using known concentrations of these gases in the oceanic crust:

$$C_{co,j} = n_j A_{sf} d_{crust} \rho_{basalt}$$

where $C_{co,j}$ is the reference crustal content for noble gas $j$, $d_{crust}$ is the crustal thickness, $\rho_{basalt}$ is the density of basalt and $n_j$ is the crustal concentration of the noble gas species $j$, measured in mol/g (moles of gas per gram of crustal rock). The concentrations of each noble gas are calculated from the geochemical data reported by Smye et al. (2017). We specified three distinct sections of the crust: Sediments, extrusive and intrusive basalt. The concentration of the total crustal package is determined by a weighted average of the concentrations of the three sections, assuming 1 km of sediments, 3 km of extrusive basalt and 4 km of intrusive gabbro. The total amount of each noble gas in our system, $C_{tot}$, is then defined by the sum of the reference mantle, crustal, and atmospheric values, and the partition coefficient, $K$, can be calculated from Eq. 5 using the known $C_{Ao}$ and $C_{co}$. The values are listed in Table 1.

### 2.2 Thermal Evolution

Both $F_{sub}$ and $F_{ridge}$ depend on the average plate velocity $v$. As the average plate velocity is a function of the vigor of mantle convection,
we must model the thermal evolution of the mantle in order to capture the time evolution of plate velocity, and hence noble gas cycling. We assume pure internal heating as internal heat sources account for the majority of present day mantle heat loss (Jaupart et al., 2007), when heat production by radioactive isotopes and secular cooling of the mantle are combined. In addition, neglecting heat transfer from the core to the mantle greatly simplifies the model. Accounting for core cooling would require a core evolution model coupled to that of the mantle (Stevenson et al., 1983). Moreover, the plate velocity and temperature time evolution is, to first order, unchanged with the inclusion of core cooling (Davies, 2007), so our simplification of ignoring core heat flux is unlikely to significantly change the results. The effects of heat transport by melts or hydrothermal fluids is also ignored, as it is a negligible component of Earth’s present day heat budget (Jaupart et al., 2007; Driscoll & Bercovici, 2014), though heat transport via mantle melting may have been significant during the Hadean or Eoarchean (Nakagawa & Tackley, 2012; Moore & Webb, 2013). With these assumptions, the temperature evolution of the mantle is given by:
\[
\frac{dT}{dt} = \frac{M_m H_m - Q_m}{\chi_m M_m C_m}
\]

where \(M_m\) and \(C_m\) are the mass and specific heat of the mantle, respectively, \(H_m\) is the heat production rate in the mantle, \(\chi_m\) is the upper mantle to mean mantle potential temperature ratio (we assume \(\chi_m = 1\)) and \(Q_m\) is the heat loss out of the mantle through the lithosphere.

Mantle heat loss is

\[
Q_m = 4\pi R_e^2 q
\]

where \(q\) is heat flux through the lithosphere and \(R_e\) is the radius of the Earth. The heat flux out of the mantle is determined by a scaling law for heat flux of a viscous, convecting fluid. Numerical models, boundary layer theory and laboratory experiments indicate a scaling law of the form (Howard, 1966; Solomatov, 1995; Schmalzl et al., 1996; Turcotte & Schubert, 2002):

\[
q = \frac{a k \Delta T}{D} \left( \frac{Ra_i}{Ra_c} \right)^{\beta}.
\]

Here, \(a\) is an empirically determined constant, \(k\) is thermal conductivity, \(D\) is the thickness of the mantle, \(\beta\) is a scaling exponent discussed below, \(Ra_i\) is the internal Rayleigh number and \(Ra_c\) is the critical Rayleigh number. The critical Rayleigh number is the threshold value that the internal Rayleigh number must reach in order for the fluid to convect, and is on the order of 1000 (Strutt, John William (Lord Rayleigh), 1916; Turcotte & Schubert, 2002; Chandrasekhar, 2013). The internal Rayleigh number itself is:

\[
Ra_i = \frac{g \rho_m \alpha \Delta T D^3}{\kappa \mu_i}
\]

where \(g\) is the gravitational constant, \(\rho_m\) is the mantle density, \(\alpha\) is the thermal expansion coefficient of the mantle and \(\Delta T\) is the difference between the mantle potential temperature and the surface temperature.

The scaling factor \(\beta\) determines how strongly heat flux is linked to the vigor of convection. However, the value of \(\beta\) that best captures the complexities of convection in Earth’s mantle is debated. For a simple convecting fluid with constant material properties, \(\beta = 1/3\) (Howard, 1966; Turcotte & Oxburgh, 1967; Solomatov, 1995; Korenaga, 2006). However, the viscosity of Earth’s mantle varies by many orders of magnitude due to variations in temperature, grain size, stress, and water content (Karato & Wu, 1993; Hirth & Kohlstedt,
which might alter this scaling factor. For example, studies have suggested that a temperature-dependent viscosity will result in lower $\beta$ values (Christensen, 1984). However, Solomatov (1995) showed that strongly temperature-dependent viscosity results in stagnant lid convection where $\beta = 1/3$. $\beta$ less than 1/3 only occurs in a “sluggish lid” convection regime, between the stagnant lid and constant viscosity convection regimes, where the temperature dependence of viscosity is only modest. However, whether plates and plate boundaries or the underlying mantle provide the dominant resistance to plate motions for the Earth is not well known, and thus $\beta$ values less than 1/3, including $\beta \approx 0$ or even negative, have been proposed on the basis that tectonic plates are intrinsically strong (Conrad & Hager, 1999; Korenaga, 2006). For our initial model, we assume a weak lithosphere where the primary resistance to motion is the mantle viscosity, and thus we use $\beta = 1/3$. However we also perform test cases for strong plate models, one where $\beta$ is effectively 0 and one where it is negative, and find no significant change to our overall results (see Section 3.4).

We also assume the mantle deforms by diffusion creep. While dislocation creep dominates the upper mantle, caused by the high shear stresses imparted by moving and subducting slabs, diffusion creep dominates under the lower stress regimes and smaller grain sizes of the lower mantle. As the lower mantle makes up the bulk of the mantle, it thus sets the average interior mantle viscosity that descending slabs feel, and we therefore assume it controls Earth’s thermal evolution (Karato & Li, 1992; Karato & Wu, 1993; Hirth & Kohlstedt, 1995). We calculate internal viscosity $\mu_i$ by

$$\mu_i(T) = \mu_0 \exp \left[ \frac{E}{R_g \left( \frac{1}{T} - \frac{1}{T_r} \right)} \right]$$

where $\mu_0$ is the viscosity at the reference temperature, $T_r$ (the current upper mantle potential temperature of 1300°C), $E$ is the activation energy for the mantle, and $R_g$ is the universal gas constant. Current estimates for $E$ are approximately 400 kJ/mol (Karato & Wu, 1993; Hirth & Kohlstedt, 2003). Plate velocity is calculated from the following scaling law, which is analogous to the heat flux scaling law (Eq. 11)(Solomatov, 1995):

$$v(T) = \frac{ck}{D} Ra^{2\beta}$$

where an empirically derived constant $c = 0.09$. Substituting in for the Rayleigh number from Eq. 12

$$v(T) = \frac{ck}{D} \left( \frac{\rho g \alpha \Delta T D^3}{\kappa \mu_i(T)} \right)^{2\beta}.$$
Table 2: Quantities used in calculations.

<table>
<thead>
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<th>Symbol</th>
<th>Value</th>
<th>Description</th>
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<tr>
<td>$M_m$</td>
<td>4x10^{24}$ kg</td>
<td>Mass of the mantle</td>
</tr>
<tr>
<td>$C_m$</td>
<td>1000 J kg^{-1}K^{-1}</td>
<td>Specific Heat of the mantle</td>
</tr>
<tr>
<td>$V_m$</td>
<td>9.0989x10^{20}$ m^3</td>
<td>Volume of the mantle</td>
</tr>
<tr>
<td>$R_e$</td>
<td>6371 km</td>
<td>Radius of the Earth</td>
</tr>
<tr>
<td>$D$</td>
<td>2890 km</td>
<td>Depth of the mantle</td>
</tr>
<tr>
<td>$\rho_m$</td>
<td>4000 kg m^{-3}</td>
<td>Density of the mantle</td>
</tr>
<tr>
<td>$\rho_{basalt}$</td>
<td>3000 kg m^{-3}</td>
<td>Density of basalt</td>
</tr>
<tr>
<td>$g$</td>
<td>9.8 m s^{-2}</td>
<td>Gravitational acceleration of the Earth</td>
</tr>
<tr>
<td>$E$</td>
<td>400 kJ mol^{-1}</td>
<td>Activation energy of the mantle</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>8.6x10^{-7}$ m^2 s^{-1}</td>
<td>Thermal diffusivity of the mantle</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>2x10^{-5}$ K</td>
<td>Thermal expansion of the mantle</td>
</tr>
<tr>
<td>$\mu_0$</td>
<td>10^{21}$ Pa s</td>
<td>Current viscosity of the mantle</td>
</tr>
<tr>
<td>$Ra_c$</td>
<td>1000</td>
<td>Critical Rayleigh number</td>
</tr>
<tr>
<td>$\beta$</td>
<td>1/3</td>
<td>Rayleigh scaling parameter</td>
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<td>$\alpha$</td>
<td>1.4</td>
<td>Empirically derived constant</td>
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<tr>
<td>$T_r$</td>
<td>1300 °C</td>
<td>Current upper mantle potential temperature</td>
</tr>
<tr>
<td>$T_s$</td>
<td>15 °C</td>
<td>Surface temperature</td>
</tr>
<tr>
<td>$R_g$</td>
<td>8.31 J mol^{-1}K^{-1}</td>
<td>Universal gas constant</td>
</tr>
<tr>
<td>$d_m$</td>
<td>70,000 m</td>
<td>Depth of melting beneath spreading ridge</td>
</tr>
<tr>
<td>$d_{crust}$</td>
<td>8000 m</td>
<td>Depth of the oceanic crust</td>
</tr>
<tr>
<td>$A_{sf}$</td>
<td>5.1x10^{14}$ m^2</td>
<td>Area of the seafloor</td>
</tr>
<tr>
<td>$L_{ridge}$</td>
<td>6x10^{7}$ m</td>
<td>Total length of spreading ridges</td>
</tr>
<tr>
<td>$L_{trench}$</td>
<td>6x10^{7}$ m</td>
<td>Total length of subduction zones</td>
</tr>
</tbody>
</table>

There are four radioactive isotopes within the Earth that power mantle convection. These isotopes are $^{235}$U, $^{238}$U, $^{232}$Th and $^{40}$K. We track the heat production in the mantle by the decay of these four radioisotopes (Turcotte & Schubert, 2002; Davies, 2007).

$$H_m(t) = \sum_i H_i X_i \exp \left( \frac{-t \ln 2}{\tau_i} \right)$$ (16)

where the index $i$ is summed over each radioactive isotope, $\tau_i$ is the half-life of isotope $i$, $H_i$ is the heat output and $X_i$ is the concentration of each isotope at Earth’s formation. Modern day concentrations are taken from Turcotte & Schubert (2002) and extrapolated back 4.5 Ga to determine $X_i$; these values can be found in Table 3.
<table>
<thead>
<tr>
<th>Isotope</th>
<th>$\tau$ (years)</th>
<th>$H_i$ (W kg$^{-1}$)</th>
<th>$X_i$ (kg kg$^{-1}$)</th>
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</thead>
<tbody>
<tr>
<td>$^{238}U$</td>
<td>4.47x10$^9$</td>
<td>9.46x10$^{-5}$</td>
<td>6.189x10$^{-8}$</td>
</tr>
<tr>
<td>$^{235}U$</td>
<td>7.04x10$^8$</td>
<td>5.69x10$^{-4}$</td>
<td>1.848x10$^{-8}$</td>
</tr>
<tr>
<td>$^{232}Th$</td>
<td>1.4x10$^{10}$</td>
<td>2.64x10$^{-5}$</td>
<td>1.55x10$^{-7}$</td>
</tr>
<tr>
<td>$^{40}K$</td>
<td>1.25x10$^9$</td>
<td>2.92x10$^{-5}$</td>
<td>4.447x10$^{-7}$</td>
</tr>
</tbody>
</table>

Table 3: Concentrations, heat production and half lives of the primary radioisotopes in the Earth. Concentrations are extrapolated to Earth’s formation (Turcotte & Schubert, 2002).

2.3 Noble Gas Fluxes and the Thermal State of the Subduction Zone

Noble gases are incorporated into hydrothermal minerals such as amphibole during fluid-rock interactions on the seafloor. As this oceanic crust subducts, the noble gases within will be subjected to rising temperatures and pressures. Some of these gases will be released from carrier minerals like amphibole, through diffusion and dissolution during dehydration. These gases are then returned to the atmosphere through arc volcanism, and thus avoid becoming part of the mantle reservoir. Due to the temperature dependence of noble gas diffusion through mineral lattices, the release of noble gases from the subducting slab is strongly influenced by the thermal regime of the subduction zone. In hot subduction zones, such as Cascadia, the lighter gases, such as Ar, are preferentially lost through slab dehydration in comparison to the heavier gases, such as Kr. These lighter gases are returned to the atmosphere through arc volcanism, whereas more of the heavier gases remain and become locked in as they descend, ultimately being incorporated into the mantle. Alternatively, in cold subduction zones such as Honshu, there is no preferential loss in Ar compared to Kr or Xe, and all are eventually locked in and retained in the mantle (Smye et al., 2017). However, the point at which the length scale of grain boundary gas transport becomes shorter than the width of the slab, preventing noble gases from being lost from the slab, is unclear. Accordingly, we make a simple depth assumption: All gases that escape the slab above a specified depth return to the atmosphere, and all that escape below this depth are retained in the mantle. We will call this the retention depth.

In Fig 1, we see fractionation in hot subduction zones occurring in the 55 – 75 km depth range. As the retention depth for the Earth is not known, we test a series of retention depths in this range, specifically 60, 65, and 70 km, encompassing the range of depths where fractionation
Figure 1: Mass fraction of each noble gas species that remains in the slab as a function of depth for cold (A), intermediate (B), and hot (C) subduction zones. We assume a retention depth, above which any gases released from the slab escape into the atmosphere through arc volcanism, and below which they are retained in the mantle. The retention depth range explored in this study (between 60 and 70 km) is highlighted. This chosen range shows the most significant light noble gas fractionation during hot subduction. Source: (Smye et al., 2017)

is the most significant and where amphibole breaks down. The fraction of a particular noble gas that remains in the slab at the retention depth is incorporated into the mantle. This fraction is characterized by introducing the retention factor $N$, first seen in Eq. 1, which ranges from 0 (all gases escape to the atmosphere) to 1 (all gases are retained in the mantle). $N$ values for each noble gas species at each retention depth during hot subduction are shown in Table 4. Intermediate and cold subduction zones have $N$ values of 1 for each noble gas at each retention depth.

<table>
<thead>
<tr>
<th>Noble Gas</th>
<th>60 km</th>
<th>65 km</th>
<th>70 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>$N_{Ar}$</td>
<td>0.5</td>
<td>0.3</td>
<td>0.2</td>
</tr>
<tr>
<td>$N_{Kr}$</td>
<td>1</td>
<td>0.95</td>
<td>0.9</td>
</tr>
<tr>
<td>$N_{Xe}$</td>
<td>1</td>
<td>0.6</td>
<td>0.2</td>
</tr>
</tbody>
</table>

Table 4: Hot subduction $N$-values for each noble gas at each of the three retention depths.
3 Results

3.1 Steady State Solutions

To gain an understanding of how the three reservoir system behaves, we first consider a simplified model that permits an analytical solution for the time evolution of mantle noble gas content. When a constant plate speed and fixed crustal noble gas reservoir are assumed, Eq. 1 can be integrated to yield

$$C_M(t) = NC_c \frac{F_{sub}}{F_{ridge}} \left[1 - e^{-F_{sub}t}\right].$$  \tag{17}

As \( t \to \infty \), we find the steady state:

$$C_M(t \to \infty) = NC_c \frac{F_{sub}}{F_{ridge}}$$  \tag{18}

Substituting in Eq. 3 and Eq. 2 and solving for \( C_c \) in terms of \( C_{tot} \) (see Eq. 4), we can solve for a steady-state solution entirely in terms of known input parameters:

$$C_{M,j} = C_{tot,j} \left(\frac{N_j K_j V_m}{2A_s f d_m(1 + K_j) + N_j K_j V_m}\right)$$  \tag{19}

where \( N \) and \( K \) are the fractionation factor and partition coefficient of a specific noble gas species \( j \), respectively. By taking the ratio of the predicted steady state mantle contents we find the steady state solutions in ratio space:

$$S_{Kr} = \frac{C_{tot,Kr}}{C_{tot,Ar}} \left[\frac{N_{Kr} K_{Kr}}{N_{Ar} K_{Ar}} \left(\frac{2A_s f d_m(1 + K_{Ar}) + N_{Ar} K_{Ar} V_m}{2A_s f d_m(1 + K_{Kr}) + N_{Kr} K_{Kr} V_m}\right)\right]$$  \tag{20}

$$S_{Xe} = \frac{C_{tot,Xe}}{C_{tot,Ar}} \left[\frac{N_{Xe} K_{Xe}}{N_{Ar} K_{Ar}} \left(\frac{2A_s f d_m(1 + K_{Ar}) + N_{Ar} K_{Ar} V_m}{2A_s f d_m(1 + K_{Xe}) + N_{Xe} K_{Xe} V_m}\right)\right]$$  \tag{21}

where \( S_{Kr} \) and \( S_{Xe} \) are the \(^{84}\text{Kr}/^{36}\text{Ar} \) and \(^{130}\text{Xe}/^{36}\text{Ar} \) steady state solution ratios, respectively. We plot these in ratio space along with the measured MORB mantle, seen in Fig 2. For hot subduction, \((N_{Ar} = 0.5, \text{all others are equal to 1})\), we find the steady state differs significantly from that of cold subduction (all \( N \) values equal 1). The modern MORB mantle is not at either the hot or cold subduction steady states, and is in fact closer to the hot subduction steady state.
than the cold. As the Earth is currently in a cold subduction thermal regime, we can see the modern mantle in not in equilibrium.

Our model uses modern reference values for noble gas abundances to set the partition coefficients, $K_j$, and the total amount of each gas that we keep conserved, $C_{tot}$. As discussed before, crustal concentrations are calculated by a weighted average of the oceanic crust data from Smye et al. (2017). Mantle contents are calculated from noble gas concentrations per $10^6$ Si atoms, presented by (Halliday, 2013), atmospheric contents are obtained from (Ozima & Podosek, 2002) and the uncertainty ranges are shown in Table 5. As these values come from direct experiments and observations, they have their own uncertainty ranges. Our calculated hot and cold subduction steady-states

Figure 2: Modern MORB value from (Holland & Ballentine, 2006) is plotted along with idealized hot and cold subduction steady states: Ellipses represent associated uncertainties derived from crustal reference value uncertainties (Table 5). Modern subduction, described in section 3.4.1, is plotted alongside idealized cold subduction.
depend on these input reference values, so we investigate the effects these uncertainty ranges have on the model results.

We find that adjusting the reference atmospheric and mantle noble gas values in our model within their respective uncertainty ranges produce negligible effects on model predictions. However, varying the crustal concentrations of each noble gas produces large differences in the calculated steady states. In Fig. 2, ellipses represent the uncertainty ranges of the hot and cold subduction steady states, due to uncertainties in the modern crustal concentrations of $^{36}$Ar, $^{84}$Kr and $^{130}$Xe.

<table>
<thead>
<tr>
<th>Element</th>
<th>Symbol</th>
<th>$n_{\text{min}}$ Range</th>
<th>$n_{\text{max}}$ Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Argon</td>
<td>$n_{\text{min}}$</td>
<td>$2.2928 \times 10^{-12}$ mol g$^{-1}$</td>
<td>$2.8678 \times 10^{-12}$ mol g$^{-1}$</td>
</tr>
<tr>
<td>Krypton</td>
<td>$n_{\text{min}}$</td>
<td>$7.9281 \times 10^{-14}$ mol g$^{-1}$</td>
<td>$9.5581 \times 10^{-14}$ mol g$^{-1}$</td>
</tr>
<tr>
<td>Xenon</td>
<td>$n_{\text{min}}$</td>
<td>$1.6731 \times 10^{-15}$ mol g$^{-1}$</td>
<td>$2.1111 \times 10^{-15}$ mol g$^{-1}$</td>
</tr>
</tbody>
</table>

Table 5: Ranges in noble gas crustal concentrations (Smye et al., 2017).

As an aside, the mantle steady state ratios can be compared to the overall crustal ratios. The crustal ratios are calculated as follows:

$$\left( \frac{K_r}{Ar} \right)_{\text{crust}} = \frac{N_{Kr}}{N_{Ar}} \left( \frac{C_{co,Kr}}{C_{co,Ar}} \right)$$  \hspace{1cm} (22)

$$\left( \frac{X_e}{Ar} \right)_{\text{crust}} = \frac{N_{Xe}}{N_{Ar}} \left( \frac{C_{co,Xe}}{C_{co,Ar}} \right).$$  \hspace{1cm} (23)

These ratios differ from the predicted mantle steady state (keep in mind that this is the cold subduction steady state, therefore $N_j = 1$ for each noble gas $j$). This difference lies in the extra factors of Eq. 20 and 21, primarily the differences in the partition coefficient $K$ between crust and atmosphere for the different noble gases.

The steady state results indicate mantle is transitioning from hot to cold steady states. Characteristic time scale to come to steady state then dictates how long ago the switch to cold subduction could have happened. This time scale is the e-folding time. From the general solution seen in Eq. 17, the e-folding time is set by $F_{\text{sub}}$ and is simply:

$$\tau = \left[ F_{\text{sub}} \right]^{-1}$$  \hspace{1cm} (24)
Assuming a fixed plate velocity, the e-folding time can be calculated (Fig. 3). The plate velocity scales with mantle potential temperature following the mantle convection scaling law given in Eq. 15. A decrease in temperature results in a lower plate velocity, not only from the decrease in $\Delta T$ between the mantle and the surface, but also from the increase in viscosity. Therefore, we can calculate plate velocity, and hence e-folding time, as a function of mantle potential temperature (Fig. 3).

The results of the e-folding time calculations are significant. Fig 3 shows that the modern mantle has an e-folding time of approximately 1 Gyrs. Higher mantle potential temperatures, which likely prevailed in the past, result in faster e-folding times. Thus the mantle reaches a steady state relatively quickly in response to a change in mantle input flux brought about by a change in subduction zone thermal
regime. Therefore, in order to explain the modern mantle $^{84}\text{Kr}/^{36}\text{Ar}$ and $^{130}\text{Xe}/^{36}\text{Ar}$ ratios, the switch from hot to cold subduction must be within the last 1 Gyrs.

### 3.2 Full Thermal Evolution Model Results

Steady state and characteristic timescale analysis gives us a rough guide for when the hot to cold subduction transition must have been, but we need a full thermal evolution model for a more precise estimate. We employ this model to confirm the results of the simple, analytical model and to provide more precise estimates of the transition from hot to cold subduction that explains the modern mantle $^{84}\text{Kr}/^{36}\text{Ar}$ and $^{130}\text{Xe}/^{36}\text{Ar}$ ratios. To do this, we manually impose a switch from hot to cold subduction at varying times, ranging from 0 Gyrs (the time of Earth’s formation, in which case cold subduction would have prevailed throughout Earth’s history) to 4.5 Gyrs (the present day age,
Figure 5: The calculated ratios of $^{130}\text{Xe}/^{36}\text{Ar}$ to $^{84}\text{Kr}/^{36}\text{Ar}$ for hot-cold transition times between 0 – 4.5 Ga with a retention depth of 60 km, color coded by the time of the imposed hot-cold switch. Also plotted is the current mantle MORB ratio (purple dot), with its uncertainties. Reference crustal concentrations used: $^{36}\text{Ar} = 2.2928 \times 10^{-12}$ mol g$^{-1}$, $^{84}\text{Kr} = 9.5581 \times 10^{-14}$ mol g$^{-1}$ and $^{130}\text{Xe} = 2.1111 \times 10^{-15}$ mol g$^{-1}$ in which case hot subduction would have prevailed throughout Earth’s history. This will allow us to place bounds on the transition time frame. The predicted results are then plotted and compared to the measured MORB mantle and shown in Fig 5. Predicted mantle values that fall within the error of the measured MORB mantle are considered acceptable models that have hot to cold subduction transition times consistent with the observed mantle $^{84}\text{Kr}/^{36}\text{Ar}$ and $^{130}\text{Xe}/^{36}\text{Ar}$ ratios.

We characterize plate velocity over Earth history by modeling the temperature evolution of the upper mantle. By solving Eq. 9, we find a temperature evolution shown in Fig 4. After an initial cooling period, the temperature tracks the rate of heat production from ra-
dioactive elements. We set the initial mantle potential temperature at 1800°C, which is consistent with upper mantle potential temperatures after magma ocean solidification (Abe, 1997). However, after an initial adjustment phase, mantle temperature follows the same time evolution regardless of the initial temperature, in this “classic” model with $\beta = 1/3$ (Davies, 2007). The effect of this result on plate velocity is substantial, as viscosity increases by two orders of magnitude between 1800°C and 1300°C and the plate velocity correspondingly decreases by two orders of magnitude. As e-folding time is dependent on plate speed, the effect of a cooling mantle on e-folding time is also significant (these effects are shown in Fig 3). Early on when the mantle was hot, the mantle noble gas ratios would have reached steady state relatively quickly. However, mantle temperatures through most of Earth history were comparable to today, with e-folding times on the order of 1 Gyr. Consequently, the onset of cold subduction must have occurred within the last 1 Ga.

The exact time of this transition varies depending on input parameters, primarily the modern crustal concentrations of noble gases. As these concentrations have uncertainties, there is a wide range of values that can be input, yielding differing transitions times. Since the analytical solution suggests a recent transition time, we therefore choose crustal concentrations in order to yield the earliest onset of cold subduction possible. This is achieved by assuming a minimum $^{36}$Ar value ($2.928 \times 10^{-12}$ mol g$^{-1}$) and maximum $^{84}$Kr and $^{130}$Xe values ($9.581 \times 10^{-14}$ mol g$^{-1}$ and $2.1111 \times 10^{-15}$ mol g$^{-1}$, respectively). Our model predicts $^{84}$Kr/$^{36}$Ar and $^{130}$Xe/$^{36}$Ar ratios in the modern day mantle as a function of the transition time from hot to cold subduction (Fig 5). We find a transition time of 3.5 to 4 Gyrs after Earth’s formation. Here we see that hot-cold subduction transition times during the first 3 Gyrs of Earth history predicts a modern mantle in a cold subduction steady state. This is consistent with e-folding times of 1 Gyr, as the mantle would have had plenty of time to reach an equilibrium state in such a case. Transition times of 3 Gyrs after formation or later produce noticeable deviations from the cold steady state, with ranges from 3.5 - 4 Gyrs after formation corresponding to the modern MORB mantle (the best fit being 3.7 Gyrs). This places the hot-cold transition time approximately 500 Ma to 1 Ga, during the Neoproterozoic. Varying the reference crustal noble gas abundances within their respective uncertainty ranges yield different results. However, those combinations that generate predictions consistent with the observed MORB mantle all push the cold subduction onset time closer to the present. For example, transition times of 200 to 400 Ma are consistent with the observed mantle noble gas ratios if midpoint noble
gas concentration reference values (midway between \( n_{\text{min}} \) and \( n_{\text{max}} \)) are chosen. In order to find an exact transition time, more precise measurements of crustal noble gas concentrations will be required. Nonetheless, the time frame for hot-cold subduction transition can be constrained to sometime in the last 1 Ga.

### 3.3 Retention Depths

The results depicted in Fig 5 are for a retention depth of 60 km. However, we also explored the consequences of 65 and 70 km depths. For consistency, these depths were also modeled using minimum \( ^{36}\text{Ar} \) and maximum \( ^{84}\text{Kr} \) and \( ^{130}\text{Xe} \) crustal concentrations. As can be seen in Fig 6, both the 60 and 65 km retention depths models are able to match, within error, the modern day mantle. Not only that, they both make similar predictions for the hot to cold subduction transition time frame, between 3.5 and 4 Gyrs after formation, though the 65 km depth does narrow the range to approximately 3.5 to 3.8 Gyrs. With a 70 km retention depth, however, none of our models can match the modern day mantle \(^{84}\text{Kr}/^{36}\text{Ar}\) and \(^{130}\text{Xe}/^{36}\text{Ar}\) ratios, as the modern mantle does not lie on a mixing line between the 70 km hot and cold subduction steady states. The loss of \(^{130}\text{Xe} \) in hot subduction zones is significant enough at deep retention depths that the loss in \(^{36}\text{Ar} \) cannot make up for it, preventing the \(^{130}\text{Xe}/^{36}\text{Ar}\) ratio from rising high enough to explain the modern mantle. At depths greater than 75 km, all noble gases are lost from the slab before they can enter the mantle in hot subduction zones. Therefore, noble gases will only be able to enter the mantle in cold subduction zones, where they are unfractionated. In this case, our model will also be unable to match the observed modern day mantle noble gas ratios. An shallow constraint on the retention depth is also possible. 50 km, for example, would involve an \( N \) value of 0.9 for \(^{36}\text{Ar} \), and \( N \) values of 1 for the other two gases. As both Fig 5 and Fig 6 are plots in ratio space with both \(^{130}\text{Xe} \) and \(^{84}\text{Kr} \) being ratioed to \(^{36}\text{Ar} \), we can easily predict where this new hot subduction steady state will lie. When \( N \) for \(^{36}\text{Ar} \) is 0.5, the value of each ratio is doubled, yielding the hot subduction zone steady state depicted in Fig. 5. With an \( N \) value of 0.9, the ratios will increase only by 10 percent. This will place the new, 50 km retention depth, hot subduction steady state closer to the cold steady state that the current MORB mantle. This suffers from the same fate as the 70 km depth, failing to predict the modern mantle.
3.4 Sensitivity Testing

3.4.1 Not Quite Cold

Modern day subduction zones range from hot, to intermediate, to cold. Cold and intermediate subduction zones are far more common than hot subduction zones (van Keken et al., 2011), therefore the majority of noble gas input into the mantle is unfractionated in the modern day. However, our model assumes that the current subduction regime of the Earth is entirely cold. Here we justify that assumption by comparing the difference between the actual modern day noble gas input flux and a purely cold input flux. To do this, we calculate an “effective” $^{84}\text{Kr}/^{36}\text{Ar}$ and $^{130}\text{Xe}/^{36}\text{Ar}$ steady state of the mantle for modern subduction, taking a global average of all subduction zones. As seen in Fig 1, cold subduction would indicate $N$ values of 1 for each noble gas species. Since some of the current subduction flux of

Figure 6: The calculated ratios of $^{130}\text{Xe}/^{36}\text{Ar}$ to $^{84}\text{Kr}/^{36}\text{Ar}$ for hot-cold transition times between 0 – 4.5 Ga, with retention depths of 60, 65 and 70 km.
noble gases occurs in hot subduction zones, we calculate effective $N$ values to properly capture degree of fractionation that occurs when integrating over all modern day subduction zones. The current input flux is determined by integrating the individual fluxes of every subduction zone. The flux of each zone is calculated by adjusting Eq. 2 in the following way:

\[ F' = \frac{NC_c'v'L_{trench}'}{A_{sf}} \]  

(25)

where $F'$ is the subduction flux of any specific subduction zone. Only $A_{sf}$ and $C_c$ are the same as in our full model calculation, as the other terms are subduction zone specific. The value of $N$ depends on the thermal parameter of the subduction zone, described by Syracuse et al. (2008). As depicted in Fig 1, both cold and intermediate subduction zones have similar $N$ values, making them effectively cold, for our purposes. We differentiate between hot and intermediate subduction zones at a thermal parameter of 1000 km. The velocity $v'$ is the descent rate per Ma and $L_{trench}'$ is the length of the subduction zone’s trench, both described by van Keken et al. (2011).

$F'$ is calculated for each subduction zone, with $N = 0.5$ for $^{36}$Ar in hot subduction zones, and $N = 1$ for every other thermal state and noble gas; that is, we assume a retention depth of 60 km in calculating effective $N$ values. The flux from each subduction zone is summed together to produce the total flux into the modern mantle, $F'_{earth}$. To find an effective set of $N$ values, we compare $F'_{earth}$ to the total flux if every subduction zone was cold ($N = 1$ for each noble gas), $F_{cold}$. By taking the ratio of these, we can find effective $N$ values for each noble gas:

\[ N_{eff} = \frac{F'_{earth}}{F_{cold}}. \]  

(26)

As we see in Eq. 26, if $N = 1$ for each zone, then there is no difference between numerator or denominator, and $N_{eff} = 1$. This is true for $^{84}$Kr and $^{130}$Xe, as there is no difference in their $N$ values in hot versus cold subduction. However, $^{36}$Ar does fractionate in hot subduction zones, and as a result we find that $N_{eff} = 0.9452$ for $^{36}$Ar in modern day Earth subduction.

The difference between the hot and cold subduction steady states (seen in Fig 2) is the hot subduction elemental fractionation of argon ($N = 0.5$) as opposed to no fractionation ($N = 1$) for cold subduction. With our new $N_{eff}$, the cold subduction steady state shifts closer to the hot steady state by approximately 6 percent (shown in Fig 2). This change has a negligible effect on the overall results.
3.4.2 Strong Plate Model

Thus far we have assumed a “classic” model of thermal evolution of the mantle, with a scaling factor of $\beta = 1/3$. In this model, higher temperatures of the early Earth cause more vigorous convection, a thinner rigid lithosphere and a higher plate velocity. This results in a very quick initial e-folding time for the noble gases in the mantle reservoir. As the Earth cools, the plate velocity decreases, increasing the e-folding time until reaching the present value of approximately 1 Gyrs. However, an alternative models suggest a very different situation where plate velocities have have remained constant (Conrad & Hager, 1999) or actually gradually increased over Earth history (Korenaga, 2006). In Korenaga’s model, for instance, higher mantle potential temperatures would result in a larger depth of melting beneath spreading ridges. Water in this region will preferentially partition into the melt and be lost to the surface, creating a region of dehydrated mantle beneath the thermal lithosphere. This will behave rigidly and act as part of the lithosphere. Thicker lithosphere is harder to bend at subduction zones and therefore causing plates to move more slowly than the “classic” mantle convection scaling laws suggest, yielding a negative $\beta$ (Korenaga, 2006).

We test how the alternative thermal history models with lower $\beta$ values would influence our results by setting a constant plate speed, as it is plate speed that determines how quickly the mantle equilibrates and therefore predicting constant e-folding times throughout history. We find no significant difference between our results with $\beta = 1/3$ and the constant plate speed model, because in either case the mantle spends the last 2-3 Gyrs with plate speeds about equal to today, and thus cold subduction has to start within the last 1 Gyr to keep the mantle from having fully equilibrated with the cold subduction input.

4 Discussion

Although our overall results are robust to uncertainties in the modern day noble gas contents of the crust, atmosphere, and mantle, noble gas retention depths, and changes in fundamental mantle convection scaling laws, there are further important model assumptions that could affect the results. In this model we assume instantaneous mixing of the mantle, when in reality the mantle mixing time is finite. The timescales of mantle mixing is not well understood, but is estimated to be on the order of 1 Gyrs at modern day Earth conditions (Tackley, 2007). As our models indicate a switch from hot to cold subduction $\approx$ 1 Gyrs ago, mantle mixing could have an important effect on our re-
sults. In particular, the hot to cold subduction transition could have happened earlier than we estimate, with the fact that the modern mantle is not in steady-state with the cold subduction input of noble gases due to incomplete mixing. This could be tested with additional mantle measurements from different sites across the planet. If they all possess a similar noble gas ratio to Bravo Dome, then it would be reasonable to conclude that the mantle is well mixed and our results hold.

There is also evidence that our assumption of a closed system might also be inaccurate. The Earth’s total Xe content is significantly reduced compared to solar and meteoric abundances (Owen et al., 1992). Zahnle et al. (2019) described how Xe can be lost by coupling to $H^+$ ions during hydrogen escape. Geochemical data suggests this was an ongoing process until the loss of free hydrogen at the end of the Archean. This implies Xe escape does not influence our results, as we predict the hot to cold subduction transition time to be in the late Proterozoic or Phanerozoic. During these more recent times, there was no appreciable quantity of free hydrogen in the atmosphere, preventing Xe loss and thus the assumption of a closed system is reasonable.

Our model assumes the presence of plate tectonics on the Earth during its entire history. However, it is not known if this is the case. Numerous alternate tectonic histories have been proposed, such as a cyclic or episodic tectonic history, were the planet cycled between the stagnant and mobile lid regimes (Cawood et al., 2013; Wyman, 2018). Other proposals suggest the Earth was locked in a stagnant lid regime early on (Debaille et al., 2013; Brown, 2014; Griffin et al., 2014). They argue the transition to a mobile lid regime occurred at the end of the Archean. Mobile lid onset times during the late Archean do not affect our results, however, as it takes place over 1 Ga earlier than our earliest proposed hot-cold transition time. As the e-folding time of the mantle is either 1 Ga or faster, the mantle will have had enough time between the onset of mobile lid tectonics and the onset of overall cold subduction for the mantle to reach the hot subduction steady state, thus erasing whatever the Archean noble gas signature was. The Earth would then have been able to undergo the transition from hot to cold subduction in the last Ga, resulting in the noble gas ratios of the mantle we currently observe. Whatever tectonic regime was present during the Hadean and Archean Eons, as long as the Earth was in a state of hot subduction for $\approx 1$ Gyrs afterwards, it will not affect our overall results.

High pressure metamorphic rocks, such as blueschists, first appear in the rock record within the last 1 Ga (Brown, 2014). These rocks form at approximately 100 - 500 °C and at depths of 15 - 60 km, or
a pressure range of 0.5 - 2.5 GPa, and thus form more readily in cold subduction zones (Philpotts & Ague, 2009). Although preservation bias is always an important consideration in interpreting the geologic record, the appearance of high pressure metamorphism at approximately 1 Ga is an additional independent constraint that modern style cold subduction began on Earth approximately 1 Ga, as our models based on mantle noble gas ratios also predict. However, there is evidence that the emergence of blueschist in the rock record is instead due to a change in oceanic crustal composition as a result of secular mantle cooling, and the correlation between our predictions and the appearance of blueschist is therefore coincidental (Palin & White, 2016).

5 Conclusion

We constrain the history of subduction zone thermal regime on Earth using $^{84}$Kr/$^{36}$Ar and $^{130}$Xe/$^{36}$Ar ratios. In a subduction zone, all noble gases are eventually lost from the slab, but when they are lost depends on the subduction zone’s thermal regime. Noble gas loss in hot subduction zones is elementally fractionated, while loss in cold subduction zones is unfractionated. Specifically, $^{36}$Ar is preferentially lost in comparison to $^{130}$Xe and $^{84}$Kr in hot subduction zones and therefore the mantle noble gas input ratios and thermal history of subduction can be deduced. With a constant plate speed, the model permits a simple analytical solution for the time evolution of mantle noble gas contents. From this solution two end-member steady states for the mantle are found: one for a mantle that has come to equilibrium with input via hot subduction zones, and one in equilibrium with input from cold subduction zones. The modern MORB mantle lies in between these steady states. As the Earth is currently in a primarily cold subduction thermal regime, the mantle is thus not at steady state with the current input of noble gases into the mantle at cold subduction zones, but instead transitioning from the hot subduction steady state to the cold. The simple analytical solution has a characteristic time scale to reach steady state, that depends on plate speed. For modern day plate speeds, this time scale is $\approx 1$ Gyr, and thus the onset of cold subduction must have been within the last 1 Gyr, or the Earth’s mantle would be at the cold subduction steady-state. To more precisely estimate the onset of cold subduction and confirm the predictions of the simple analytical solution, we simulate Earth’s thermal evolution over 4.5 Gyr. We find that the transition from hot to cold subduction occurred between 500 Ma to 1 Ga ago at the earliest, during the Neoproterozoic.
While our model predicts that this transition occurred within the last Ga, it is a simplified description of subduction history. In reality, the thermal regime of global tectonics is gradually shifting from hot to cold, with the most significant changes occurring in the latter half of Earth history, the transition point representing when cold subduction became dominant. This result is robust, as the same general conclusion holds despite a wide variety of retention depths, heat flux scaling laws, and reference values. Future work to better constrain the noble gas abundances of the ocean crust would tighten our estimates of the onset of predominantly cold subduction on Earth. Moreover, additional measurements of the noble gas contents of the modern mantle at different sites around the world are needed to better constrain the modern mantle $^{84}\text{Kr}/^{36}\text{Ar}$ and $^{130}\text{Xe}/^{36}\text{Ar}$ ratios, and reveal whether the mantle is strongly heterogeneous in these species or well-mixed.
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