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Continental recycling: The oxygen isotope point of view

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[1] Mass balance calculations based on a data compilation lead us to estimate a $\delta^{18}O$ value of 8.9 ± 0.7‰ for the continental crust. This isotopic composition is the product of two competing processes, namely, the erosion and growth of continental masses. Erosion tends to enrich the continental crust in ^{18}O through low-temperature isotopic exchange with the hydrosphere, whereas continental growth maintains the composition of the crust close to its mantle-derived precursor ($\delta^{18}O$ = 5.7‰). Box modeling of the oxygen isotope exchange between the continents, mantle, and seawater leads us to calculate a flux of subducted sediments averaged over the Earth's history of 0.4 km^3 yr^{-1}, significantly lower than most other recent estimates.

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1. Introduction

[2] The rate at which the continental crust is recycled into the mantle is still a matter of great debate. Most geological estimates for the rate of subduction of sediments into the mantle range between 0.5 and 0.7 km^3 yr^{-1} [*Rea and Ruff*, 1996; *Plank and Langmuir*, 1998]. When the mechanical erosion of the crust at subduction zones is considered, the present-day rate of continental material recycled into the mantle is 1.6 km^3 yr^{-1} [*von Huene and Scholl*, 1991]. This value is similar to the estimated rate of present-day crustal growth from the mantle (1.6 km^3 yr^{-1} [*Reymer and Schubert*, 1984]), which is based on material accreted during arc volcanism at convergent margins. However, these modern fluxes of continental growth and recycling cannot account for the present-day mass of the continental crust when averaged over the Earth's history and do not provide any information about its change through time. Crustal recycling rates deduced from Nd isotope models range between 0.8 ± 0.5 km^3 yr^{-1} [*Albarède*, 1989] and 2.5 km^3 yr^{-1} [*DePaolo*,

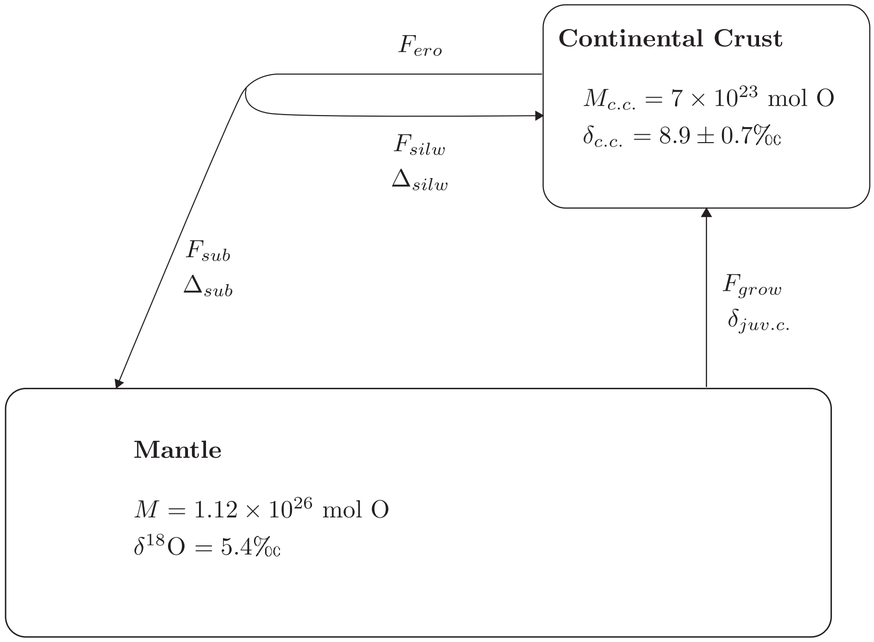

Figure 1. Schematic description of the model used to calculate the oxygen isotope evolution of continental crust. Parameters are presented in Table 2.

1983], but may be sensitive to the assumed structure of the mantle [*Coltice et al.*, 2000]. Moreover, a recent study based on Ar isotope data [*Coltice et al.*, 2000] yields estimates lower than 1 km^3 yr^{-1}. The knowledge of the recycling rate of continental crust into the mantle has important consequences for the evolution of continental growth, thermal regime of the Earth and chemistry of the mantle. In this paper, the fractionation properties of oxygen isotopes that operate during the low-temperature interactions between continental crust and hydrosphere are used to quantify sediment recycling into the mantle.

[3] At Earth's surface temperatures, weathering products such as clays are ^{18}O-enriched at the expense of the hydrosphere. The oxygen isotope evolution of the continental crust results from two competing processes. At active margins, sediments are injected into the mantle in subduction zones, preferentially recycling crustal components enriched in ^{18}O relative to the mantle source of continental crust ($\delta^{18}O \sim 6‰$). The remaining part of the total sediment discharge, characterized by high $\delta^{18}O$ values, is reincorporated into the continents via accretion prisms and passive margins, and contributes to its ^{18}O enrichment relatively to its mantle derived precursor. An ^{18}O enrichment of the continental crust through time is thus expected to reflect a minimal rate of sedimentary recycling into the mantle.

2. Long-Term Oxygen Isotope Cycle

[4] The mean flux of recycled sediments is quantified through box modeling of the oxygen isotope geochemical cycle, which involves the continental crust, mantle, and hydrosphere (Figure 1). Compared to previous models based on radiogenic trace elements, the oxygen isotope cycle presents some unique properties. Oxygen is equally shared by the various lithologies constituting the continental crust, thus reducing the uncertainty of the estimate of its isotopic composition. As low-temperature processes cause high oxygen isotope fractionation between silicates and fluids, the $\delta^{18}O$ value of the continental crust should reflect the amount of eroded continental mass that is reincorporated into the continents. Modeling the oxygen isotope cycle requires estimations of the $\delta^{18}O$ value of the continental crust, isotopic fractionation between the continental crust and seawater, and erosional fluxes.

[5] Oxygen isotope data, compiled from the literature, have been weighted according to the proportions of the various lithologies found within the continental crust (Table 1). The $\delta^{18}O$ values of samples from such lithologies have been summa-

Table 1. Compilation of Data Used to Calculate $\delta^{18}O$ Value of Continental Crust[a]

Reservoirs	Mass (kg) or Fraction	$\delta^{18}O$ (‰ SMOW)	±σ
Upper crust	**0.53**	**9.7***	**1.3**
Sedimentary rocks	*0.14*	**14.0***	**1.0**
Shales	0.440	15	2.0
Sandstones	0.209	11	2.0
Mafic/Volcanic	0.203	6.4	1.1
Carbonates	0.122	25.8	2.0
Felsic intrusives	*0.50*	9.4	2.4
Gabbros	*0.06*	6	2.0
Metamorphic rocks	*0.30*		
Gneisses	0.640	8.5	2.0
Schists	0.154	11.0	1.8
Amphibolites	0.178	6.4	2.0
Marbles	0.026	14.3	3.9
Lower crust	**0.47**	**8.1***	**1.4**
Felsic granulites	*0.615*	8.5	2.0
Mafic granulites	*0.385*	7.5	1.8
Bulk continental crust	2.5×10^{22}	**8.9***	**0.7**

[a] The fraction of various lithologies are from *Wedepohl* [1995]. Numbers in bold with an asterisk are mass balance calculated values.

rized in frequency histograms (Figure 2). Mafic granulites, which constitute a significant part of the lower crust, have a mean $\delta^{18}O$ of 7.5 ± 1.8‰ based on 171 selected samples taken from strictly mafic xenoliths and surface terranes (Table 1; Figure 2). The oxygen isotope composition of the felsic granulites, constituting the other part of the lower crust [*Wedepohl*, 1995], is considered equal to the mean $\delta^{18}O$ of felsic gneisses. This leads to a calculated $\delta^{18}O$ value for the lower continental crust of 8.1 ± 1.4‰. The surficial part of the upper crust, which consists of sedimentary rocks and volcanics, is thoroughly documented and data sets are already well-established [*Savin and Epstein*, 1970; *Veizer and Hoefs*, 1976; *Savin and Yeh*, 1981; *Longstaffe*, 1987; *Harmon and Hoefs*, 1995]. A compilation of oxygen isotope analyses of intrusive and metamorphic rocks of the continental crust reveals that the $\delta^{18}O$ of the upper continental crust equals 9.6 ± 1.3‰ (Figure 2; Table 1).

[6] According to the composition given by *Wedepohl* [1995], the average $\delta^{18}O$ value of the continental crust is found to be 8.9 ± 0.7‰ (Table 1). The isotopic enrichment of the continents relative to the precursor rocks derived from the mantle by partial melting is documented by the pioneering work of *Silverman* [1951]. With regard to the current study, the computed $\delta^{18}O$ value for the continental crust is in agreement with the previous estimates provided by *Taylor* [1974] and *Shieh and Schwarcz* [1977] ($\delta^{18}O \geq 7.5$‰). The initial composition of the continental crust generated from a mantle source having a $\delta^{18}O$ value of 5.5 ± 0.2‰ [*Eiler*, 2001] is estimated to be close to 5.7‰ according to the small isotopic fractionation that occurs during magmatic differentiation [*Taylor*, 1974; *Taylor and Sheppard*, 1986; *Weis et al.*, 1987; *Eiler*, 2001].

[7] In order to estimate the influence of changing the percentage of the various crustal lithologies on the average oxygen composition of the continental crust, a parametric resampling procedure closely related to the nonparametric bootstrap technique was applied. This procedure consists of pseudo-randomly and independently generating n equiprobable pseudosamples from the fractions of the continental crust lithologies, taking into account an arbitrary error associated with these proportions (i.e., a standard deviation σ that corresponds to an uncertainty of 30% on the fractions). Having a fraction $f^i_{obs} \pm \sigma^i$ for each crustal lithology, n = 50000 pseudosamples were generated by pseudo-random resampling of each value of the "observed" fractions via a simple Gaussian Model. For each fraction of the continental crust, a pseudo-random value f^i_{boot} was computed, such as

$$f^i_{boot} = f^i_{obs} + \left(\sigma^i \times \nu\right) \quad \text{where} \quad \nu \sim \mathcal{N}(0;1). \qquad (1)$$

Each fraction is then divided by the sum of all fractions of the continental crust (the sum of the fractions should be equal to unity) and the $\delta^{18}O$ value of the continental crust is calculated. The median $\delta^{18}O$ value equals 8.90 ± 0.04 with a 95% confidence interval. Our estimation of the average oxygen isotope composition is statistically very robust with regards to the uncertainty associated with the proportions of rock types in the continental crust.

[8] The oxygen isotope fractionation between the continental crust and seawater cannot be easily calculated from mineral-water equations that assume a mean temperature of isotopic exchange. Moreover, some rock-forming minerals, such as quartz, experience very limited oxygen isotope exchange with water at low temperature for kinetic reasons [*Clayton et al.*, 1978]. Therefore the isotopic compositions and proportions of sedimentary rocks (15.9 ± 1.3‰; Table 1) are used to estimate the apparent fractionation factor between continental crust and seawater. As the metamorphism of sediments, which is responsible for the release of

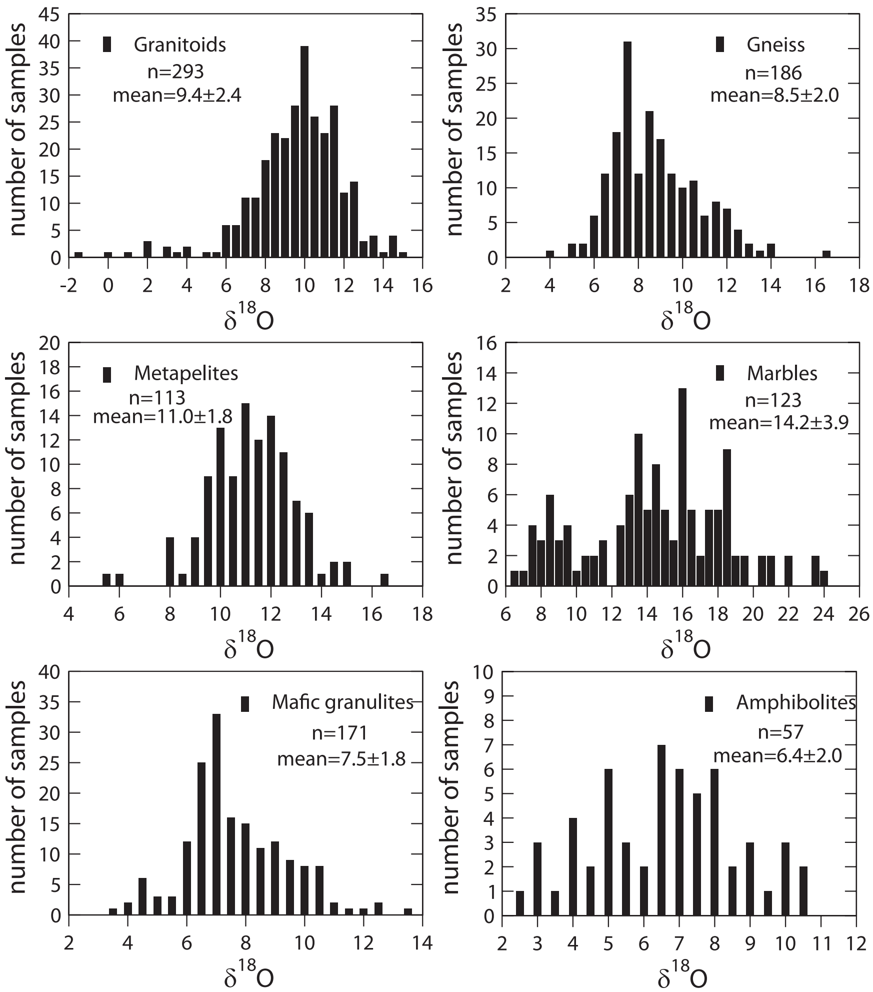

Figure 2. Frequency distribution of $\delta^{18}O$ values of selected rock samples of various lithologies from the continental crust. References used for the data compilation are listed in the auxiliary material.

^{18}O-rich fluids back to seawater [*Sheppard*, 1981], must be taken into account in the $\delta^{18}O$ budget of the continent, a simple mass balance between sediments and metasediments (principally schists and marbles) is used to calculate a bulk fractionation factor of 15.7 ± 1.0‰ corrected from the metamorphic cycle. It is noteworthy that this value corresponds to the average composition of shales (15‰), suggesting that it approximates well the amplitude of oxygen isotope exchange between the crust and the hydrosphere.

[9] Sediments entering subduction zones suffer high pressure metamorphism and dehydration, responsible for a decrease of their $\delta^{18}O$. High-pressure metamorphised rocks in the blueschist and eclogitic facies are remains of subducted rocks. The mean $\delta^{18}O$ of metasediments that have been metamorphised in the blueschist and eclogitic facies is calculated to be 11 ± 2‰ [*Taylor et al.*, 1963; *Devereux*, 1968; *Blattner and Cooper*, 1974; *Longstaffe and Schwarcz*, 1977; *Magaritz and Taylor*, 1981; *Park et al.*, 1999; *De et al.*, 2000]. This oxygen isotope value is based only on metasediment isotopic compositions and should well approximate the composition of subducted sediments into the mantle.

[10] The oxygen isotope composition of the hydrosphere and its evolution during the 4.56 Gyr of Earth's history is still the subject of intense controversy. Most researchers who modeled the ^{18}O cycle have concluded that the oxygen isotope composition of the hydrosphere is buffered by the mantle during hydrothermal activity at deep-sea floor and weathering. The upper part of the altered oceanic crust is ^{18}O-enriched relative to the MORB mantle source and is balanced by a ^{18}O depletion

[1] Auxiliary material is available at ftp://ftp.agu.org/apend/gc/2005GC000958.

Table 2. Main Parameters and Forcing Functions of the Model

Parameter	Description	Value
$M_{c.c.}$	oxygen content of the continental crust	7×10^{23} mol at $t = 0$
$Area_{c.c.}$	continental crust area	
F_{grow}	oxygen flux from the mantle to the continental crust during continental growth	$f(t)$
F_{ero}	oxygen flux of erosion	$(F_{sub} + F_{silw}) \propto Area_{c.c.}$
F_{sub}	oxygen flux of sediments subducted into the mantle	*model output*
F_{silw}	oxygen flux of sediments reincorporated into the continental crust	*model output*
$\delta_{c.c.}$	$\delta^{18}O$ of the continental crust	8.9 ± 0.7‰ at $t = 0$
δ_{sw}	$\delta^{18}O$ of the oceans	−1‰ (CST run) variable (WALL run)
$\delta_{juv.c.}$	$\delta^{18}O$ of the juvenile crust	5.7 ± 0.2‰
Δ_{sub}	oxygen isotope fractionation between subducted sediments and seawater	12 ± 2‰
Δ_{silw}	oxygen isotope fractionation between continental crust and seawater	15.7 ± 1‰

of the lower crust [*Muehlenbachs and Clayton*, 1976; *Gregory*, 1991]. Results of the mass balance between the oceanic crust and seawater allow for the conclusion that the oxygen isotope composition of seawater might have been constant throughout the Earth's history [*Gregory and Taylor*, 1981; *Holland*, 1984; *Jean-Baptiste et al.*, 1997; *Muehlenbachs*, 1998]. Within the framework of a dynamic model of seawater-crust interaction, and taking into account the expansion rate, permeability profile, mineralogical mode of the crust and kinetics of oxygen isotope exchange, *Lécuyer and Allemand* [1999] reached the conclusion that seawater had a nearly constant $\delta^{18}O$ value of 0 ± 2‰. However, *Walker and Lohmann* [1989], *Veizer et al.* [1999], and *Goddéris and Veizer* [2000] showed that significant long-term seawater $\delta^{18}O$ change would occur if the ratio between high-temperature, on-axis hydrothermal processes and low-temperature, off-axis oceanic crust alteration has changed. By using a revisited box model for global water and ^{18}O cycling, *Wallmann* [2001] proposed a secular evolution of the oxygen isotope composition of Phanerozoic seawater with a minimal $\delta^{18}O$ value of −8‰ during the early Cambrian.

3. Formulation of the Model

[11] The oxygen isotope evolution of the continental crust, interacting with both mantle and hydrosphere (Figure 1), is modeled by using a time-dependent box model that applies a first-order kinetic law of mass transfer. The ^{18}O budget of the continents can be written as follows:

$$\begin{aligned} M_{c.c.} \cdot \frac{d\delta_{c.c.}}{dt} &= F_{grow} \cdot (\delta_{juv.c.} - \delta_{c.c.}) \\ &\quad - F_{sub} \cdot (\delta_{c.c.} - \Delta_{sub} + \delta_{sw}) \\ &\quad + F_{silw} \cdot (\Delta_{silw} + \delta_{sw} - \delta_{c.c.}), \end{aligned} \qquad (2)$$

where $M_{c.c.}$ is the oxygen mass of the continental crust (about 7×10^{23} mol); $\delta_{c.c.}$ is the $\delta^{18}O$ of the continents; $\delta_{juv.c.}$ is the $\delta^{18}O$ of the juvenile crust; δ_{sw} the $\delta^{18}O$ of seawater; F_{grow} is the flux of oxygen transferred from the mantle to the continental reservoir during continental growth; F_{sub} and F_{silw} are the respective oxygen fluxes for eroded continental material subducted into the mantle and incorporated back to the crust; Δ_{sub} is the oxygen isotope fractionation between subducted sediments and seawater; and Δ_{silw} is the isotopic fractionation between continental crust and seawater.

[12] The present-day global erosion flux is comprised between 2 and 4 km^3 yr^{-1} according to most estimates based on river fluxes and chemical balances [*Gilluly*, 1964; *Berner et al.*, 1983; *Drever et al.*, 1988; *Walker and Lohmann*, 1989; *Hay*, 1998]. Because both interactions between solid Earth and hydrosphere are implied, the global erosion flux ($F_{sub} + F_{silw}$) is considered proportional to the area of the continental crust through time. This is a very simplified model since in reality silicate weathering is dependent on air temperature [*Walker et al.*, 1981; *Berner et al.*, 1983; *Dessert et al.*, 2003] and on other processes such as runoff [*Amiotte-Suchet*, 1995], land plant coverage [*Berner*, 1998], uplift and relief [*Goddéris and François*, 1995], and possibly oxygen atmospheric content. However, the quantification of the impact of these processes over 4.56 Gyr is not possible so far, and even erosion flux variations of 50% will not drastically change conclusions when averaged over the entire Earth's history.

[13] Three models of continental growth have been taken into account (Figure 3). In the first scenario (model A), the mass of the continental crust increased linearly during the first 1.5 Gyr and then remained constant until now, as proposed by

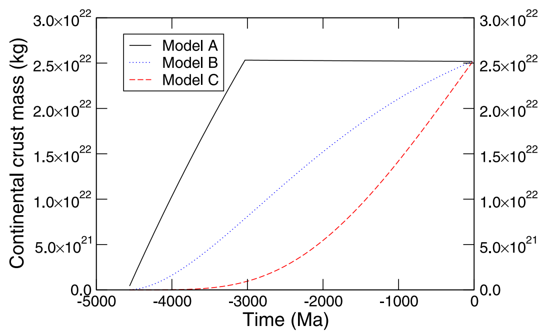

Figure 3. Evolution of continental mass through Earth's history according to three models: A, early crustal accretion followed by steady state; B, maximum crustal growth during Late Archean; C, hypothetical late crustal growth.

Armstrong [1968]. Such a model implies the presence of a large mass of continental crust very early in the Earth's history and a steady-state of the crust-mantle system. In the second scenario (model B), the formalism developed by *Albarède* [1998] is used. It has the attractive property to fix a maximum growth rate whose time and amplitude can be adjusted:

$$\frac{\phi_{cc}(t)}{\phi_{cc}(t_0)} = k \cdot \left(\frac{t}{t_0}\right)^{\alpha-1} \cdot \exp\left(-\beta\frac{t}{t_0}\right), \qquad (3)$$

where ϕ_{cc} is the continental growth rate, t_0 = 4.56 Gyr, and α and β are adjustable parameters. According to the procedure of *Albarède* [1998], the constant k is evaluated using incomplete gamma functions [*Press et al.*, 1992] in such a way that the right-hand side of equation (3) sums to 1 when integrated for t over $[0, t_0]$. Hereby, the maximum growth rate of the continents is fixed at 2.5 Gyr ago (Figures 3 and 4a). This model of maximum crust accretion during the Late Archean is supported by most of the geological and geochemical evidence available [*Hurley and Rand*, 1969; *Nelson and DePaolo*, 1985; *Patchett and Arndt*, 1986; *Taylor and McLennan*, 1995]. In the third scenario (model C), α and β from equation (2) were adjusted such as the maximum continental growth took place about 1 Ga ago (Figure 4a). This scenario of late continental crust growth was considered to test the influence of the timing of continental extraction on the recycling rate of subducted sediments, although it is unrealistic.

[14] The calculated evolution of the continental $\delta^{18}O$ depends on the variations in the oxygen isotope composition of the oceans through time. Two runs have thus been performed for each scenario of continental growth. The CST runs assume a constant $\delta^{18}O$ value of seawater of −1‰ during the Earth's history. In contrast the WALL runs use a $\delta^{18}O$ value of seawater that evolves as proposed by *Wallmann* [2001]: a constant marine $\delta^{18}O$ value of −3‰ during the Archean and Proterozoic, a minimal value of −8‰ 600 Ma ago and a progressive increase in the $\delta^{18}O$ value of seawater up to the Cenozoic value of −1‰.

4. Quantification of Crustal Recycling

[15] Given a global erosion flux, a unique solution for the recycling rate of subducted sediments is calculated for the $\delta^{18}O$ value (8.9 ± 0.7‰) of the continental crust after 4.56 Gyr of evolution of the geochemical cycle. Curves have been computed to show the evolution of the continental mass (Figure 3), rates of continental growth (Figure 4a), and global erosion fluxes and recycling rates of subducted sediments (Figure 4b). When averaged

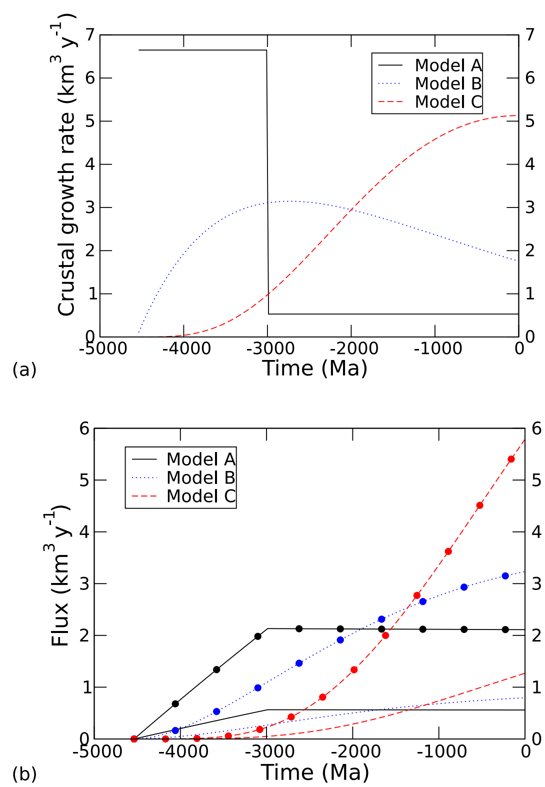

Figure 4. Evolution of growth, erosion, and recycling of the continental crust. (a) Computed curves depict the three considered models of continental growth through time. (b) Computed curves show the evolution of erosion (●) and recycling fluxes (without dot) for the three considered models. Fluxes are calculated using a global erosion flux of 1.7 km^3 yr^{-1} averaged over 4.56 Gyr.

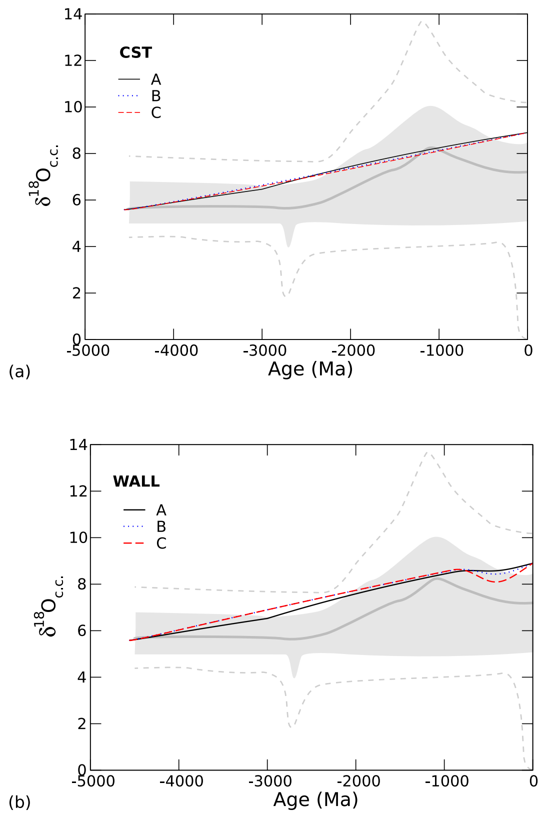

Figure 5. Evolution of the oxygen isotope composition of the continental crust from a juvenile δ^{18}O value of 5.7‰ to a present-day value of 8.9‰ for (a) CST and (b) WALL runs, and for the three considered models A, B, and C of continental growth. The shaded area and the gray curve represent the averaged trends of oxygen isotope composition of zircons [*Valley*, 2003]. The dotted envelope includes all zircon samples from the compilation of *Valley* [2003].

over 4.56 Gyr, an erosional flux of 1.7 km^3 yr^{-1} corresponds to present-day fluxes comprised between about 2 and 6 km^3 yr^{-1}, depending on the considered scenario of crustal growth (Figure 4b). The later the continental crust maximum growth rate takes place (Model C), the lower the calculated averaged flux is for a given present-day growth rate. Obviously, in the case of the CST runs (constant seawater δ^{18}O), whatever the considered model of continental growth, the same averaged recycling flux is obtained for the given averaged erosion flux of 1.7 km^3 yr^{-1}. The calculated flux equals 0.4 km^3 yr^{-1} and corresponds to a present-day crustal recycling rate ranging from 0.5 km^3 yr^{-1} (Model A) to 1.2 km^3 yr^{-1} (Model C) (Figure 4b). The computed increase of the continental crust δ^{18}O from 5.7 to 8.9‰ is supported by the oxygen isotope composition of magmatic zircons [*Peck et al.*, 2000; *Valley*, 2003], whose δ^{18}O increase from mantle values (5.3‰) during the Archean to values commonly higher than 8‰ after 2.0 Gyr [*Valley*, 2003]. The nature of the zircon δ^{18}O increase (sharp or gradual) is not known because of the lack of Paleo-Proterozoic zircon analyses [*Valley*, 2003], making difficult any precise comparison between the zircon δ^{18}O trend and our modeled δ^{18}O evolution of the continental crust. However, the increase of the δ^{18}O of magmatic zircons reflects recycling of low-temperature altered crustal material in the source of the magmas. The computed continental crust δ^{18}O evolution remains similar and cannot be used to decipher the various continental growth modes or the two proposed models of the evolution of seawater δ^{18}O (Figure 5).

[16] Taking into account the geologically plausible range of averaged erosion fluxes from 0.5 to 3.5 km^3 yr^{-1}, which corresponds to the estimated range of present-day fluxes from 2 to 4 km^3 yr^{-1}, calculated recycling rates vary from 0 to 1.4 km^3 yr^{-1} (Figure 6a) for the CST runs. The uncertainty on the recycling flux are deduced from the sensitivity of the model to the critical parameters: oxygen isotope composition of the continental crust (±0.7‰), crust-water fractionation value (±1‰) and fractionation between subducted sediments and water (±2‰). Considering the cumulated uncertainties associated with these three parameters, the maximum recycling flux into the mantle is equal to 1.4(±0.6) km^3 yr^{-1} (Figure 6a) for a maximum averaged erosion flux of 3.5 km^3 yr^{-1}. It is also noteworthy that the uncertainties associated with the global erosional flux lead a wider range of calculated recycling rates. The WALL runs display different results as a function of the continental growth model (Figure 6b). Late crustal accretion (model C) leads to lower calculated recycling rates for a given erosion flux. In this case, the maximum recycling flux equals 1.2 ± 1.0 km^3 yr^{-1} and is obtained for the continental growth model A. More generally, calculated recycling rates are lower for WALL runs than for CST runs, but the sensitivity to isotopic composition and fractionation factors is larger (Figure 6).

[17] Late crustal accretion, as described in the scenario C, is not reasonable in the light of geological and geochemical data. With such a model, the present-day ^{18}O enrichment of the bulk continental crust is difficult to obtain in the framework of the present-day erosion flux that does not exceed 4 km^3 yr^{-1}, even without crustal recycling (Figure 6). A model of maximum crustal accretion

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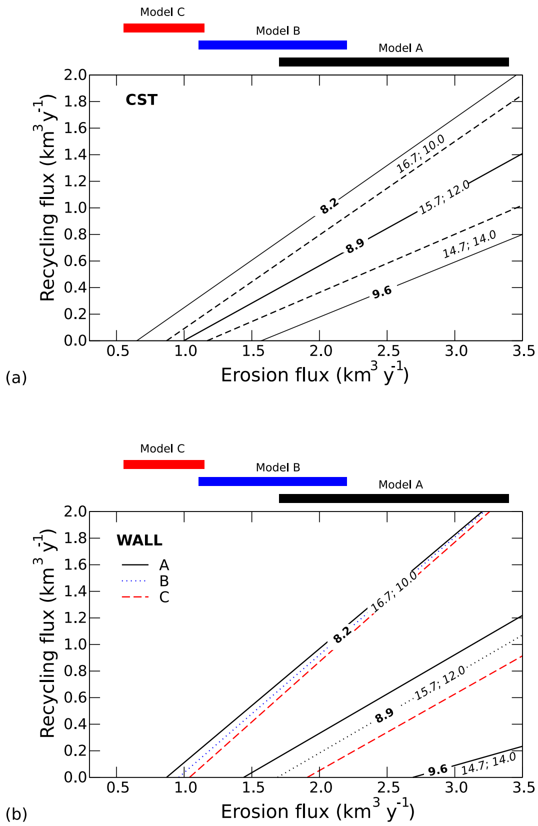

Figure 6. Recycling fluxes as a function of erosion fluxes. The bold numbers are the $\delta^{18}O$ values of the continents (8.9 ± 0.7‰). Slanted numbers are Δ_{silw} (15.7 ± 1‰) and Δ_{sub} (12 ± 1‰) values. The black bars above both figures indicate the range of averaged erosion fluxes that are calculated from the present-day erosion rates between 2 and 4 km^3 yr^{-1} for the three considered models of crustal growth. (a) CST runs. (b) WALL runs.

during late Archean (model B) is preferred on the basis of geological [*Hurley and Rand*, 1969; *Veizer and Jansen*, 1979; *Taylor and McLennan*, 1995] and geochemical [*Allègre and Rousseau*, 1984; *Nelson and DePaolo*, 1985; *Patchett and Arndt*, 1986; *Kramers and Tolstikhin*, 1997; *Collerson and Kamber*, 1999] studies. In the case of model B, the computed recycling rate lies between 0.1 and 0.6 km^3 yr^{-1} for continents with a $\delta^{18}O$ of 8.9‰, a Δ_{silw} of 15.7‰ and a Δ_{sub} of 12‰, having a constant seawater $\delta^{18}O$ of −1‰. With a variable $\delta^{18}O$ value of seawater since the Archean (WALL runs), recycling rates are lower than 0.4 km^3 yr^{-1}. Even when considering the cumulated uncertainties of the parameters, the recycling rate remains lower than 1 km^3 yr^{-1} (Figure 6). The continental growth model A, close to the curve proposed by *Armstrong* [1968], generates higher computed recycling rates (CST runs: 0.4–1.4 km^3.a^{-1}; WALL runs: 0.2–1.2 km^3 yr^{-1}). If we consider that the mass of the continental crust is at a steady-state, the present-day recycling rate should be equal to the actual continental crust production from the mantle (1.6 km^3 yr^{-1}; [*Reymer and Schubert*, 1984]). Averaged over the entire Earth's history, a recycling flux of 1.4 km^3 yr^{-1} is then required in the framework of model A. In this case, the $\delta^{18}O$ value of the bulk continental crust of 8.9‰ is explained by a present-day global erosion flux higher than 4 ± 1 km^3 yr^{-1}, which lies in the upper range of the geological estimates.

5. Conclusion

[18] Calculated recycling rates based on the $\delta^{18}O$ value of the continental crust since the Archean are lower than 2 km^3 yr^{-1}, regardless of the model considered for crustal growth and the evolution of the seawater $\delta^{18}O$ throughout the Earth's history. Moreover, the present-day enrichment of the continents relative to the mantle-derived rocks is unlikely to be achieved with a model of early crustal growth and steady-state of the crust-mantle system. Given a more realistic model of maximum continental accretion during the late Archean, a recycling rate of sediments into the mantle of $0.4 \pm ^{0.6}_{0.4}$ km^3 yr^{-1} is deduced for the present-day erosion flux, which is estimated between 2 and 4 km^3 yr^{-1}. These recycling fluxes are lower than the estimates previously proposed from radiogenic isotope modeling [*DePaolo*, 1983; *Albarède*, 1989]. It should be emphasized that these low recycling rates of sediments are compatible with both geological data [*von Huene and Scholl*, 1991; *Rea and Ruff*, 1996; *Plank and Langmuir*, 1998] and more especially with recent estimates given by *Coltice et al.* [2000] who established terrestrial inventories of ^{40}K and ^{40}Ar. A better knowledge of the global erosion flux should allow a refinement of the calculation of sediment recycling rates into the mantle, based on the modeling of the long-term oxygen isotope geochemical cycle. These results also suggest that the chemical evolution of the continental crust is mainly dominated by cannibalistic reworking of sediments.

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