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The fundamental role of island arc weathering in the oceanic Sr isotope budget

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ABSTRACT

We have re-assessed the Sr isotopic budget of the modern ocean taking into account the high erosion rates of volcanic islands, and especially of island arcs, emphasizing important contribution from subsurface weathering to global budgeting. We propose that intensive weathering on volcanic islands, island arcs and oceanic islands, coupled with large surface and subsurface water fluxes is the missing source of mantle-derived $^{87}Sr/^{86}Sr$ (0.703) in seawater Sr isotope balance. In our approach, it represents 60% of the actual mantle-like input of Sr to the oceans, the remaining 40% supplied by ridge-crest hydrothermal activity and sea-floor low-temperature alteration of basalts. The seawater Nd isotopic budget is consistent with this interpretation and explains well the regional contributions from ridge crest and island arc activity among the oceans.

1. Introduction

The $^{87}Sr/^{86}Sr$ isotopic ratio measured in a marine limestone registers the ocean water composition at the time of the limestone deposition. Limestone $^{87}Sr/^{86}Sr$ ratios do vary with geological time. The question asked since the pioneering work of Veizer and Compston (1974) and Brass (1976) is how to interpret it? How can we extract information relative to ocean geological history?

The answer to this fundamental challenge to geochemistry depends on one key question: what is the quantitative explanation of the present-day oceanic $^{87}Sr/^{86}Sr$, which is uniform worldwide.

For a long time it was admitted that seawater Sr isotopic ratio is inherited from the mixture of continental erosion products and ridge-crest hydrothermal fluids (Brass, 1976; Brevard and Allègre, 1977; Albarède et al., 1981; Chaudhuri and Clauern, 1986; Palmer and Edmond, 1989). This simple interpretation has been challenged recently and unsuccessful efforts have been made to give alternative explanations (e.g. Davis et al., 2003).

In the present paper, we give a new interpretation of the oceanic $^{87}Sr/^{86}Sr$ ratio, using the $^{87}Sr/^{86}Sr$ measurements on major rivers all over the world accumulated by our group and others. This model shows the fundamental importance of island arc contribution and opens the door to a reinterpretation of the geological oceanic $^{87}Sr/^{86}Sr$ record in limestone during the Whole Phanerozoïc (Allègre et al., 1996, in preparation).

2. Present day $^{87}Sr/^{86}Sr$ in the ocean

The actual $^{87}Sr/^{86}Sr$ ratio of seawater is 0.709176 ± 0.000003. This ratio is nearly constant all over the oceans as the strontium residence time in seawater is larger than 3 Ma and the mixing time of the oceans is about 1500 years (or longer) (Broecker and Peng, 1982).

Since the first studies of the seawater Sr isotopic variations through time as recorded in accumulated marine limestones (Gast, 1955; Hurley et al., 1965; Peterman et al., 1970; Veizer and Compston, 1974; Burke et al., 1982), it was admitted that the two major sources of strontium in the ocean were continental erosion transported by rivers and another component from the mantle, since the oceanic $^{87}Sr/^{86}Sr$ is different from the average continental values. Later on, with the discovery of hydrothermal circulation at ridge crests, it was admitted that this was the source of the Sr mantle component (Brass, 1976; Brevard and Allègre, 1977; Albarède et al., 1981; Palmer and Edmond, 1989). Different mass-budget estimates have since questioned this possibility on quantitative grounds (Fehn et al., 1983; Morton and Sleep, 1985; Sleep, 1991): the hydrothermal Sr flux is too low to explain the present-day $^{87}Sr/^{86}Sr$ of the ocean. Davis et al. (2003) tried to work out this problem by involving ophiolites or metamorphic processes, without success; they could not close the budget.

We have measured $^{87}Sr/^{86}Sr$ in 25 rivers of the world, completing the already existing data bank, which now gathers more than 30 principal rivers of the world (Table 1, data from Gaillardet et al., 1999 and references therein). The river discharge weighted average $^{87}Sr/^{86}Sr$ for this database is 0.7136 ± 0.0002. Mid-oceanic ridge hydrothermal $^{87}Sr/^{86}Sr$, measured by several authors, follows Albarède et al. (1981) and is 0.7029 ± 0.0003.

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Table 1
Strontium ($^{87}Sr/^{86}Sr$) and neodymium (ε_{Nd}) isotopic ratios and discharge of the main rivers worldwide. Data are gathered from literature and from unpublished results of IPGP group: (1) Gaillardet et al. (1999), (2) Goldstein and Jacobsen (1987)), (3) IPGP Geochemistry and Cosmochemistry group, (4) Goldstein et al. (1984). D stands for dissolved load (fraction finer than 0.2 or 0.45 µm) of the river water and SS for suspended sediment (fraction larger than 0.2 or 0.45 µm). Values of $(\varepsilon_{Nd})_D$ in square brackets have been estimated as proposed by Goldstein and Jacobsen (1987). Measurements errors (2 S.D.) are typically 0.00005 for $^{87}Sr/^{86}Sr$ and 0.2 for ε_{Nd}. Average values used in the discussion are discharge weighted.

	River	Discharge (km^3/a)	Ref	$(^{87}Sr/^{86}Sr)_D$	Ref	$(\varepsilon_{Nd})_D$	Ref	$(^{87}Sr/^{86}Sr)_{SS}$	Ref	$(\varepsilon_{Nd})_{SS}$	Ref
Atlantic	Amazon	6590	(1)	0.71148	(3)	{−10.7}	(3)	0.72146	(3)	−10.7	(3)
	Congo	1200	(1)	0.71918	(3)	−16.1	(2)	0.73160	(3)	−16.4	(3)
	Orinoco	1135	(1)	0.71830	(1)	{−14}	(3)			−14	(2)
	Parana	725	(2)	0.71390	(1)	−10.3	(4)			−10.3	(4)
	St. Lawrence	439	(2)	0.70959	(2)	−12.0	(2)	0.71265	(2)		
	Mackenzie	308	(1)	0.71138	(3)	−14.3	(2)	0.71886	(3)	−14.3	(3, 2)
	Magdalena	237	(1)			−8.3	(4)			−8.3	(4)
	San Francisco	94	(2)			{−12.9}	(4)			12.9	(4)
	Greenland	266	(2)			{−19}	(2)				
	Baffin Island	148				{−19}	(2)				
	Hudson Bay	273	(2)			{−25}	(2)				
	Guyana	240	(2)			{−20}	(2)				
	Niger	154	(1)	0.71433	(3)	−10.5	(2)			−10.5	(3)
	Mississippi	580	(2)	0.70953	(2)	{−11.2}	(2)	0.71918	(2)	−11.2	(2)
	Colorado	23	(2)	0.71075	(2)	−8.0	(2)	0.71077	(2)	−9.2	(2)
	Seine	13	(1)	0.70810	(3)	−10.7	(3)	0.71301	(3)	−11.7	(3)
Pacific	Fraser	112	(1)	0.7188	(3)	{−3}	(2)	0.72967	(2)		
	Columbia	200	(1)	0.71210	(2)	−3	(2)	0.71161	(2)	−4.5	(2)
	Mekong	467	(1)	0.71049	(3)	{−11.2}	(3)	0.71883	(3)	−11.2	(3)
	Chang Jiang	928	(1)	0.71074	(3)	{−10.6}	(3)	0.72194	(3)	−10.6	(3)
	Xi Jiang	363	(1)	0.71068	(3)	{−11.2}	(3)	0.72779	(3)	−11.2	(3)
	Huang He	41	(1)	0.71114	(3)	{−10.8}	(3)	0.71489	(3)	−10.8	(3)
	New Zealand	314	(2)			{0}	(2)				
	New Guinea	1360	(2)			{−3}	(2)				
	Japan	550	(2)	0.7076	(2)	{0}	(2)	~0.707	(2)	−0	(2)
	Philippines	332	(2)	0.7056	(2)	+6	(2)	~0.705	(2)	+7	(2)
	New Caledonia	368	(2)			{+8}	(2)				
	Indonesia	1000	(2)			{+2}	(2)				
Indian	Indus	90	(2)	0.71116	(2)	{−12.2}	(2)				
	Brahmapoutra	510	(1)	0.71970	(3)	{−10.5}	(3)	0.72442	(3)	−13.5	(3)
	Ganges	493	(1)	0.72490	(3)	{−14.5}	(3)	0.74486	(3)	−14.3	(3)
	Irrawady	486	(1)	0.71366	(3)	{−10}	(3)	0.71420	(3)	−14.4	(3)
	Salween	211	(1)	0.71408	(3)	{−8.7}	(3)	0.71458	(3)	−8.6	(3)
	Zambeze	103	(1)			{−10}	(2)				
	Murray	23.6	(1)	0.71075	(2)	−6	(2)	0.71428	(1)	−6	(2)
	Indonesia	1000	(2)			{+2}	(2)				
	India	182	(2)			{−20}	(2)				

The seawater (SW) mass balance equation between inputs from both continental crust (CC) and mantle (M) can be written:

$$\left(\frac{^{87}Sr}{^{86}Sr}\right)_{SW} = x\left(\frac{^{87}Sr}{^{86}Sr}\right)_{M} + (1-x)\left(\frac{^{87}Sr}{^{86}Sr}\right)_{CC} \quad (1)$$

x being the mass fraction of Sr coming from the mantle.

With the above Sr isotopic ratios for seawater, river and hydrothermal ridge end-members, we can compute that 41% of Sr is coming from the mantle and 59% from the continental crust.

The Sr input from the weathering of continental crust (CC) can also be split between carbonate (carb) and silicate (sil) rock weathering and a new mass-budget for strontium isotopic ratio assessed:

$$\left(\frac{^{87}Sr}{^{86}Sr}\right)_{CC} = y\left(\frac{^{87}Sr}{^{86}Sr}\right)_{sil} + (1-y)\left(\frac{^{87}Sr}{^{86}Sr}\right)_{carb} \quad (2)$$

y being the proportion of Sr arising from silicate rock weathering.

For the silicate rock weathering end-member, we have calculated an average $^{87}Sr/^{86}Sr$ ratio of the suspended loads carried by big rivers of 0.721 ± 0.004, using our own measurements and other published results (Table 1). Phanerozoïc limestones show $^{87}Sr/^{86}Sr$ variations from 0.707 to 0.709 (Peterman et al., 1970; Veizer and Compston, 1974) and since the Sr recycling time in the sedimentation–erosion cycle is about 100 to 150 Ma (Ronov, 1964; Brevard and Allègre, 1977, Berner et al., 1983; Palmer and Edmond, 1989; Veizer and Mackenzie, 2004; Wilkinson et al., 2009), we will take an average value of 0.708 ± 0.001 for the limestone contribution into the present ocean.

With the chosen values Eq. (2) leads to 43% of the continental river Sr coming from silicate rock weathering (without any apportioning to granitoïds, metamorphic rocks, shales or volcanic rocks) and 57% from carbonate rock weathering (limestone and dolomite) (Fig. 1).

In terms of fluxes, continental rivers, with a water flux of $3.74 \cdot 10^{19}$ g/a (Meybeck and Ragu, 1997) and a mean Sr concentration of 44 ppb $\pm$ 15 (data from Gaillardet et al., 1999 and references therein), bring $1.65 \cdot 10^{12}$ g of Sr per year to the oceans. The 41% Sr mantle input, previously defined by mass balance Eq. (1), leads then to a supposed ridge-crest hydrothermal Sr flux of $11.6 \cdot 10^{11}$ g of Sr per year. However, heat flow data have shown that hydrothermal flux at ridge crest is much smaller than previously thought (Fehn et al., 1983; Morton and Sleep, 1985; Sleep, 1991) and consequently not quantitatively capable of balancing the continental input to the seawater Sr isotope budget. The most recent estimate of Davis et al. (2003) gives a Sr flux from ridge-crest hydrothermal and low-temperature sea-floor basalt alteration of $3.15 \cdot 10^{11}$ g/a, which represents only 27% of the mantle flux needed. We have to find another mantle-like source of Sr accounting for $8.47 \cdot 10^{11}$ g of Sr per year.

3. Island arc (IAs) and oceanic island (OIs) weathering in the seawater strontium isotope budget

Classical estimations of the silicate weathering, and strontium, fluxes are based on studies of the largest rivers worldwide. However,

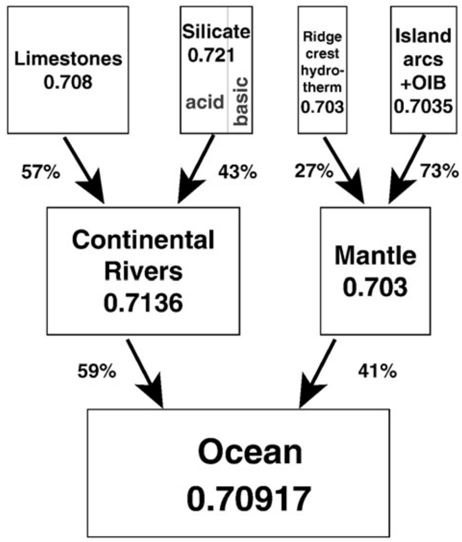

Fig. 1. Strontium isotope oceanic budget. 59% of the oceanic strontium is derived from continental weathering (both of silicate and carbonate rocks). For the remaining 41% of Sr from mantle sources, only 27% can be attributed to ridge-crest hydrothermalism. The missing 73% originates from island arc and OIB weathering (surface and underground), an important source of riverine material towards the oceans (Rad et al., 2007), which has been missed by previous studies.

subsequent to publication by Milliman and Meade (1983), demonstrating the particularly high sediment fluxes from small mountainous rivers in South-East Asia, and the publication of Meybeck (1986) pointing out the variable weatherability of different lithologies, small and medium size "basaltic" rivers have received an increased attention (e.g. Benedetti et al., 1994; Gislason et al., 1996; Négrel and Deschamps, 1996; Louvat and Allègre, 1997, 1998; Louvat, 1997; Dessert et al., 2001, 2009; Rad et al., 2007, 2008; Das et al., 2005; Louvat et al., 2008). These studies demonstrated that weathering rates for basalts are the highest among silicate rocks other than the ultramafics, and that rivers draining volcanic rocks have to be considered as a separate input in the worldwide geochemical budgets. For example the atmospheric CO_2 consumption by weathering of volcanic rocks represents 30 to 40% of the global consumption by silicate rock weathering (Dessert et al., 2003). While some rivers draining continental basalts are included in the average calculations of global weathering budgets, small or medium size volcanic islands are not, and they are also not included in the digitalized geographical/geological world maps (e.g. Amiotte-Suchet et al., 2003).

In terms of water fluxes, the classical approach to the budget of water flowing into the ocean is to consider the river water fluxes as established by hydrological survey statistics. This is not satisfactory for volcanic islands. First, the riverine hydrological budgets on these islands, when they exist, often underestimate the annual quantities of water flowing in the river catchments because the biggest events during storms or cyclones are difficult to monitor. Second, as pointed out by several authors (e.g. Milliman and Meade, 1993), the amount of water that is transmitted outside river channels is great particularly for volcanic islands because their hydrological network is immature. As an example, on Réunion Island, on the more recent part of the Piton de la Fournaise volcano with no rivers, the infiltrated water represents almost 90% of the rain water once the evapo-transpiration is subtracted and for the most mature part of this island with the largest rivers it still represents 30% (Folio, 2001). In the Lesser Antilles, where pyroclastic flows widely distributed on the Martinique and Guadeloupe highly enhance the porosity within the river catchments, subsurface water flow is enhanced accordingly (Rad et al., 2007). For the estimation of IAs and OIs water flux to the oceans we have chosen another approach based on mean rainfall averages.

From geographical and geological maps, we have calculated an approximate cumulative surface of $2.02 \cdot 10^6$ km^2 for IAs and OIs worldwide, only the volcanic part of the island surface areas has been taken into account (Table 2). Extending this surface to a surface area weighted average rainfall of 2800 mm/a (Table 2), we deduce a water flux of $5.66 \cdot 10^{18}$ g/a from volcanic islands (IAs plus OIs). Subtracting an evapo-transpiration ratio of 25%, yields about $4.24 \cdot 10^{18}$ g of water per year. Thus, accepting that all this water ultimately returns to the oceans via permanent rivers, subsurface flows and groundwaters, a mean Sr concentration of 200 ppb in the IAs and OIs waters is needed to close the Sr isotope oceanic budget (required to generate a flux of $8.45 \cdot 10^{11}$ g of Sr per year).

Our study of subsurface water fluxes on the IAs of Antilles and the OI of Réunion (Rad et al., 2007) has shown that due to underground alteration, the underground contributions of Sr to the ocean are about 2 to 3 times higher than is the case for surface rivers. With subsurface water fluxes to the evaluated systematically for the whole Earth, the global geochemical budgets may be quite different from what they are today.

We compiled Sr concentrations for rivers of different IAs and OIs (Fig. 2a). Despite the fact that this list includes only few island arc rivers (mostly Java, Lesser Antilles and Kamchatka), the distribution of Sr has a near-Gaussian shape centered at 35 ± 10 ppb. This summary does not include hot and cold underground waters with longer residence time, many injected directly to the sea as springs. As shown by Rad et al. (2007) for the Antilles and Réunion islands, these underground waters are far more concentrated in Sr than the surface waters. The computed average Sr concentration for island arc and OI waters is thus a minimum value. Compilation of strontium concentrations for some thermal springs in Réunion, Iceland and Lesser Antilles (Louvat and Allègre, 1997; Rad et al., 2007; Louvat et al., 2008) yields a mean Sr concentration of 450 ± 150 ppb (Fig. 2b). With such a concentration, an admixture of 40% of thermal springs and/or underground waters to the river waters is sufficient to reach 200 ppb, the needed Sr concentration.

The above considerations suggest that the continents and islands discharge $37.4 \cdot 10^{18}$ g and $4.24 \cdot 10^{18}$ g of water/a, respectively. Intuitively that ratio of 8.8/1 seems to be rather small. However, as pointed out by Milliman and Meade (1993)) the contribution of the island arc and small mountainous rivers with high erosion potential (e.g. small tropical rivers weathering basalts) to the global suspended sediment load is disproportionally high. We claim the same for the dissolved load and, if so, weathering of the island arc and OI is the missing flux that is necessary to close the seawater strontium isotopic balance.

4. Validation with the neodymium isotopic budget

Nd has a much shorter residence time than Sr in seawater and, while not well established, its isotopic composition is variable among different oceans (Arsouze et al., 2007). For Nd, in contrast to Sr, hydrothermal circulation at ridge crest is not a significant source (e.g. Michard et al., 1983; Piepgras and Wasserburg, 1985; Rudnicki and Elderfield, 1993). Ultimately its isotopic composition in seawater reflects erosion of continental and volcanic sources.

Utilizing the average Nd isotope ratio in seawater for the different oceans given in Arsouze et al. (2007) (Fig. 3), and weighing by the relative size of each ocean, we compute a world oceanic average $(\varepsilon_{Nd})_{SW}$[2] of -6.9 ± 1, which is meaningful only for a global mass balance exercise and its comparison to Sr.

Goldstein and Jacobsen (1987) in their study of rivers worldwide demonstrated that for river, the ε_{Nd} of the suspended material was on average similar to the ε_{Nd} of the dissolved load. We have used this

[2] $(\varepsilon_{Nd})_X = \{(^{143}Nd/^{144}Nd)_X/(^{143}Nd/^{144}Nd)_{CHUR} - 1\}*10{,}000$, with CHUR = primitive mantle.

Table 2
Surface area, proportion of volcanic rocks and average rainfall of oceanic islands in the Atlantic, Pacific and Indian Oceans, compiled from various geographic, geologic and hydrologic maps and databases after Louvat (1997).

	Island name	Type	Area (km^2)	% volca rocks	Rainfall (mm/a)		Island name	Type	Area (km^2)	% volca rocks	Rainfall (mm/a)
Atlantic	Iceland	OI	103,000	100	1600	Pacific	1/2 Indonesia (without PNG)	IA	1,033,250	25	3500
	Svalbard	OI	62,049	100	300		Papua New Guinea (PNG)	IA	471,700	35	3500
	Franz Josef Land	OI	20,700	100	200		Japan	IA	372,313	60	2000
	Canary Islands	OI	7273	100	3500		Kamchatka	IA	350,000	70	900
	Cape Verde	OI	4033	100	250		Malaysia	IA	330,434	10	2500
	Azores	OI	2355	100	1300		Philippines	IA	300,000	70	3000
	Faroe Islands	OI	1400	100	1500		New Zealand	IA	268,046	40	2500
	Madeira	OI	800	100	700		Aleutian Islands	IA	150,000	100	2000
	Jan Mayen	OI	380	100	500		Sakhalin	IA	76,400	20	600
	Bear Island	OI	179	100	400		Taiwan	IA	36,174	60	3000
	St Helena	OI	122	100	900		Solomon Islands	IA	29,785	70	3000
	Tristan da Cunha	OI	104	100	1700		New Caledonia	IA	19,103	40	1200
	Ascension Island	OI	88	100	500		Kuril Islands	IA	15,600	100	1200
	Bermuda	OI	53	100	1500		Vanuatu	IA	14,760	70	3000
	Bouvet Island	OI	49	100	600		Small Indonesia islands	IA	9313	40	3000
	Jamaica	IA	10,991	5	2000		Guam	IA	549	70	2500
	Puerto Rico	IA	9104	100	1500		Northern Mariana Islands	IA	477	100	2100
	Trinidad and Tobago	IA	5128	5	2500		Palau	IA	458	70	3700
	Guadeloupe	IA	1438	60	3500		South China Sea Islands	IA	250	20	2000
	Martinique	IA	1079	100	3500		Fidji	OI	18,376	70	3000
	St Lucia	IA	619	100	2000		Hawaii	OI	16,705	100	2500
	Virgin Islands	IA	497	98	1000		Galapagos Islands	OI	7927	100	600
	Antigua and Barbuda	IA	442	40	1200		French Polynesia	OI	4200	75	2000
	St Vincent and Grenadines	IA	389	100	3000		Samoa	OI	2842	70	2900
	Leeward Islands	IA	353	100	1200		Kiribati	OI	811	100	2000
	Grenada	IA	344	100	3000		Tonga	OI	747	70	1700
	St Christopher and Nevis	IA	261	100	1500		Micronesia (Fed. States)	OI	702	70	4800
	Montserrat	IA	102	100	1500		Cook Islands	OI	240	70	2100
	Island volcanic rock surface area = 217,169 km^2						American Samoa	OI	199	70	3000
	Island surface area = 233,332 km^2						Marshall Islands	OI	181	70	3000
	Annual waterflux from volcanic islands = 283 km^3/a						Eastern	OI	162	100	1200
	Average rainfall (surf area weighted) = 1210 mm/a						U.S. Minor Outlying Islands	OI	34,5	70	1000
Indian	Kerguelen Islands	OI	7215	100	1800		Tuvalu	OI	26	70	3000
	Reunion Island	OI	2500	100	4000		Nauru	OI	21	70	2000
	Mauritius	OI	1865	100	2500		Island volcanic rock surface area = 1,551,670 km^2				
	Comoro Islands	OI	1862	100	3500		Island surface area = 3,531,785 km^2				
	Heard Island	OI	680	100	2500		Annual waterflux from volcanic islands = 9512 km^3/a				
	Crozet Islands	OI	500	100	2200		Average rainfall (surf area weighted) = 2690 mm/a				
	Prince Edward Islands	OI	316	100	3000						
	Rodrigues	OI	104	100	1500						
	1/2 Indonesia (without PNG)	IA	1,033,250	23	3300						
	Island volcanic rock surface area = 252,690 km^2										
	Island surface area = 1,048,292 km^2										
	Annual waterflux from volcanic islands = 3448 km^3/a										
	Average rainfall (surf area weighted) = 3290 mm/a										

Worldwide island volcanic rock surface area = 2.02 10^6 km^2.
Worldwide volcanic island surface area = 4.81 10^6 km^2.
Annual waterflux from volcanic islands worldwide = 13,243 km^3/a.
Mean annual rainfall on volcanic islands worldwide = 2800 mm/a.

relationship and the database in Table 1), from which we compute a discharge weighted average $(\varepsilon_{Nd})_{CC}$ for the rivers of -12.0 ± 1 (CC stands here for continental crust). The Nd input by dissolved loads from continental rivers thus cannot explain the seawater ε_{Nd} and we need a complementary mantle type input with positive ε_{Nd}. Since the attribution to ridge-crest hydrothermal input is precluded by empirical observations, weathering of volcanic islands is an opportune candidate, as already advocated or local considerations (Dia et al., 1992; Jeandel et al., 1998; Sholkovitz et al., 1999; Amakawa et al., 2000; Lacan and Jeandel, 2001; Goldstein and Hemming, 2003; Jones et al., 2008) but not at a global scale.

A simple global mass balance based on continental erosional input with an average $(\varepsilon_{Nd})_{CC}$ of -12 and volcanic island erosional component with a $(\varepsilon_{Nd})_{VI}$ of $+6$, suggests that 70% of oceanic Nd comes from the continents and 30% from volcanic islands. Thus the ratio of continental to volcanic Nd input is 2.3, similar to a ratio of 2 for Sr, an acceptable agreement in view of the approximations of the present calculations.

An examination of the regional distribution of ε_{Nd} among different oceans is also quite informative. Using the ε_{Nd} of the principal rivers worldwide (Table 1), we can tentatively calculate the erosional input for each ocean. We did the computation by weighing each river ε_{Nd} with its discharge and assuming that all rivers have the same Nd concentrations. Exceptions are the Arctic rivers, where we double the concentration because measured values appear to be very high (see Goldstein and Jacobsen, 1987). We obtain ε_{Nd} of -12.8, -3.0 and -8.0 for the Atlantic, Pacific and Indian oceans, respectively, very close to the measured average ε_{Nd} for each ocean (e.g. Arsouze et al., 2007). With the ε_{Nd} summarized in Fig. 3, we propose that continental erosion accounts for 96% of the Nd in the Atlantic ocean (vs 4% from volcanics), 52% in the Pacific ocean (vs 48%) and 73% in the Indian ocean (vs 27%).

For the Atlantic this is not surprising since the world largest rivers (Amazon, Orinoco, Parana, Congo, Niger, and St Lawrence) discharge into this ocean, islands arcs are relatively scarce (mainly Antilles and Sandwich islands), and the largest oceanic islands are in relatively low

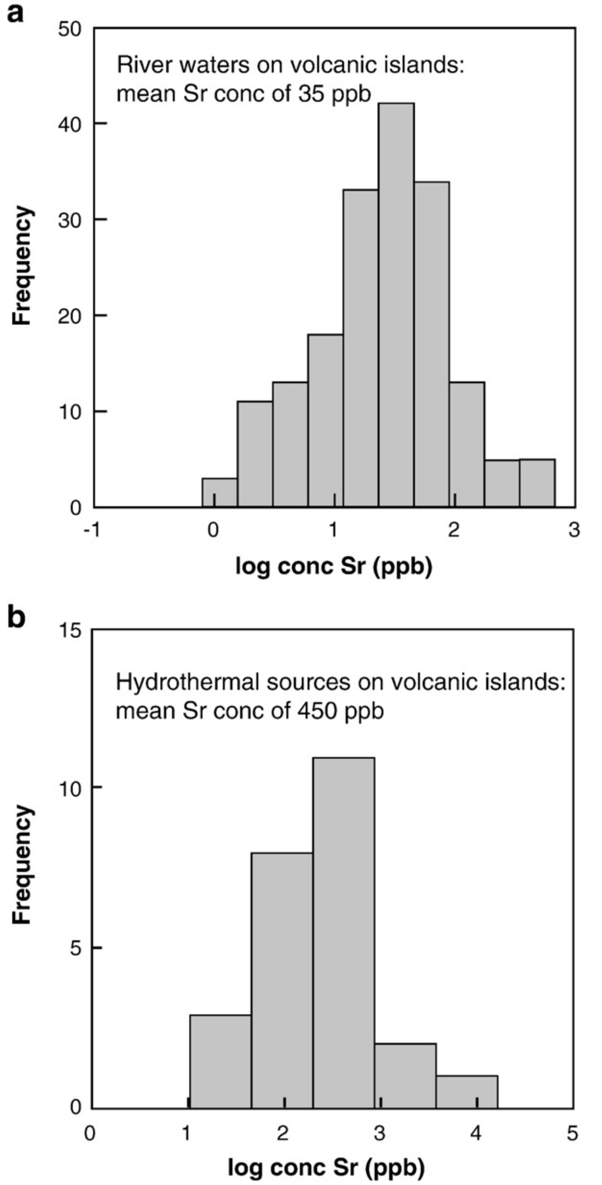

Fig. 2. Histograms of strontium concentrations in rivers (a) and hydrothermal sources (b) from volcanic islands: Iceland (Gislason et al., 1996), Réunion (Louvat and Allègre, 1997), Azores (Louvat and Allègre, 1998), Java (Louvat, 1997), Kamchatka (Dessert et al., 2007), Lesser Antilles (Rad et al., 2006). Sr concentrations follow a log-normal distribution centered around 35 ppb for rivers and 450 ppb for hydrothermal sources.

rainfall regions (Iceland, Svalbard, Franz Josef Land…). In the Pacific, on the other hand, the ε_{Nd} of seawater clearly reflects the presence of the largest active island arc systems of the world, from Kermadec to Tonga, Papua New Guinea, Philippines, Japan, all the way to Kamchatka. The Indian Ocean, with an average ε_{Nd} of -7.0 ± 2.7, is slightly more complex because it receives water from the Pacific Ocean via the Indonesian Gateway. However, it also receives large quantities of erosional products from the Himalayan Rivers (Ganges, Brahmapoutra, Irrawaddy, Salween…) with average ε_{Nd} of -13.3 ± 1) as well as from a portion of the very active Indonesian island arc that has rainfall among the largest in the world.

We contest the earlier assertion that the differences in ε_{Nd} among the oceans are due to differences in the ages of the continents that surround them. The average continental ε_{Nd} is almost the same for all oceans, or only slightly higher for the Pacific (Fig. 3) and the major difference is the relative distribution of island arcs among the oceans. It is the surface weathering of these and transport by surface and underground waters that controls the ε_{Nd} seawater mass balance of each ocean.

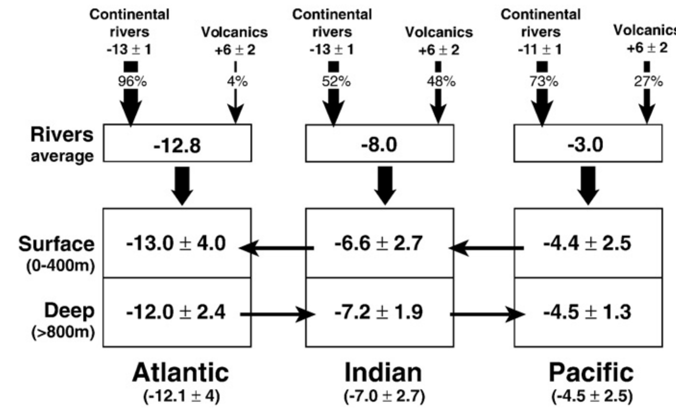

Fig. 3. Neodymium isotope budgets of the Atlantic, Indian and Pacific oceans. Surface, deep and average seawater ε_{Nd} are from the compilation of Arsouze et al. (2007). Average ε_{Nd} for major rivers are from Table 1. If continental river ε_{Nd} is -13 for the Atlantic and Indian Oceans and -11 for the Pacific Ocean, and if volcanic island weathering ε_{Nd} is $+6$, then 4%, 48% and 27% of the Nd comes from volcanic inputs in the Atlantic, Indian and Pacific Oceans. This reflects well the relative proportions of volcanic islands, and particularly volcanic island arcs, among the oceans.

Such island arc contribution is clearly evident from regional studies of Nd transport through the Indonesian strait and in the vicinity of Papua New Guinea (Jeandel et al, 1998; Sholkovitz et al., 1999; Amakawa et al., 2000; Lacan and Jeandel, 2001), a feature not strongly emphasized in the original papers.

5. Conclusion

Following studies by Rad et al. (2007) in Lesser Antilles and Réunion that documented the fundamental importance of underground water input to the oceans, we argue that volcanic islands are indeed a major source of Sr in seawater, a scenario similar to Nd isotopes budget where ridge-crest hydrothermal input is negligible. This approach enables us to refine the models that are based on $^{87}Sr/^{86}Sr$ evolution of seawater through geological time. During collision of two tectonic plates, island arcs are active prior to the continental collision but with the commencement of the actual collision continental erosion, linked with mountains building, increases while island arc erosion stops because the arcs are buried in the suture zone. Both effects tend to increase the seawater $^{87}Sr/^{86}Sr$ ratio during convergence and can be used to interpret $^{87}Sr/^{86}Sr$ record through the Phanerozoïc.

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