

The globally integrated view of Earth systems evolution from seawater isotope records: How does it work?

S.B. Jacobsen

Department of Earth and Planetary Sciences, Harvard University, 20 Oxford Street, Cambridge, MA 02138

Introduction

Major questions relating to the evolution of the Earth's oceans and atmosphere and sedimentary cycling can potentially be resolved by studying isotopic variations of seawater through time. Relatively complete isotopic records now exist for Sr and C and a less well constrained Nd isotope record exists for the past 800Ma. We have developed simple first order models to interpret these records (cf. Jacobsen, 1988; Keto and Jacobsen, 1988; Asmerom *et al.*, 1991, Derry *et al.*, 1992, Kaufman *et al.*, 1993) and here provide an overview of our modeling approach and most important results.

Constraints on paleocean evolution and the mean age of the continental material supplied to the oceans

Because of the short residence time of Nd in seawater ($\tau_{Nd} \sim 10^3$ years) large $^{143}\text{Nd}/^{144}\text{Nd}$ (expressed as ϵ_{Nd}) are often found between the various ocean basins through time. Smaller variations are found within single ocean basins. The isotopic composition of Nd in an ocean basin in general reflects whether it is predominantly surrounded by active (generally high ϵ_{Nd}) versus passive continental margins (generally low ϵ_{Nd}). A global average ϵ_{Nd} curve for the past 800 Ma has been constructed by averaging ϵ_{Nd} of various paleoceans. These results demonstrate large fluctuations of the global seawater ϵ_{Nd} value through this time period. The present Nd budget of the oceans is dominated by the river water (RW) flux from continental sources, while the mid-ocean ridge hydrothermal water (HW) Nd flux today contributes only about 1% of the total Nd input to the oceans. Thus variation of $^{143}\text{Nd}/^{144}\text{Nd}$ in seawater (SW) is primarily due to changes in the Nd-isotopic composition of the continental flux to the oceans and is therefore a proxy for the mean age of the continental material flux into the oceans.

Constraints on variations in the global erosion rate

Changes in $^{87}\text{Sr}/^{86}\text{Sr}$ (expressed as ϵ_{Sr}) of seawater are primarily controlled by changes in the river water flux (J_{Sr}^{RW}) (the erosional flux) of Sr from the continents and the hydrothermal flux (J_{Sr}^{HW}) of Sr through ocean ridges, as well as changes in the isotopic composition of these fluxes. In particular, variation in continental erosion due to tectonic events is often assumed to have a strong effect on the variation of $^{87}\text{Sr}/^{86}\text{Sr}$ in seawater. The mass balance for Sr isotopes, Nd and Sr isotopic variations in river waters, and the seawater Sr and Nd isotope curves can be used to constrain J_{Sr}^{RW} :

$$\left(\frac{J_{Sr}^{RW}}{J_{Sr}^{HW}}\right) = \left[\frac{(\epsilon_{Sr}^{HW} - \epsilon_{Sr}^{SW}) - \alpha\gamma(\epsilon_{Nd}^{HW} - \epsilon_{Nd}^{SW})}{\tau_{Sr} \left(\frac{d\epsilon_{Sr}^{SW}}{dt}\right) - (\alpha\epsilon_{Nd}^{SW} + \beta - \epsilon_{Sr}^{SW})} \right] \quad (1)$$

where $\gamma = (\text{Nd/Sr})_{HW}/(\text{Nd/Sr})_{RW}$ and τ_{Sr} (~ 4.1 Ma) = 0.09 is the residence time of Sr in the oceans. The ϵ_{Nd} and ϵ_{Sr} variations in present river waters roughly obey the following relationship: $\epsilon_{Sr}^{RW} \approx \alpha \epsilon_{Nd}^{RW} + \beta$ where $\alpha = 6.44$ and $\beta = 52$ are constants and are consistent with a present average $^{87}\text{Sr}/^{86}\text{Sr} = 0.711$ for the river water flux. Estimates of ϵ_{Nd} and ϵ_{Sr} in the hydrothermal component can be evaluated using the Sr- and Nd-isotopic evolution of the depleted mantle through time. Since the global dissolved flux of Sr in rivers is proportional to the global erosion rate, the J_{Sr}^{RW} value constrained this way is a proxy of erosion rates through time. We note that this does not necessarily imply that $^{87}\text{Sr}/^{86}\text{Sr}$ in seawater is a proxy for erosion rates. Most likely the $^{87}\text{Sr}/^{86}\text{Sr}$ of seawater is a function of both varying erosion rates as well as changes in the $^{87}\text{Sr}/^{86}\text{Sr}$ of the river flux. Model results indicate three distinct episodes of high global continental erosion rates due to uplift caused by continental collision at these times: ~ 0 Ma (Himalayan-Tibetan collision), ~ 0.4 Ga (Caledonian-Appalachian collision) and ~ 0.6 Ga (Pan-African collision) ago. This analysis

suggests that there were two times of very rapid increase in $^{87}\text{Sr}/^{86}\text{Sr}$ (0 Ga and 0.6 Ga), which appear to be caused by high erosion rates. The third event (~ 0.4 Ga) is, however, not associated with particularly high $^{87}\text{Sr}/^{86}\text{Sr}$. Thus, there is, in general, a limited correlation between the Sr isotope curve and high vs. low erosional fluxes. The idea that $^{87}\text{Sr}/^{86}\text{Sr}$ predominantly reflects variations in erosional fluxes from the continents thus appears incorrect. The processes operating during the Vendian and Cambrian periods resulted in the largest change observed in $^{87}\text{Sr}/^{86}\text{Sr}$ of seawater at any time during Earth history. The resulting $J_{\text{Sr}}^{\text{RW}}$ curve show a peak at ~ 585 Ma suggest that part, but not all of the change is the result of a change in $^{87}\text{Sr}/^{86}\text{Sr}$ of the continental flux. In contrast, the Cenozoic change of $J_{\text{Sr}}^{\text{RW}}$ yields a peak at ~ 5 Ma, appears to be associated with very little change in the Sr isotopic composition of the input to the oceans.

Constraints on variations in organic C burial

The variation of $\delta^{13}\text{C}$ of seawater (and marine carbonates) over timescales of crustal recycling is controlled both by changes in the global organic C erosion rate and by changes in the rate of C burial ($J_{\text{C}_{\text{org}}}^{\text{burial}}$). To couple the Sr and C cycles it is assumed that the erosion rate for Sr is proportional to that of organic C (i.e. $k_{\text{C}_{\text{org}}} = k_{\text{C}_{\text{org}}}^*$ ($J_{\text{Sr}}^{\text{RW}}/J_{\text{Sr}}^{\text{RW}*}$) where $k_{\text{C}_{\text{org}}}$ is the time parameter for erosion of organic C and the asterisk (*) denotes the present-day value). Assuming that the rate of recycling of carbonate rocks and sedimentary organic C is equal, it can then be shown that the burial rate of organic C shows a simple relationship to overall erosion rates and secular variations in $\delta^{13}\text{C}$:

$$J_{\text{C}_{\text{org}}}^{\text{burial}} = \frac{M_{\text{tc}}}{\Delta k_{\text{C}_{\text{org}}}^*} \left(\frac{J_{\text{Sr}}^{\text{RW}}}{J_{\text{Sr}}^{\text{RW}*}} \right) (\delta^{13}\text{C}_{\text{carb}} - \delta^{13}\text{C}_{\text{tc}}) \quad (2)$$

Here the $\delta^{13}\text{C}$ values of the carbonate ($\delta^{13}\text{C}_{\text{carb}}$) and organic ($\delta^{13}\text{C}_{\text{org}}$) C reservoirs are related by the total exogenic carbon isotope value of $\delta^{13}\text{C}_{\text{tc}} = -5.5$ and $\Delta = \delta^{13}\text{C}_{\text{carb}} - \delta^{13}\text{C}_{\text{org}}$ is considered to be reasonably constant at 28.5. Equations (1) and (2) provide the basis for obtaining changes in erosion ($J_{\text{Sr}}^{\text{RW}}$) and organic C burial rates ($J_{\text{C}_{\text{org}}}^{\text{burial}}$) as

a function of time based on the Sr-, C-, and Nd-isotopic records of seawater. The calculated rate of organic C burial depend on the product of the erosion rate and ($\delta^{13}\text{C}_{\text{carb}} - \delta^{13}\text{C}_{\text{tc}}$) values in carbonates. Thus it is possible to have relatively low $\delta^{13}\text{C}$ values (+1 to +2) in marine carbonates during periods of high organic carbon burial if the erosion rate is very high. We note that the $\delta^{13}\text{C}_{\text{carb}}$ is not a proxy for organic carbon burial rates while $J_{\text{C}_{\text{org}}}$ estimated from (2) is more likely to reflect real global variations in organic carbon burial rates. A curve for the burial rate of organic C constrained by the C- and Sr- isotopic variations shows a prominent peak at ~ 575 Ma. In the latest Proterozoic these high erosion rates, likely coupled with high organic productivity and anoxic bottom water conditions, contributed to a significant increase in the burial rate of organic C. The evolution of atmospheric O_2 is, in part, controlled by $J_{\text{C}_{\text{org}}}^{\text{burial}}$. The long-term accumulation of O_2 in the atmosphere is primarily due to inputs related to the reduction of C, Fe or S in the exogenic cycle and to burial of these in sediments. For each mole of C buried in sediments one mole of O_2 is released to the atmosphere. The Vendian peak in organic C burial coupled with lower fluxes of reducing hydrothermal fluids gave rise to a large increase in O_2 in the atmosphere after the Varanger glaciation. The peak in organic C burial at this time is sufficient to generate most of the O_2 in the present atmosphere. Several peaks in ($J_{\text{C}_{\text{org}}}^{\text{burial}}$) inferred for the Phanerozoic need not cause large changes in atmospheric O_2 levels as they may be balanced by the S-cycle.

References

- [1] Jacobsen, S.B., (1988) *Earth Planet. Sci. Lett.*, **90**, 315–29.
- [2] Keto L.S. and Jacobsen S.B. (1988) *Earth Planet. Sci. Lett.*, **90**, 315–29.
- [3] Asmerom, Y., Jacobsen, S.B., Butterfield N.J., and Knoll A.H., (1991) *Geochim. Cosmochim. Acta*, **55**, 2883–94.
- [4] Derry L.A., Kaufman A.J., and Jacobsen S.B., (1992) *Geochim. Cosmochim. Acta*, **56**, 1317–29.
- [5] Kaufman A.J., Jacobsen, S.B. and Knoll, A.H. (1993) *Earth Planet. Sci. Lett.*, **120**, 409–30.