

Interactive comment on “Sediment $^{231}\text{Pa}/^{230}\text{Th}$ as a recorder of the rate of the Atlantic meridional overturning circulation: insights from a 2-D model” by Y. Luo et al.

Reply to R. Anderson’s comments

“Most important, the model has no representation of lateral mixing. As such, the model tends to overestimate lateral concentration gradients that are otherwise reduced by lateral mixing”

There is already an inherent mixing in our model, resulting from the upwind scheme that we use and the grid size (horizontal grid spacing = 2.5° latitude; vertical grid spacing = 250 m). We can calculate the horizontal diffusivity (K_{diff}) inherent to the model as follows:

Assume a simplified steady-state advective-reactive equation of the form

$$u \frac{dC}{dx} = -kC$$

which has an analytical solution $C = C_o \exp(-kx/u)$ for a constant flow u and a constant k . If we expand the derivative using the upwind scheme, we find however, that the solution has modified exponential term and that the upwind approximation to the simplified advective-reactive equation is equivalent to the solution of a diffusive-advective-reactive equation:

$$u \frac{dC}{dx} = -kC + K \frac{d^2C}{dx^2}$$

where the diffusivity $K = u \Delta x / 2$.

u is the horizontal (meridional) velocity:

$$\text{Flow: } 4 \cdot 10^6 \text{ m}^3/\text{sec}$$

$$\text{Cross section: } 250 \text{ m} \times 3,000,000 \text{ m} = 7.5 \cdot 10^8 \text{ m}^2$$

$$\rightarrow u = 5.3 \cdot 10^{-3} \text{ m/s}$$

Δx is the horizontal grid spacing = $2 \times \pi \times 6.378 \cdot 10^3 \times 360 / 2.5 = 278 \cdot 10^3 \text{ m}$

$$\rightarrow K_{\text{diff}} = 5.3 \cdot 10^{-3} \times 278 \cdot 10^3 / 2 = 737 \text{ m}^2/\text{s}$$

This is in the upper range of the along-isopycnal tracer diffusivities reported for the southern ocean (100-800 m^2/s ; Zika et al., 2009). Therefore, with our grid size, the upwind scheme is producing an inherent diffusivity which is comparable to that found in the ocean and there is no need to add an additional diffusive term in our model. Initial tests with smaller grid sizes to lower the model’s diffusivity within the mid range of the reported oceanic values showed little difference in the model’s results.

“For example, Chase et al. (2003; cited by Luo et al) showed that there is no detectable meridional gradient in the concentration of ^{231}Pa or of ^{230}Th across the “opal belt” of the Southern Ocean when data are compared along isopycnals, indicating that lateral

mixing smoothes out any concentration minima associated with enhanced scavenging under the region of high diatom production”

If we follow the evolution of ^{230}Th and ^{231}Pa activity in our model along flow lines across the opal belt (from 42.5°S to 57.5°S ; Fig. 1), we find small variations in activity that are not much larger than those found by Chase et al. (2003). This further confirms that our simple model captures reasonably well the spatial distribution of ^{230}Th and ^{231}Pa in the Atlantic water column.

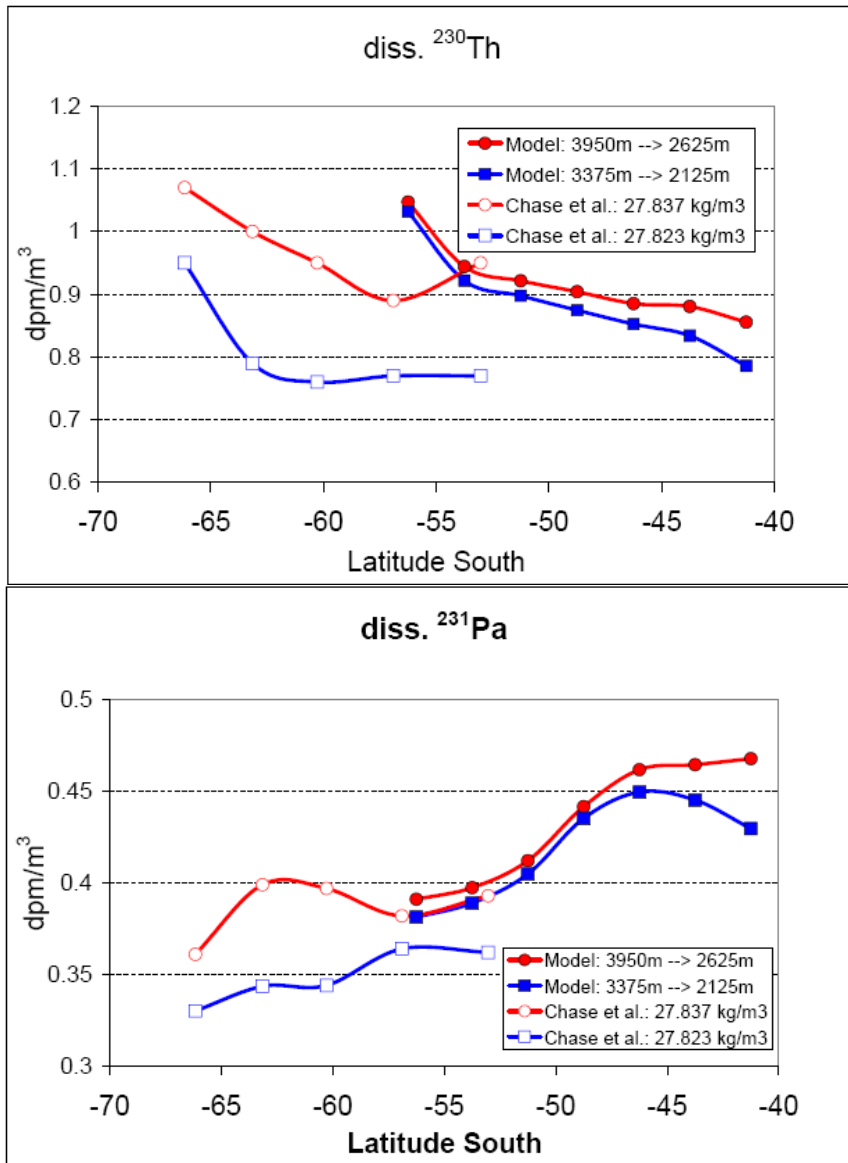


Fig. 1: Latitudinal variations in dissolved ^{230}Th and ^{231}Pa following two flow lines in the model (starting from a depth of 3375 m and 3950 m at 42°S and upwelling to depths of 2125 m and 2625 m, respectively). Also shown are the data from Chase et al. (2003) for two different isopycnals. The shift in latitude reflects the difference in the position of the opal belt in the model and in the SW Pacific sector of the southern ocean.

“Similarly, I pointed out at the GEOTRACES model-data synergy workshop, attended by Luo, that the dissolved Pa/Th ratio in deep water exhibits no detectable change along the flow path of the “deep conveyor”, from the North Atlantic to the North Pacific Ocean. This, too, must reflect a strong influence of lateral mixing on deep-sea dissolved Th and Pa distributions”

We would only expect a gradual increase in dissolved Pa/Th in deep water from the Atlantic to the Pacific if circulation between these two oceans could be described by a simple overturning cell, which is clearly not the case. Deep Pacific Water comes from the Southern Ocean, not directly from the Atlantic. Existing data and Fig. 9.g in our paper indicate that dissolved Pa/Th drops to lower values (< 0.4) in the southern ocean due to the presence of opal. At station ALOHA dissolved Pa/Th increases to values > 0.4 between 1000 and 3000m depth (data unpublished). The differences in Pa/Th between the southern ocean and the North Pacific can also be explained in terms of deep water circulation. We are preparing a manuscript that discusses these observations.

“My strongest recommendation for changes to the manuscript of Luo et al. prior to publication is to add text discussing how the model results would change if lateral mixing were included”

A discussion in the inherent diffusivity of our model has been added to the paper

DETAILED COMMENTS:

1) p. 2757, lines 15-16: Clarify here that the statement about sedimentary Pa/Th ratios being about half the production ratio (0.093) refers to the average for the North Atlantic Ocean. Sedimentary Pa/Th ratios at individual N Atlantic sites vary from approximately 0.03 to 0.14.

Done

2) p. 2761, eqn. 4, and elsewhere: Clarify that the representation of the fractionation factor (F) expressed here is numerically equivalent to the original definition by Anderson et al., (1983; EPSL V62, pp. 7-23; EPSL V66, pp. 73-90), although the arrangement of the terms is different.

Done

3) p. 2761, line 27: Doesn't the expression for $[X]_p/[X]_d$ require a term for the concentration of particles?

As per equation (1) in our paper:

Adsorption rates = $K_1[X]_d$

Desorption rates = $K_{-1}[X]_p$

Therefore, when particles are in equilibrium with seawater: $K_1[X]_d = K_{-1}[X]_p$

4) p. 2765, lines 3-4: See comment 2 above in referring to the original definition of F as well as to document the original measurements of F .

Done

5) p. 2768, line 16: *The modeled fraction of Pa associated with particles, described here, seems a little high for two reasons. First, although they are not yet published, results from the GEOTRACES intercalibration exercise for a depth of 2000 m at the Bermuda/BATS site suggest that particulate ^{231}Pa represents about 1% of the total, not 4 to 5% as indicated here. Second, if about 20% of the ^{230}Th is associated with particles and F (fractionation factor) is about 10 then one expects about 2% of the ^{231}Pa to be associated with particles.*

The latter calculation is valid only if particles are in equilibrium with seawater, which is not the case for sinking particles. Decreasing the fraction of ^{231}Pa associated with particles in our model would require decreasing the Pa adsorption rate constant which would produce water column profiles that would deviate significantly from those measured.

6) p. 2768, line 21, and Figure 11: Note that in the caption of Figure 11, “western” should be changed to “eastern” when referring to Scholten’s data.

Done

8) p. 2771, line 6: change “that” to “than”

Done

9) p. 2771, line 13 and Figure 13: The units in Figure 13B are illegible in my copy. Please ensure that all graphics are legible

We have modified Fig. 13b

11) p. 2772, lines 15-17: *I think the statement that the residence time increases linearly with depth is incorrect, because the definition of residence time is incorrect. Residence time (t) should be defined as [concentration]/(total rate of supply). Here, the authors define residence time as $t = [\text{concentration}]/(\text{production by } U \text{ decay})$. However, desorption from sinking particles contributes to the overall source too, so residence time should be expressed as: $t = [\text{concentration}]/([\text{production}] + [\text{desorption}])$. If that formulation is used, then the residence time remains constant with depth rather than increasing with depth.*

We clearly need to better define what we mean by “residence time”.

The lateral transport along isopycnals of water with low ^{230}Th or ^{231}Pa concentration (i.e. lower than the steady state concentration at that depth with respect to scavenging)

can be described by adding a lateral transport term to the original scavenging model (Rutgers van der Loeff and Berger, 1993).

$$\partial[X]_t/\partial t = P_X - S \partial(K[X]_t)/\partial Z + ({}^i[X]_t - [X]_t)/\tau_w = 0$$

Where P_X is the production rate, ${}^i[X]_t$ and $[X]_t$ are total ${}^{230}\text{Th}$ or ${}^{231}\text{Pa}$ concentration measured at two locations on the same isopycnal with ${}^i[X]_t$ the concentration in the upstream source region and τ_w is the “transit time” of water between these two sites. In deep waters, K ($[{}^{230}\text{X}]_p/[X]_t$) is nearly constant and to a first approximation can be removed from the derivative. Integrating this simplified equation thus gives:

$$[X]_t \cong (P_X \tau_w + {}^i[X]_t) (1 - e^{-Z/\tau_w SK})$$

This equation predicts that, as one moves downstream (increasing τ_w) the radioisotope profiles relax back to linearity slower as Z/SK increases. Since S and K change little with depth, the transit time required for relaxation of the profiles increases with depth, i.e. linearity is regained closer to the source at shallower depth

P_X/SK is the slope of the $[X]_t$ profile predicted by the reversed scavenging model in the absence of circulation or mixing (assuming a constant K).

$$[X]_t \cong P_X Z / SK$$

Z/SK is the residence time with respect to addition by uranium decay and removal by scavenging when the profile has regained linearity (i.e. steady state with respect scavenging), defined as:

$$\tau_{ss} = [X]_t / P_X$$

Then,

$$\tau_{ss} = Z / SK$$

Which increases linearly with depth.

Therefore:

$$[X]_t \cong (P_X \tau_w + {}^i[X]_t) (1 - e^{-\tau_{ss}/\tau_w})$$

The ${}^{230}\text{Th}$ and ${}^{231}\text{Pa}$ profiles relax back to linearity more slowly with increasing τ_{ss} and water depth ($\tau_{ss} = Z/SK$). Profile linearity is thus regained closer to the source at shallower depths (Fig. 2), and ${}^{230}\text{Th}$ regains linearity faster than ${}^{231}\text{Pa}$ because of its shorter τ_{ss} . For instances, at 4500m depth, ${}^{230}\text{Th}$ virtually regains its steady state linear profile within ~ 200 years, while it takes ~ 1000 years for ${}^{231}\text{Pa}$ to reach the same stage. This explains why we never find linear profiles for ${}^{231}\text{Pa}$, since it necessitates a water column that has been vertically stable for a period equivalent to the mixing time of the ocean.

We have modified the text in the paper to more clearly define “residence time”, now referred to as “ τ_{ss} ”.

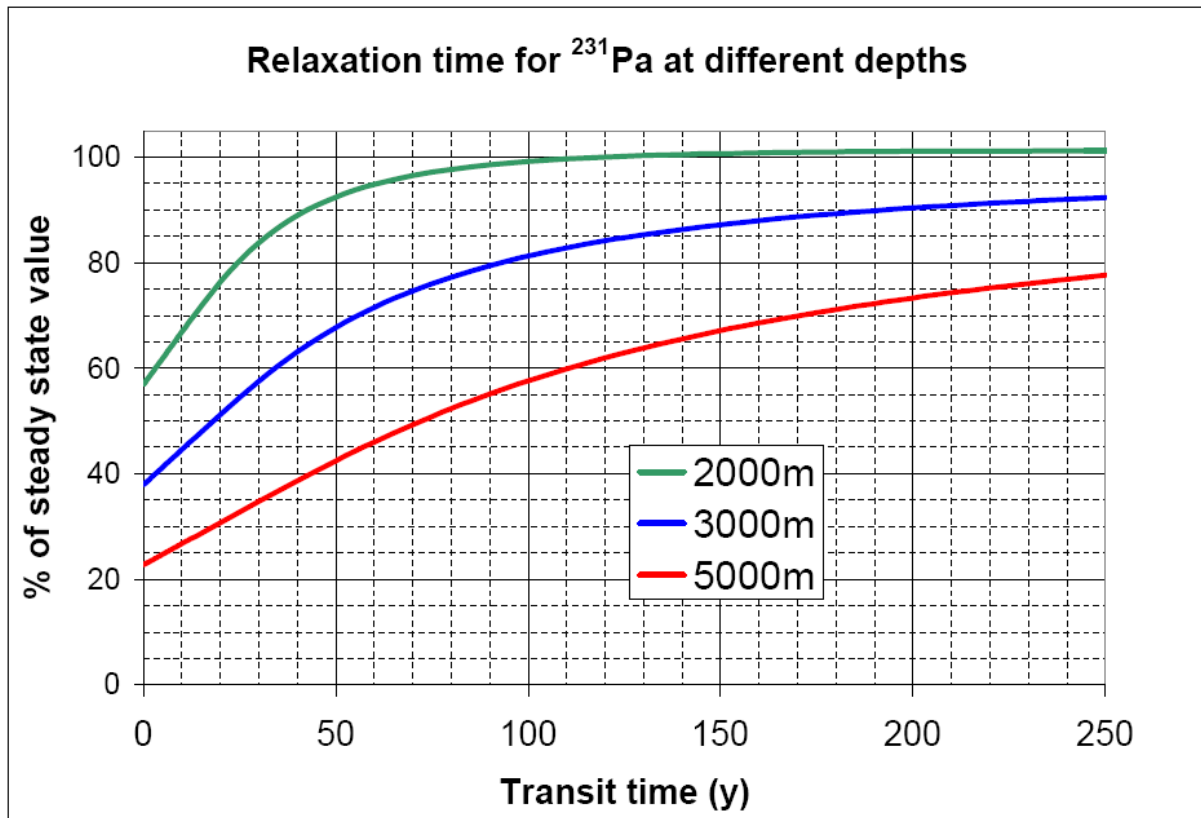


Fig. 1: Relaxation of ^{231}Pa activity at different depths as described by equation (7) with $^i[\text{X}]_t = 0.14 \text{ dpm}\cdot\text{m}^{-3}$, $S = 500\text{m}$, and $K = 0.04$. After 50 years of transit from the upstream source region, $^{231}\text{Pa}_t$ has reached 92.5% of its steady state value with respect to scavenging at 2000m. It takes increasingly longer transit times to reach a similar level of concentration (relative to the steady state value) in deeper water (250 years at 3000 m; 850 years at 5000 m).

12) p. 2772, lines 20-21: *Here the authors infer that dissolved Pa/Th ratios decrease with increasing water depth due to the lateral advection of ventilated deep water from the high-latitude source in the North Atlantic. However, dissolved Pa/Th ratios also decrease with depth in the North Pacific Ocean, where there is no comparable lateral advection of ventilated deep water. While lateral advection of ventilated deep water may contribute to the trend of decreasing dissolved Pa/Th ratios with increasing water depth, results from the N Pacific indicate that other factors may be involved as well, and this should be acknowledged.*

When we shut off circulation in our model, dissolved Pa/Th decreases with depth from 0.97 to 0.76. However, most of the decrease (from 0.97 to 0.82) occurs in the upper 1000m. The sharp decrease observed below 1000m in our model is almost entirely due to the overturning circulation. The reason for the decrease in absence of circulation is the presence of sinking particles that are out of equilibrium with surrounding seawater. Since ^{230}Th is scavenged faster, Pa/Th is relatively high in surface waters. Deeper in the water column, Pa “catches up” with Th and the changes in dissolved Pa/Th with depth become less pronounced. The decrease in dissolved Pa/Th found in the North Pacific (particularly

at station ALOHA) is much more pronounced than what we observe in the model without circulation. We believe that the sharp decrease also reflects the formation of Pacific Deep Water and we have a manuscript in preparation addressing this question.

13) p. 2773, line 11: change “a” to “an”

Done

15) p. 2775, all of Section 7.1.5: *Would these conclusions be changed if one considers lateral mixing? See general comments at the beginning of this review.*

Our model has inherent lateral mixing which is in the upper range of the diffusivity measured in the ocean (see above). Decreasing mixing by decreasing grid size does not significantly affect the results. Therefore, we don't think that further consideration of lateral mixing would change these conclusions

16) p. 2777, line 1: *“opposite effect” on what? The meaning of the comparison here is unclear.*

Reducing the fractionation factor in the southern ocean increases the Southern Ocean ^{231}Pa sink and decreases $^{231}\text{Pa}/^{230}\text{Th}$ in the Atlantic. This has now been clarified in the manuscript