Shallow particulate organic carbon regeneration in the South Pacific Ocean

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Particulate organic carbon (POC) produced in the surface ocean sinks through the water column and is respired at depth, acting as a primary vector sequestering carbon in the abyssal ocean. Atmospheric carbon dioxide levels are sensitive to the length (depth) scale over which respiration converts POC back to inorganic carbon, because shallower waters exchange with the atmosphere more rapidly than deeper ones. However, estimates of this carbon regeneration length scale and its spatiotemporal variability are limited, hindering the ability to characterize its sensitivity to environmental conditions. Here, we present a zonal section of POC fluxes at high vertical and spatial resolution from the GEOTRACES GP16 transect in the eastern tropical South Pacific, based on normalization to the radiogenic thorium isotope [230]Th. We find shallower carbon regeneration length scales than previous estimates for the oligotrophic South Pacific gyre, indicating less efficient carbon transfer to the deep ocean. Carbon regeneration is strongly inhibited within suboxic waters near the Peru coast. Canonical Martin curve power laws inadequately capture POC flux profiles at suboxic stations. We instead fit these profiles using an exponential function with flux preserved at depth, finding shallow regeneration but high POC sequestration below 1,000 m. Both regeneration length scales and POC flux at depth closely track the depths at which oxygen concentrations approach zero. Our findings imply that climate warming will result in reduced ocean carbon storage due to expanding oligotrophic gyres, but opposing effects on ocean carbon storage from expanding suboxic waters will require modeling and future work to disentangle.

biological pump | ocean carbon storage | oxygen-deficient zones | GEOTRACES | thorium

The oceanic biological pump encompasses a series of processes by which phytoplankton at the sea surface photosynthetically fix carbon dioxide (CO_2) to form particulate organic carbon (POC), a portion of which is exported from the upper ocean and sinks to depth, where it is regenerated by microbial respiration (1, 2). The first two components of the biological pump, primary production and export of POC from the upper ocean, have been sufficiently characterized to enable their parametrization in terms of variables that can be measured by satellites, allowing for comprehensive estimates of their global rates and spatiotemporal variability (3–6). However, the fate of exported POC upon sinking into the ocean interior has proved to be an elusive oceanographic target. Because the time scale that waters are sequestered from the atmosphere increases with depth, the length scale over which POC regeneration occurs exerts a strong control on oceanic carbon storage and atmospheric CO_2 levels (7). Consequently, assessing how environmental conditions influence POC regeneration length scales provides critical insights that can be incorporated into ocean carbon cycle models to improve projections of future oceanic CO_2 uptake, including the response to global warming.

Historical estimates of carbon regeneration in the ocean interior have come from POC flux profiles generated either by compilations of sediment traps (8); by individual free-floating sediment trap profiles, typically with three to six depths in the upper 500 m (9, 10); or by combining [234]Th-based euphotic zone POC fluxes with those from bottom-moored sediment traps below 1,500 m (11, 12). POC regeneration length scales are then determined by fitting either power laws (8) or exponential functions (13) to the vertical profiles of POC flux. However, these methods are respectively limited by their spatial resolution, vertical resolution, and integration across different temporal and spatial domains. The methods also provide conflicting results on the spatial patterns of regeneration depths, precluding the development of a comprehensive mechanistic understanding of the processes that control POC regeneration (9, 11).

We determine POC regeneration length scales in the eastern tropical South Pacific by adapting the paleoceanographic [230]Th-normalization method (14) to the water column. Our study is the first application of this approach to generate internally consistent, high-resolution POC flux profiles that resolve differences in POC flux characteristics across biogeochemical gradients on annual to multiannual time scales. By analyzing particulate [230]Th ([230]Th$_p$) and POC collected by in situ filtration, we calculate POC fluxes, integrated across ~1- to 3-y time scales, at each measurement depth (15, 16) (see *Materials and Methods*). A

recent intercomparison of sediment trap and radiochemical methods at the Bermuda Atlantic Time-Series Station found that ^{230}Th$_p$-normalized POC fluxes agreed (within 2-σ uncertainty) with other radiochemical methods for estimating POC flux in the upper water column (17). In further support of this approach, we find that ^{230}Th$_p$-derived POC fluxes on the GEOTRACES GP16 transect are within uncertainty of nearby annually averaged sediment trap POC fluxes (*SI Appendix*, Fig. S1).

Samples were collected on the GP16 transect (Research Vessel *Thomas G. Thompson*, cruise TN303) spanning from Peru to Tahiti (Fig. 1). The GP16 section traversed a strong zonal gradient in upper water column conditions, particularly in productivity and subsurface O_2 (Fig. 1). The Peru oxygen-deficient zone (ODZ) in the eastern portion of the section hosts nanomolar to subnanomolar O_2 levels, making it functionally anoxic (18). Oxygen concentration minima from GP16 were below the detection limit of 1 μmol/kg at stations 1 to 13 (Fig. 1*B*). Pigment and fluorescence data indicate that there is a transition in microbial community structure moving offshore within the ODZ, from autotrophic at station 9 to heterotrophic at station 11 (19). Our ^{230}Th$_p$-normalized POC flux profiles have sufficient vertical resolution to provide statistically significant constraints on the spatial variability and mechanisms controlling POC regeneration length scales and carbon transfer to the deep ocean across the sharp biogeochemical gradients spanning from the Peru ODZ to the highly oligotrophic South Pacific subtropical gyre (SPSG).

Results and Discussion

GP16 ^{230}Th$_p$-normalized POC flux profiles have highest values in the subsurface near the deep chlorophyll maximum and base of the mixed layer (*SI Appendix*, Fig. S2), and decrease subsequently with depth. Maximum POC fluxes of 5 mmol·m^{-2}·d^{-1} are found nearest to the continental shelf in the Peru upwelling region, decreasing to 2 mmol·m^{-2}·d^{-1} in the oligotrophic South Pacific gyre (Fig. 2). The greatest flux attenuation occurs in the upper 300 m of the water column, indicative of shallow POC regeneration. At stations 1 to 9, corresponding to the Peru ODZ, POC flux decreases rapidly through the upper oxycline, stays constant through the depths of lowest oxygen, then decreases again through the lower oxycline (*SI Appendix*, Fig. S3). The lack of POC flux attenuation within the ODZ must reflect negligible regeneration of vertically sourced POC supply from above (see *SI Appendix, Supplementary Information Text*).

Regeneration length scales are traditionally expressed using a power law relationship (8). We fit power laws of form $F_z = F_{z_0} \left(\frac{z}{z_0} \right)^{-b}$ to POC fluxes in the upper 1,000 m at each station to predict the POC flux (F_z) at depth z, relative to a reference depth z_0, with the exponent b describing the rate of flux attenuation with depth. Stations 15 to 36, west of the Peru upwelling region, have average b values of 1.29 ± 0.12 (Fig. 3), much higher (i.e., shallower, faster regeneration) than previous estimates for the SPSG derived from the combination of bottom-moored sediment traps and ^{234}Th ($b = 0.52$) (11, 12) and a follow-up
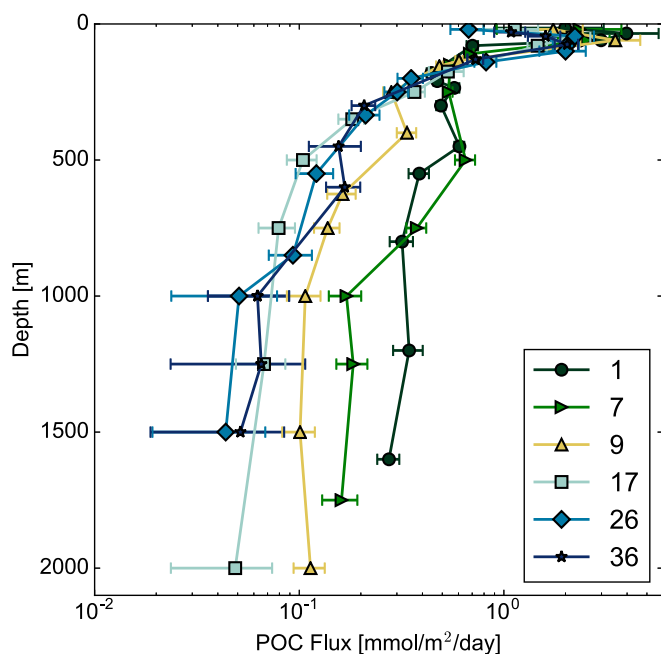

Fig. 1. Cruise track and dissolved oxygen concentrations. (*A*) TN303 cruise track showing station locations and 2013 annually averaged moderate resolution imaging spectroradiometer satellite-derived net primary productivity (NPP) from the vertically generalized productivity model (3). Locations of historical sediment trap deployments mentioned in the text are shown as stars. (*B*) Dissolved oxygen concentrations (μmol/kg) on the GP16 section.

Fig. 2. $^{230}Th_p$-normalized POC fluxes from six representative GP16 stations. The highest POC fluxes at the surface and subsurface are found at stations 1 to 9, closest to the Peru coast. Stations 17, 26, and 36, representative of the SPSG, have lower POC fluxes.

approach that included constraints from particle imaging ($b = 0.84$) (12). Our results are much more consistent with neutrally buoyant sediment trap deployments in the North Pacific subtropical gyre ($b = 1.33 \pm 0.15$), which had greater resolution through the upper water column depths at which the bulk of POC regeneration occurs (10). Transfer efficiencies derived from b values fitted to bottom-moored sediment trap observations may not be an ideal benchmark for evaluating biogeochemical model representations of POC flux and regeneration in the mesopelagic, as previously suggested (9, 20).

POC fluxes are constant with depth within the suboxic waters of the Peru ODZ (Fig. 4A). Correlations of POC flux with depth from 60 to 600 m show that the stations within the ODZ have no statistically significant decrease in POC flux with depth (Fig. 4A), while stations with no suboxic waters have highly statistically significant ($P < 10^{-10}$) decreasing POC flux in the same depth range (Fig. 4B). Stations 1 and 7 in the Peru ODZ have lower b values of 0.74 ± 0.15 and 0.66 ± 0.18, respectively, compared with a range of b values from 1.11 to 1.52 at oxic stations 15 to 36 (Fig. 3). Previous studies have also found low b values for POC flux profiles from sediment trap deployments in ODZs, attributed to greater POC preservation under low oxygen conditions (21–23). However, the goodness of fit for power laws at stations 1 to 7, where the POC flux at depth is greatest and the top of the ODZ is shallowest, was much lower than at the offshore stations (*SI Appendix*, Fig. S4). The residuals of the power-law fits to stations 1 to 7 are also correlated with depth (rho = −0.66, $P = 0.002$), indicating that a power law fails to adequately capture the functional form of POC flux profiles at ODZ stations.

We instead fit POC flux profiles using an exponential function (13, 24), $F_z = F_{z_0} \exp\left(\frac{-z}{L}\right) + F_\infty$, including an asymptotic flux F_∞ preserved as depth approaches infinity to quantify the effect of the ODZ on POC regeneration length scale (L) and transfer to the deep ocean. Unlike the power law, the residuals of exponential fits at ODZ stations are not significantly correlated with depth (rho = −0.18, $P = 0.40$). We used the exponential fits to generate bootstrapped probability distribution functions for L and F_∞ for suboxic and oxic station groupings. The distributions are nearly disjoint, with suboxic stations having both shallower regeneration (Fig. 4C) and nearly 4 times more carbon flux preserved into the deep ocean (Fig. 4D) than oxic stations. This is not simply a consequence of larger export fluxes at suboxic stations. The transfer efficiency, computed as the best-fit F_∞/F_{max}, where F_{max} is the maximum POC flux at each station, is 2 to 5 times higher at suboxic stations 5, 1, and 7 than at oxic stations (*SI Appendix*, Fig. S5). Thus, the flux profiles at suboxic stations are qualitatively and significantly different from those at oxic stations. We argue that the low b values previously inferred in ODZs overlook the importance of shallow POC regeneration in
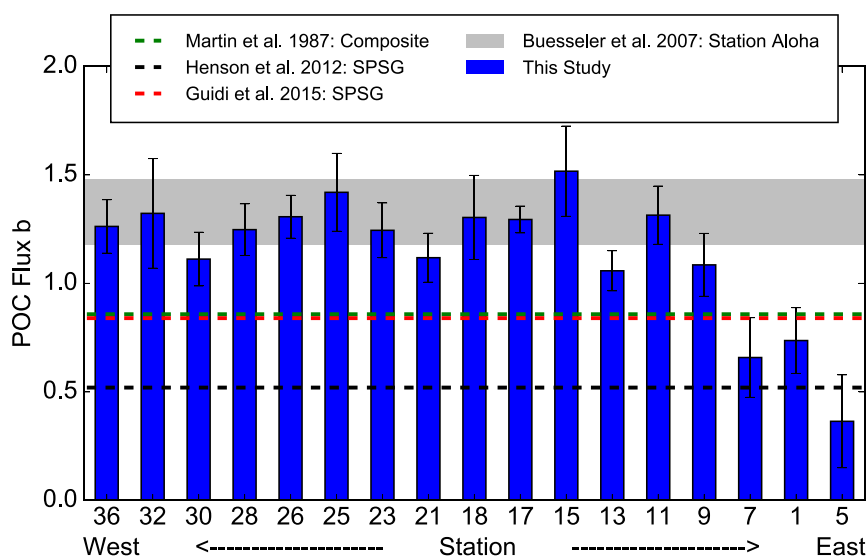

Fig. 3. Power-law b values for GP16 POC fluxes. Values from the GP16 section (blue bars, this study) are compared with a composite estimate from the North Pacific (green dashed line) (8), a neutrally buoyant sediment trap deployment from Station ALOHA in the North Pacific subtropical gyre (gray band) (10), and estimates for the SPSG derived either from combining ^{234}Th with deep-moored sediment traps (black dashed line) (11) or from particle imaging (red dashed line) (12).

Fig. 4. Statistical analysis of POC fluxes in suboxic and oxic station (St.) groupings. (*A*) POC fluxes between 60 and 600 m from suboxic stations, with correlation coefficients and *P* values for Pearson, Spearman, and Kendall tau correlation tests for POC flux with depth. (*B*) POC fluxes and correlation coefficients from oxic stations. The depth range was chosen to be broadly representative of the depths of suboxic waters in order to test for the presence of statistically significant POC regeneration at suboxic stations compared with same depths in oxic waters. The POC fluxes at suboxic stations shown here mostly come from below the depths of intense POC regeneration through the upper oxycline. (*C*) Bootstrapped probability distributions of flux preserved as depth approaches infinity from fitting an exponential model to GP16 POC flux data at suboxic and oxic station groupings (see *Materials and Methods*). (*D*) Bootstrapped probability distributions of regeneration length scale from exponential model fits to GP16 POC flux data at suboxic and oxic station groupings.

the upper oxycline. In addition to its implications for ocean carbon storage, accurately representing the vertical pattern of POC flux and regeneration in ODZs is critical for determining the depth distribution and magnitude of nitrogen loss processes (anammox and denitrification) in ODZs, which are directly linked to the supply, regeneration, and stoichiometry of organic matter flux (25).

Our results have important implications for feedbacks in the global carbon cycle under future climate change. The oligotrophic subtropical gyres are projected to expand due to increased vertical stratification (26, 27). The efficiency of POC transport to the deep ocean in the subtropics has been debated, with bottom-moored sediment trap observations suggesting efficient subtropical C storage (11), but inverse modeling (20) and neutrally buoyant sediment trap results (9, 10) suggesting the opposite. Shallow subtropical POC regeneration inferred from $^{230}Th_p$ normalization in both the North Atlantic gyre (17) and the South Pacific (this study) are consistent with inefficient carbon storage in the oligotrophic ocean. Thus, gyre expansion from CO_2 warming is predicted to drive a positive feedback involving shallower carbon regeneration and less efficient carbon sequestration in the deep ocean.

ODZs are expected to expand in area and shoal under climate warming (28, 29); however, the relative importance of increased respiration and decreased ventilation is unknown (30, 31). Both L and F_∞ are well correlated with the depth of the upper oxycline (19) (Fig. 5), indicating that changes in ODZ extent and POC regeneration will be intimately coupled. Our results show that regeneration dynamics in ODZ regions have two potentially very large, offsetting effects on ocean carbon storage. Shallower regeneration length scales will return respired CO_2 to the atmosphere more quickly, but greater POC preservation to depths below 1,000 m will result in greater abyssal carbon storage. The expansion and shoaling of ODZs, therefore, will not necessarily result in enhanced overall ocean carbon storage, as previously proposed (22). The feedbacks between ODZ expansion and ocean carbon storage will require the implementation of more flexible and spatially variable regeneration length scales in global carbon cycle models and should be a high priority target for future study.

Materials and Methods

Particulate Sample Collection. Particulate samples on the GP16 section were collected via in situ filtration using McLane pumps (WTS-LV) with two flow paths. Each flow path was equipped with a 142-mm-diameter filter holder containing baffles to ensure homogenous particle distributions on the filters (32). The holders both had a 51-μm Sefar Polyester mesh prefilter, followed by either paired acid-leached, precombusted quartz-fiber Whatman QM-A filters with a 1-μm pore size, or acid-leached paired Pall Supor800 0.8-μm polyethersulfone filters (33).

Blank filters were simultaneously deployed with the pumps on each cast, either on specially adapted filter holders disconnected from pumped water flow or in polypropylene containers zip-tied to the frame of a pump. The blank filters were in contact with ambient seawater at pump depth for the entire cast. These dipped blanks were used for background corrections of POC and Th isotopes. Previous publications (33, 34) have provided more detailed documentation of the collection of in situ pumped particles on the TN303 cruise.

Sample Analysis. Measurement techniques for dissolved oxygen (35), POC (33), and Th isotopes (36) on the GP16 section have been previously documented. We provide here a brief overview containing the salient details of the measurements techniques, but refer readers to the publications containing the original data for complete methods.

Dissolved oxygen was determined via modified Winkler titration, according to standard procedures established in the WOCE, CLIVAR, and GO-SHIP Repeat Hydrography programs (https://www.go-ship.org/Hydro-Man.html). The detection limit for discrete oxygen samples was 1 μmol/L, with ∼0.1% precision (35). The finalized oxygen dataset is archived online at the Biological and Chemical Oceanography Data Management Office (BCO-DMO, https://www.bco-dmo.org/dataset/503145), as well as the GEOTRACES Intermediate Data Product (37).

POC in the 0.8- to 51-μm small-size fraction (SSF) was measured on two 12-mm-diameter punches taken from the top QM-A filter, representing ∼20 L of pumped seawater. The filters were dried at sea, fumed with concentrated hydrochloric acid (HCl) to remove inorganic carbon, and then dried again before the punches were taken. SSF POC was measured using a FlashEA 1112 Carbon/Nitrogen Analyzer using a Dynamic Flash Combustion technique. Dipped blank QM-A filters (n = 47) were used for blank subtraction, and the SD of the dipped blank measurements was assigned as the uncertainty for SSF POC measurements. POC data are available online at BCO-DMO (https://www.bco-dmo.org/dataset/668083) and the GEOTRACES Intermediate Data Product (37).

Particulate ^{230}Th ($^{230}Th_p$) was measured in two laboratories: Lamont-Doherty Earth Observatory (LDEO), and University of Minnesota (UMN). Intercalibration showed no detectable differences between the methods of the two laboratories. At LDEO, one-fourth–filter aliquots were placed in
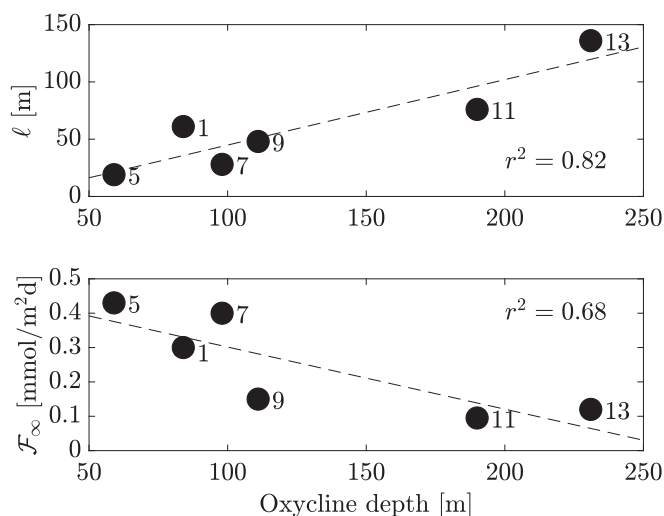
Fig. 5. Correlations of regeneration characteristics with the depth of the top of the ODZ. Both POC regeneration length scale L (*Upper*) and POC preservation to depth F_∞ (*Lower*), derived from exponential fits to POC fluxes at each station (see *Materials and Methods*), are correlated with the depth of the upper boundary of the ODZ [defined as the depth of the appearance of curvature in uncorrected upcast SBE42 oxygen sensor data (19)], indicating that POC regeneration characteristics are strongly controlled by the presence of suboxic waters.

60-mL Savillex jars, a ^{229}Th-^{233}Pa spike and 25 mg of purified iron carrier were added, and the filters sat overnight in concentrated HNO_3 at room temperature. The filters were then completely digested in concentrated perchloric acid ($HClO_4$) to dissolve the polyethersulfone material. Particles were subsequently digested in concentrated HNO_3 and HF, followed by iron coprecipitation. Thorium fractions were isolated using anion exchange chromatography (Bio-Rad AG1-X8, 100 to 200 μm). Measurements of ^{230}Th and ^{232}Th were made on a Thermo Element XR inductively coupled plasma mass spectrometer (ICP-MS) instrument, using an Aridus desolvating nebulizer for sample introduction to improve sensitivity (38). Full details of the LDEO method have been published previously (36, 39).

At UMN, one-eighth–filter aliquots were folded into 30-mL Teflon beakers, a ^{229}Th-^{233}Pa spike was added, and filters were submerged in 7N HNO_3 and 10 drops of concentrated HF. The beakers were capped and heated under pressure for 10 h at 200 °F to leach/digest the samples. After heating, the leach solution was quantitatively transferred to a separate 30-mL Teflon beaker, and five drops of concentrated $HClO_4$ were added. The leach solution was dried down and taken up in 2N HCl, followed by iron hydroxide coprecipitation. The precipitate was dissolved, dried down, and taken up again in 7N HNO_3, which was then loaded onto Bio-Rad AG1-X8 100- to 200-μm mesh resin for separation of Th fractions via anion exchange chromatography. Thorium isotope measurements were made on a Thermo Neptune multicollector ICP-MS instrument (40, 41). In both laboratories, measured ^{230}Th and ^{232}Th were blank corrected using average dipped blank values. Errors in measured ^{230}Th include uncertainties from ICP-MS counting statistics, spike concentrations, and blank corrections. Particulate ^{230}Th and ^{232}Th data are archived at BCO-DMO (https://www.bco-dmo.org/dataset/676231) and in the GEOTRACES Intermediate Data Product (37). More details on the measurements techniques in this study can be found in *SI Appendix, Supplementary Information Text.*

Application of ^{230}Th normalization to POC fluxes. ^{230}Th normalization is a widely used method in paleoceanography for correcting sediment mass accumulation rates for syndepositional redistribution (42, 43). Most ^{230}Th in seawater is produced in the water column by the decay of ^{234}U. Uranium is highly soluble in seawater, stabilized as carbonate complexes (44, 45) with a residence time of hundreds of thousands of years (46). As such, uranium is conservative in seawater, with only minor (parts per thousand) spatial variations in concentration as a function of salinity, allowing for the prediction of oceanic uranium concentrations from salinity (47, 48). These uranium–salinity relationships are used to predict the activity of the major uranium isotope, ^{238}U, which is multiplied by the seawater ^{234}U/^{238}U activity ratio of

1.1468 (49) to estimate ^{234}U. Thus, the production rate of ^{230}Th integrated to a depth horizon z can be predicted anywhere in the water column:

$$P(^{230}Th)_z = \int_0^z \lambda_{230}{}^{234}U \, dz.$$

Unlike its parent ^{234}U, ^{230}Th is highly insoluble in seawater. Upon production by ^{234}U decay, ^{230}Th rapidly adsorbs to particles, with a scavenging residence time of 20 to 40 y (50), much shorter than both its half-life [75,584 y (51)] and the time scale of whole-ocean mixing. The removal of ^{230}Th from a given location is potentially driven by two processes: scavenging removal by particles, and lateral redistribution by advective-diffusive fluxes. Where the latter can be either ignored or corrected, the concentrations of both dissolved and particulate ^{230}Th are expected to increase linearly with depth in a process known as reversible scavenging (52). In this formulation, the integrated production of ^{230}Th to a depth z is balanced in one dimension by its downward export on particles sinking through that depth.

The equation for calculating ^{230}Th$_p$-normalized POC fluxes is nearly identical to that used in paleoceanography to determine vertical constituent fluxes:

$$POCFlux = \frac{P(^{230}Th)_z * [POC]}{[^{230}Th]_p},$$

where the integrated production rate is in μBq·m^{-2}·d^{-1}, $[^{230}Th]_p$ is the activity of particulate ^{230}Th in μBq·m^{-3}, and [POC] is the concentration of POC in mmol·m^{-3}. The resulting POC fluxes we report (Dataset S1) are in units of mmol·m^{-2}·d^{-1}.

We calculate POC fluxes on particles in the 0.8- to 51-μm SSF. Due to low ^{230}Th activity on particles >51 μm, larger filter aliquots were required for analysis than could be routinely measured across the entire section. The actual size of sinking particles carrying ^{230}Th downward to balance its water column production is unknown. However, scavenging removal of ^{230}Th is a two-step process involving adsorption of ^{230}Th onto small particles, which subsequently undergo repeated cycles of aggregation into larger "sinking" particles and disaggregation into smaller "suspended" particles (53, 54). Thus, provided that the aggregation-sinking process is in equilibrium on the time scales of ^{230}Th removal, the POC fluxes recorded by ^{230}Th$_p$ normalization on 0.8- to 51-μm particles will be valid.

Statistical procedures. Power laws of form $F_z = F_{z_0} \left(\frac{z}{z_0}\right)^{-b}$ were fit to POC flux profiles at each station (Fig. 3) using data only at or above 1,000 m. Because the depths of the mixed layer, the deep chlorophyll maximum, and the oxycline varied between stations (*SI Appendix*, Fig. S2), we used the depth of maximum POC flux at each station as the reference depth z_0 rather than interpolating onto a common reference depth (e.g., the base of the euphotic zone or 100 m) across all stations. We show in *SI Appendix, Supplementary Information Text* that our findings are not sensitive to the choice of reference depth.

For Fig. 4 *A* and *B*, data from the suboxic stations at the depths where oxygen concentrations were near zero (60 to 600 m) were grouped, as were data from the oxic stations over the same depth range. Three correlation tests were performed—Pearson's correlation, Spearman's rank correlation, and Kendall's rank correlation (also known as Kendall's tau)—and associated P values were computed for both groups of data (55). By any usual significance threshold, fluxes from suboxic stations of the suboxic stations are not significantly correlated with depth, whereas fluxes from the same depths of the oxic stations significantly decrease with depth.

To quantify the differences between the flux–depth relationships in the oxic versus suboxic stations, we estimated uncertainty in the parameters of the exponential fits using a bootstrap analysis (56). For each group of stations, we generated 10,000 replicate datasets via resampling with replacement and then fit the functional form $F_z = F_{z_0} \exp\left(\frac{-z}{L}\right) + F_\infty$ via nonlinear least-squares regression to each replicate. These 10,000 estimates for each parameter are shown in Fig. 4 *C* and *D* for F_∞ and L, respectively. Based on the intersection of these parameters' estimated probability distributions, we can state with 98% and 90% confidence, respectively, that F_∞ is larger and that L is smaller for the suboxic data than for the oxic data. The median L for the suboxic data is 56 m and the median L for the oxic data is 102 m. The median F_∞ for the suboxic data is 0.36 mmol·m^{-2}·d^{-1} and the median F_∞ for the oxic data is 0.093 mmol·m^{-2}·d^{-1}.

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Supplementary Information for

Shallow Particulate Organic Carbon Regeneration in the South Pacific Ocean

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This PDF file includes:

> Supplementary text
> Figs. S1 to S7
> References for SI reference citations

Supplementary Information Text

Determining Unsupported ^{230}Th$_{xs}$ for computing POC fluxes. Calculating ^{230}Th-normalized fluxes requires that the ^{230}Th is only sourced from the decay of ^{234}U in the water column. However, there is lithogenic particulate ^{230}Th from two separate sources: ^{230}Th supported by ^{234}U in the mineral lattice of detrital particles brought to the ocean by dust or resuspended sediments, and ^{230}Th released during sediment dissolution that subsequently adsorbs onto particles (1). This correction is performed by subtracting the product of measured particulate ^{232}Th and the lithogenic ^{230}Th/^{232}Th ratio (4x10^{-6} mole/mole) (2) from measured ^{230}Th$_p$. All ^{230}Th$_p$ data reported in this paper and used in flux calculations are corrected in this manner. Additional uncertainties from the measured ^{232}Th$_p$ and a nominal 30% uncertainty in the lithogenic ^{230}Th/^{232}Th ratio are propagated into the reported error in ^{230}Th$_p$.

Total ^{230}Th is computed as the sum of small particulate and dissolved ^{230}Th. Where possible, dissolved samples were taken at identical depths as particulate samples, but otherwise were interpolated onto the depths of ^{230}Th$_p$ for calculating total ^{230}Th. Full details of the methods for collecting and measuring dissolved ^{230}Th on the GP16 section have been previously reported (3). Samples were taken from a conventional stainless-steel rosette in Niskin bottles, and 4-5 liters of water were filtered through a 0.45μm Acropak capsule filter. The samples were acidified to pH=2 at sea using 6M ultrapure HCl to prevent adsorption of Th to cubitainer walls during transport to shore (4). In the lab, samples were weighed and spiked with ^{229}Th-^{233}Pa, co-precipitated with iron oxyhydroxide, and digested in HNO$_3$, HF, and HClO$_4$. Separation of Th fractions by anion exchange chromatography and measurement by ICP-MS followed the same

methodology described for particulate samples. Measured dissolved ^{230}Th was corrected for ingrowth by ^{234}U during sample storage.

Testing Possible Biases in ^{230}Th-derived POC fluxes. We test the validity of using the 0.8-51μm particle size class by comparing ^{230}Th$_p$-normalized 0.8-51μm POC fluxes with results from annually-averaged bottom-moored sediment traps. Hayes et al. (16) found that ^{230}Th$_p$-derived POC fluxes below 1000m were within 8% and 21% of multi-year averages from collocated deep moored sediment traps at the Bermuda Atlantic Time-Series. On the GP16 section, ^{230}Th$_p$-derived POC fluxes agree closely with annual POC fluxes from nearby deep sediment traps (7, 8). At 135ºW, sediment trap POC fluxes are within 1-sigma uncertainty of ^{230}Th$_p$-derived fluxes at similar depths (*SI Appendix* Fig. S1a), while at 100ºW, ^{230}Th$_p$-normalized fluxes at 14ºS and 16ºS are halfway between the sediment trap values at 10ºS and 20ºS, capturing the regional latitudinal productivity gradient (*SI Appendix,* Fig. S1b). The close agreement between the ^{230}Th$_p$-normalized fluxes of 0.8-51 μm POC and sediment trap estimates indicates that aggregation-disaggregation between small and large particles is of primary importance for the downward transfer of POC in the mesopelagic and bathypelagic, in agreement with optical particle observations (9) and biomarker ^{13}C results (10).

 We test whether size fractionation is a potential bias by comparing >51μm and 0.8-51μm POC/^{230}Th$_p$ ratios for the limited number of samples for which it was possible to analyze both size fractions (*SI Appendix* Fig. S6). POC/^{230}Th$_p$ is 29±6% higher in the >51μm size fraction (n=23, R^2=0.95), but there is no appreciable trend in size fractionation with depth. Not including a single high POC/^{230}Th$_p$ value in both size

fractions from station 7 that drives the most of the correlation, the best-fit POC/^{230}Th$_p$ slope is 1.05 ± 0.11 (n=22, R^2=0.8), within uncertainty of unity (*SI Appendix,* Fig. S6). So, while it is possible that our fluxes derived from 0.8-51μm particles slightly underestimate the overall POC flux, this should not affect our interpretation of the shape of the profiles (i.e., the gradient with depth in the POC flux).

The POC fluxes we observe are also not influenced by lateral transport of organic matter from the continental shelf. Sanial et al. (11) measured ^{228}Ra, which is supplied by diffusion from sediments and thus traces the advection of material from the continental shelf, on the GP16 section. While ^{228}Ra is enriched in the surface stations from the shelf to the open ocean, the enrichment is mostly confined to oxic waters above the OMZ, indicating low lateral inputs of shelf-derived material to ODZ depths (11). Additionally, the offshore gradient in POC flux within the ODZ is minimal (Fig. 1, *SI Appendix* Fig. S3), whereas lateral transport of POC from the shelf would likely cause there to be much higher POC fluxes near the shelf that drop off rapidly towards the open ocean.

Application of ^{230}Th$_p$-normalization requires there to be no net advective and/or diffusive gain or loss of ^{230}Th in the water column. Lateral transport of ^{230}Th could potentially bias ^{230}Th-normalized fluxes. High particle fluxes in productive regions could result in lower dissolved ^{230}Th concentrations than in oligotrophic regions with low particles fluxes. Lateral eddy diffusion could then transport ^{230}Th from the interior towards the margin, a process known as boundary scavenging (12, 13). The short residence time of ^{230}Th generally limits its redistribution by lateral isopycnal diffusion (14), and annually-averaged sediment trap fluxes of ^{230}Th are typically within 10-15% of the production rate in the overlying water column (15). A slight zonal gradient in

dissolved and total ^{230}Th is apparent in our profiles below 400m (*SI Appendix,* Fig. S7). Net shoreward transport of ^{230}Th by mixing would act to increase the apparent ^{230}Th$_p$-normalized POC fluxes at sites further offshore, and decrease them close to shore. However, there is no discernable feature in the POC flux profiles below 400m that would indicate shoreward transport of ^{230}Th at these depths, and the bulk of the differences between POC flux profiles is above the depths where there is an offshore ^{230}Th gradient. Additionally, both surface (16) and 400m subsurface (17) net zonal flows at 8-12ºS are negligible. Thus, lateral transport by either advection or eddy diffusion does not impact our results.

Upwelling rates of 0-3 m d^{-1} have been estimated in nearshore waters off Peru using ^{3}He and ^{7}Be methods (18-20). Since total ^{230}Th increases with depth, upwelling can transport ^{230}Th upward through the base of the mixed layer. In this case, the sinking flux of particulate ^{230}Th will be balanced by its production from ^{234}U decay and its supply from upwelling. Adding an upwelling flux would increase the ^{230}Th$_p$-derived POC flux estimates in the uppermost water column, making our maximum flux estimates a lower bound. However, we focus our interpretations on the depth-dependence of POC regeneration rates, which is not influenced by upwelling. If the shape of the regeneration profile were dictated by the upwelling of ^{230}Th, we would expect this to be a consistent feature across stations influenced by upwelling, whereas the regeneration features instead track the depth of the oxycline (Fig. 5). Further, performing statistical analysis of the POC regeneration length scales only by station groupings averages out any potential effects of upwelling at a single station.

Testing Possible Biases in Choice of Reference Depth. The choice of reference depth can affect the b values for power law fits of POC flux profiles (21). We compare b and reference depth z_0 from the GP16 section (*SI Appendix,* Fig. S8), finding a weak positive correlation between b and z_0 ($R^2=0.27$, p=0.034, n=17). This correlation is in the opposite direction of what we would expect to find if shallow reference depths were driving artificially high b values. We also show that the location of reference depth relative to the MLD and DCM does not bias our b estimates. At some stations, the reference depth z_0 falls slightly above the DCM (15,21,23,25,28) or above both the DCM and MLD (5,11,13,17,26). In these cases, the POC/^{230}Th ratio could potentially be offset from the POC/^{230}Th ratio of sinking particles due to the contribution of POC from newly produced, non-sinking particles that are recycled *in-situ*. We test whether the inclusion of these stations biases our fitting procedures. The mean of b values from stations where z_0 is above the DCM and/or MLD (5,11,13,17,21,23,25,26,28) is 1.18±0.34, within uncertainty and very close to the mean b value of 1.09±0.26 from stations where z_0 is below the MLD and DCM (1,7,9,30,32,36). Limiting the comparison only to offshore stations (11-36), the stations with z_0 above the DCM only (15,21,23,25,28) have a mean b value of 1.3±0.16, within error of and similar to the mean b values of 1.24±0.12 from stations with z_0 above both the MLD and DCM (11,13,17,26), and of 1.23±0.11 from stations where z_0 is deeper than both the DCM and MLD (30,32,36). We also re-fit the exponential functions at stations 11 and 13 using z_0 as the first sampling depth below the MLD and DCM to test whether the correlation in Fig. 5 is an artifact of the choice of reference depth. In this sensitivity test, the correlation between oxycline depth and L still holds, with $R^2=0.72$, as does the correlation between oxycline depth and F_∞, with

R^2=0.69. We thus conclude that neither our b values, nor the correlation between oxycline depth and regeneration characteristics from exponential fitting, are artifacts of the choice of reference depth.
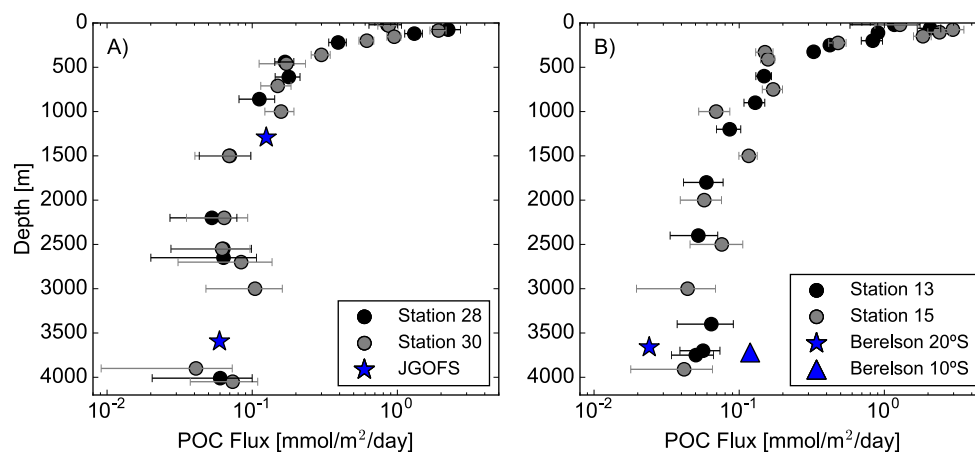
Fig. S1. Comparison between GP16 ^{230}Th$_p$-normalized POC fluxes and sediment trap estimates. ^{230}Th$_p$-normalized fluxes from stations zonally-bracketing the sediment trap longitudes are shown as black and gray dots. Annually-averaged sediment trap fluxes are shown as blue stars or blue triangles. a, Comparison between GP16 stations 28 and 30 with sediment trap fluxes from the JGOFS program (12ºS, 135ºW), deployed in 1992 (7). b, Comparison between GP16 stations 13 and 15 at 14ºS and 16ºS with sediment trap deployments from 2010-2011 at 10ºS and 20ºS along 100ºW (8).
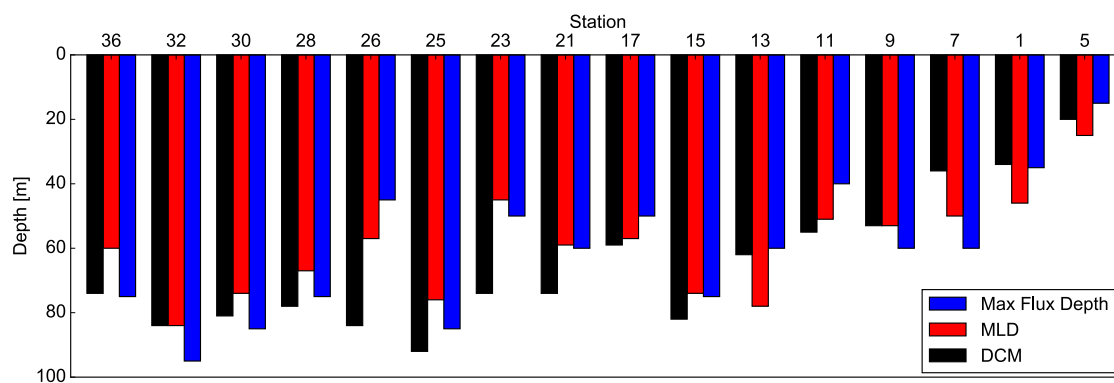
Fig. S2. Depth of Maximum POC Flux (this study), Mixed Layer Depth (MLD) **(22)** and Deep Chlorophyll Maximum (DCM) **(22)**. Station 18 is omitted because there were no $^{230}Th_p$ measurements on samples above 150m. Stations are oriented from east (right) to west (left).
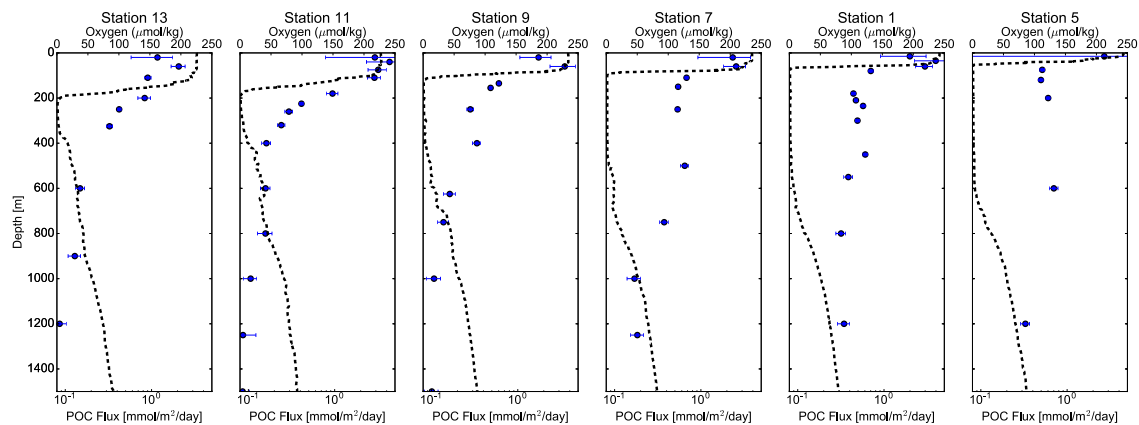
Fig. S3. POC Fluxes (blue symbols) and Oxygen Profiles (dashed lines) at GP16 Stations 1-13. Station order is shown from east to west going right to left. The length scale of POC flux attenuation tracks the depth of the upper oxycline at these stations.
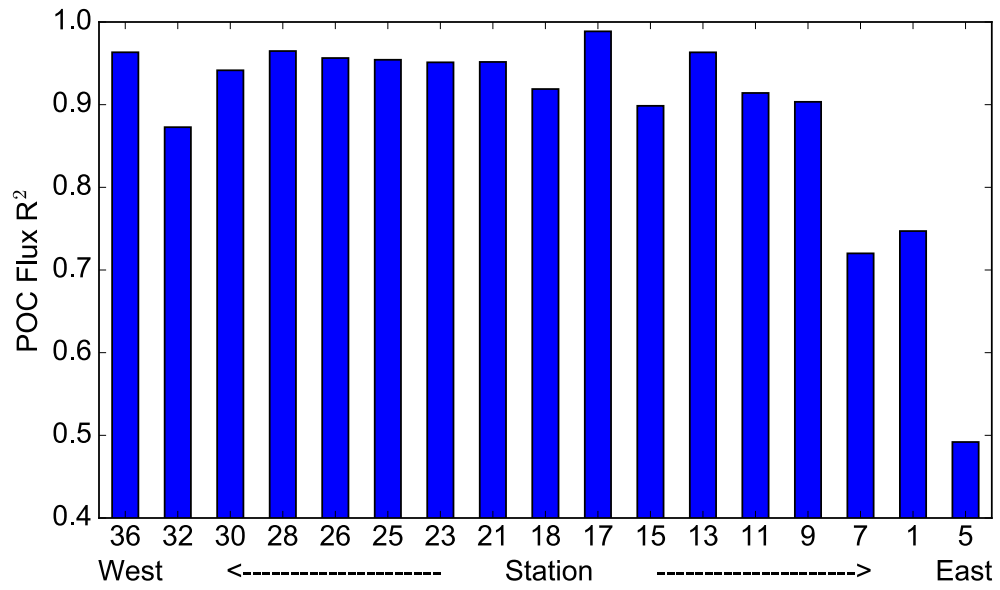
Fig. S4. R^2 values for power law fits to POC flux data. Fits are poorer at stations 5, 1, and 7, where the upper oxycline is shallowest, indicating the power laws do not properly represent the functional form of POC flux profiles at ODZ stations.
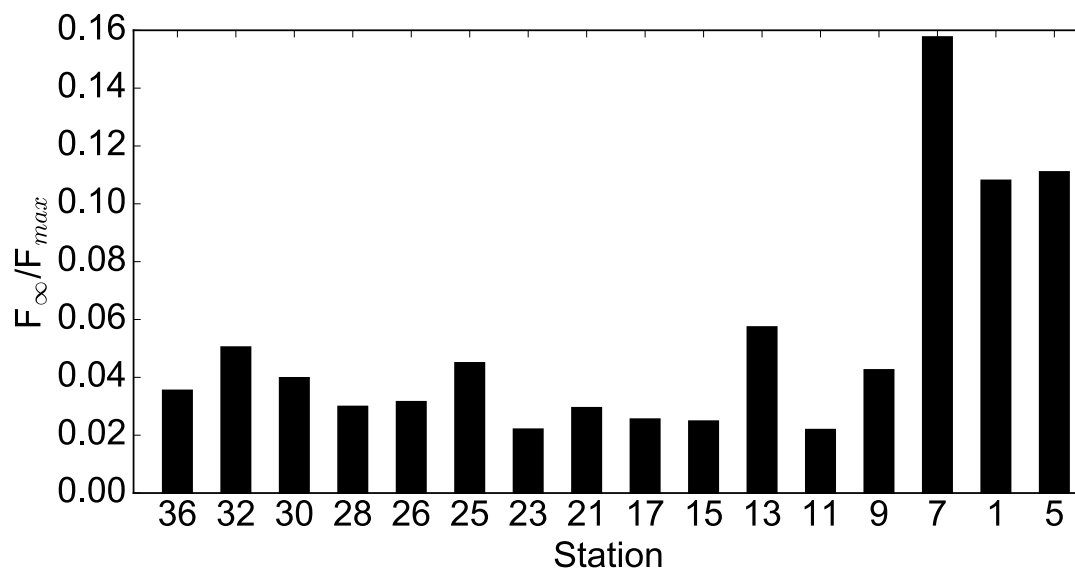
Fig. S5. Transfer efficiencies for POC fluxes. Values are computed as the ratio between the best-fit F_∞ from exponential fits to POC flux profiles and the observed maximum POC flux at each station. Highest values at suboxic stations 5, 1, and 7 indicate greater POC preservation at the stations with thickest ODZs.
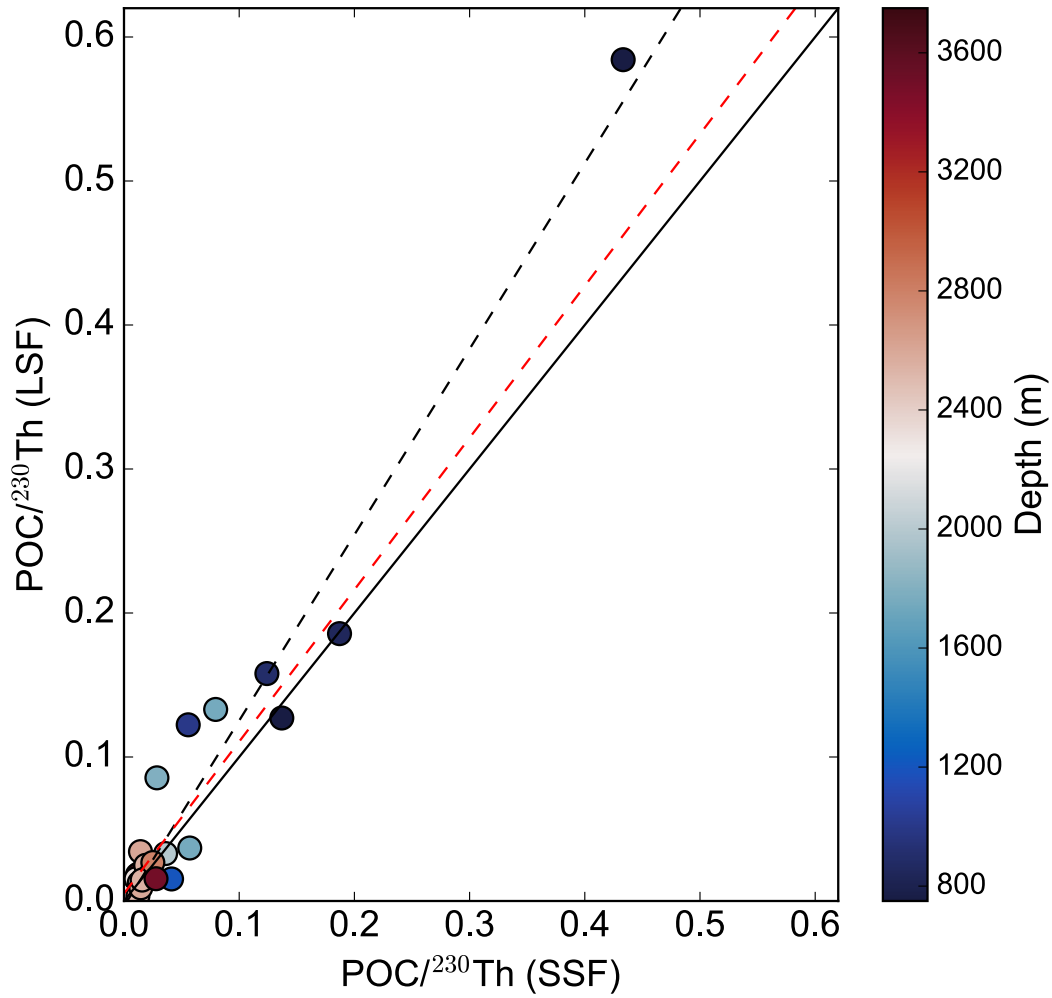
Fig. S6. Size Partitioning of POC/^{230}Th. Dots show individual data points of POC/^{230}Th ratios in large size fraction particles (LSF, >51μm) and small size fraction particles (SSF, 0.8-51μm), colored by depth. Dashed line black shows the best-fit line (slope=1.29, n=23, R^2=0.95) for all data points. Since much of the correlation is driven by a single high POC/^{230}Th ratio point from 750m at Station 7, we also fit the data without this point – the best fit line excluding this point is shown as a dashed red line (slope=1.05, n=22, R^2=0.81). The 1:1 line is shown as a solid black line.
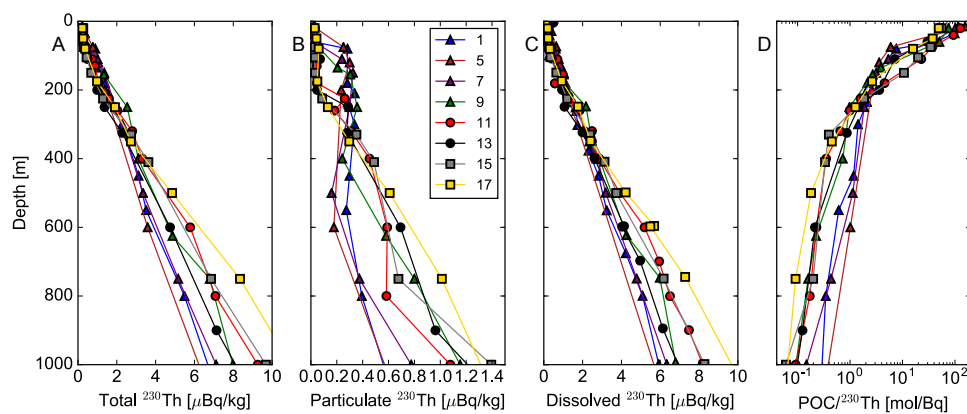
Fig. S7. ^{230}Th Profiles and POC/^{230}Th ratios. a, Total ^{230}Th. b, Particulate ^{230}Th. c, Dissolved ^{230}Th. d, POC/^{230}Th$_p$ ratios. Data from ODZ stations are shown as triangles.
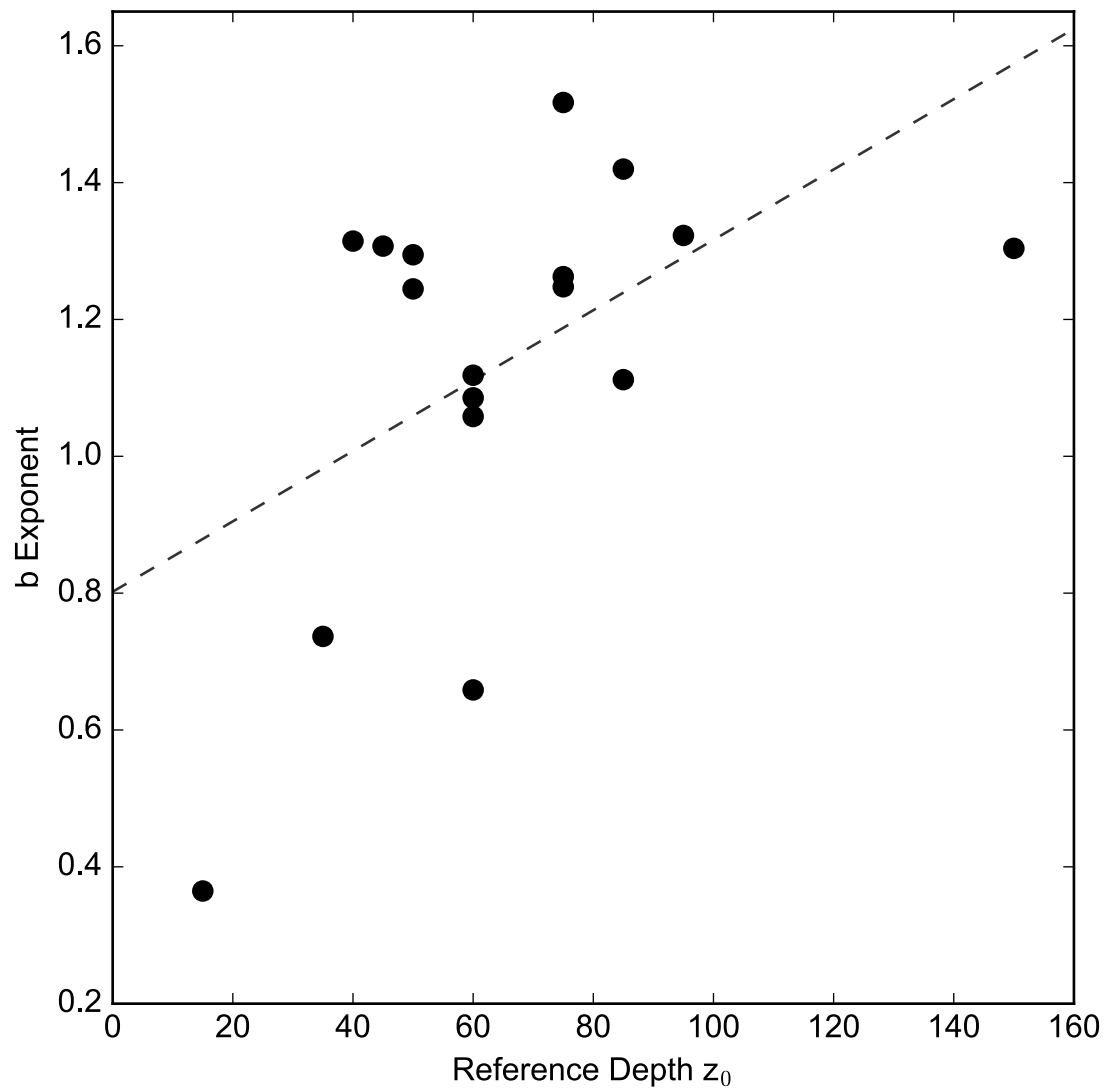
Fig. S8. Relationship of b values and reference depth z_0. Dots show points from individual stations. Dashed line shows the least-squares linear fit to the data, which are weakly correlated with slope 0.005 ± 0.002 ($R^2=0.27$, p=0.034, n=17).

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