**Dissolved and Particulate ^{230}Th - ^{232}Th in the Central Equatorial Pacific Ocean:
Evidence for Far-Field Transport of the East Pacific Rise Hydrothermal Plume**

Grecia I. Lopez, Franco Marcantonio*

Department of Geology and Geophysics, Texas A&M University, College Station, TX 77843

Mitch Lyle**

Department of Oceanography, Texas A&M University, College Station, TX 77843

Jean Lynch-Stieglitz

School of Earth and Atmospheric Sciences, Georgia Institute of Technology. Atlanta, GA 30332

*corresponding author, email address: marcantonio@tamu.edu, phone: 979-845-9240, fax: 979-845-6162

**now at: College of Earth, Ocean, and Atmos. Sci., Oregon State University, Corvallis, OR 97331

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ABSTRACT

We assess the distribution of ^{230}Th and ^{232}Th along a latitudinal gradient in the Central Equatorial Pacific Ocean ($\sim 155^\circ \text{W}$ - 159°W) at two sites: 8°N and the equator. The dissolved ^{230}Th concentration profile at 8°N increases nearly linearly from the surface to 2000 m, exhibiting behavior consistent with thermodynamic reversible scavenging. However, from 2000 m to 3000 m, the dissolved ^{230}Th concentrations exhibit little change, before increasing slightly from 3000 m to the bottom. At this site dissolved ^{230}Th concentrations range from 1.1 fg/kg at 100 m to 55.2 fg/kg at 4600 m. At the equator, dissolved ^{230}Th concentrations are slightly lower, and range from undetectable at 25 m to 19.1 fg/kg at 3038 m. The pattern in the dissolved ^{230}Th concentration profile at the equator is indistinguishable from that at 8°N . The mid-depth-water deviation from equilibrium reversible scavenging between 2 and 3 km in the ^{230}Th profiles (lower concentrations than expected) at both sites occurs in the interval of the water column that is consistent with an interval that has high concentrations of ^3He and dissolved Fe at other nearby sites. This ^3He - and Fe-rich signal has been traced to hydrothermal plumes from the East Pacific Rise, thousands of km away. We hypothesize that the lower concentrations of ^{230}Th in mid-depth waters of the Central Equatorial Pacific are a result of a 5000-km transit of waters that have had their ^{230}Th scavenged by Fe-Mn particulates close to the EPR.

Oceanic residence times of thorium combined with dissolved ^{232}Th concentrations suggest dust fluxes of about $\sim 0.5 - 0.6 \text{ g m}^{-2} \text{ yr}^{-1}$ to the sea surface.

These fluxes are in agreement with other empirical studies in the Pacific, but are higher than those suggested by global atmospheric circulation models.

1. INTRODUCTION

^{230}Th ($t_{1/2} = 75.69$ kyr) is widely used as a quantitative marine geochemical tracer due to its highly reactive chemical behavior. It is produced by the radioactive decay of ^{234}U , which is conservative in seawater and has a long residence time (~200 kyr; Henderson and Anderson, 2003). ^{230}Th is essentially insoluble in seawater, so as it is produced it is quickly removed from the water column by adsorption onto particles which then settle to the seafloor. This process of adsorption and removal of ^{230}Th from the water column is called *scavenging*. ^{230}Th concentrations in seawater provide insight into oceanographic processes such as deep oceanic circulation and mixing rates, and marine particle dynamics (e.g., Nozaki et al., 1981; Van der Loeff and Berger, 1993; Moran et al., 1997). In addition, ^{230}Th studies of marine sediments provide important information on past changes in fluxes to the seafloor, which can be used to better understand climate-related changes in marine biological productivity, dust input to the oceans, and deep water circulation once the systematics are adequately understood.

Due to the quick scavenging of ^{230}Th by particles, the residence time of ^{230}Th in the ocean is approximately 10 - 40 years (Henderson and Anderson, 2003). It is assumed that the excess flux of Th carried to the seafloor by particles is equivalent to its rate of production (P_{Th}) by the decay of ^{234}U throughout the water column (Bacon, 1984; Francois et al., 2004; Henderson et al., 1999; Krishnaswami, 1976). Model simulations (Henderson et al., 1999) and sediment trap studies (Scholten et al., 2005; Yu et al., 2001) have been used to test this assumption, yielding the hypothesis that the flux of scavenged ^{230}Th (F_{Th}) to sediment at the seafloor remains within 30% of P_{Th} in 70% of the earth's

23 oceans. A linear increase in dissolved ^{230}Th concentration with depth in the water
24 column is expected due to thermodynamic reversible exchange between dissolved and
25 particulate ^{230}Th (Bacon and Anderson, 1982; Nozaki et al., 1981; Scholten et al., 2005;
26 Van der Loeff and Berger, 1993; Moran et al., 1997). However, linear profiles are not as
27 common as expected, and have been found only in some regions of the Southern Ocean
28 (Van der Loeff and Berger, 1993), and the Central Pacific (Roy-Barman et al., 1996) and
29 Northwestern Pacific (Nozaki et al., 1981) oceans. It is more common to find deviations
30 from linearity. Such deviations indicate disturbance of the vertical equilibrium by lateral
31 fluxes, or a change in the scavenging efficiency due to changes in particulate
32 composition or flux (Hayes et al., 2014), and/or a failure of the assumption of
33 equilibrium reversible scavenging.

34 Unlike the majority of ^{230}Th , which is supplied to the water column by the in-situ
35 radioactive decay of ^{234}U , ^{232}Th is derived from the leaching in open ocean waters of
36 wind-blown dust from the continent. Wind-blown mineral aerosols (dust) contribute
37 iron and other micronutrients to the surface ocean and therefore play an important role in
38 the carbon cycle (e.g., Martin et al., 1990). In addition, dust affects the planet's radiative
39 balance because of its role in the albedo (e.g., Jickells et al., 2005; Mahowald et al.,
40 2005). Quantifying the dust flux is challenging (Jickells et al., 2005; Hsieh et al., 2011),
41 and dust input to the surface ocean has been assessed using remote sensing and modeling
42 (Mahowald et al., 2005), combined with measurements of dissolved Al and Th
43 concentrations in the surface ocean (Measures and Brown, 1996; Hsieh et al., 2011;
44 Hayes et al., 2013a). Indeed, when coupled with dissolved ^{230}Th analyses in the same

45 water samples, dissolved ^{232}Th measurements offer a way to quantify the supply of
46 dissolved ^{232}Th delivered by dust to the modern ocean (Deng et al., 2014; Hayes et al.,
47 2013a; Hsieh et al., 2011). Such indirect measurements of the dust flux to the sea
48 surface, therefore, yield potential insights into the flux of dissolved Fe, an important
49 limiting micronutrient supplied to the world's oceans via eolian processes (Jickells et al.,
50 2005).

51 Here, we investigate the ^{230}Th and ^{232}Th depth profiles in the Central Pacific
52 Basin. We find a non-linear profile with depth for ^{230}Th . We discuss several
53 possibilities that can explain the non-linearity observed in the ^{230}Th profile, including:
54 renewal of basin waters with different dissolved Th concentrations, recent mixing with
55 depleted or enriched water masses, hydrothermal scavenging, and advection or diffusion
56 of ^{230}Th in the water column (e.g. Hayes et al., 2014; Singh et al., 2013; Venchiarutti et
57 al., 2008). In addition, we combine the systematics of dissolved $^{232}\text{Th} - ^{230}\text{Th}$ seawater
58 concentrations to determine lithogenic input to the upper ocean in the Central Equatorial
59 Pacific (CEP), and compare our results to others for the Pacific Ocean.

60

61 **2. METHODOLOGY**

62 **2.1 Seawater Sample Collection**

63 Seawater cast samples were taken from two sites in the Central Equatorial Pacific
64 during an NSF-funded cruise to core the Line Islands Ridge in May 2012 (ML1208, on
65 R/V Marcus Langseth). Five-liter samples at varying depths (24.5 – 4601 meters) were
66 collected at 8°N, and at the equator using 24 PVC Niskin bottles mounted on a

67 conductivity–temperature–depth (CTD) rosette system. We have assessed the
68 distribution of total and dissolved ^{230}Th and ^{232}Th along a latitudinal gradient ($\sim 155^\circ \text{W}$ -
69 159°W ; Figure 1), from the equator to 8°N , an area that subsumes both low and high
70 particle flux in the Central Equatorial Pacific. Sample ML1208-03CTD ($00^\circ 13.166' \text{S}$,
71 $155^\circ 57.668' \text{W}$) was collected at the equator whereas sample ML1208-12CTD was
72 collected at the southern edge of the North Pacific gyre ($8^\circ 19.989' \text{N}$, $159^\circ 18.000' \text{W}$).

73 Collected seawater samples in PVC Niskin bottles were transferred via acid-
74 washed Teflon tubing to prewashed 5-L polyethylene-collapsible cubitainers promptly
75 after collection. Each cubitainer was rinsed with trace metal grade acid (10% HCl),
76 cleaned three times with Milli-Q water, air-dried in a laminar airflow hood and stored
77 separately inside the clean laboratory of the R. Ken Williams '45 Radiogenic Isotope
78 Geosciences Laboratory at Texas A&M University. $0.45\text{-}\mu\text{m}$ Acropak capsule filters
79 were used to isolate the dissolved (filtered) fraction from the particulate (unfiltered)
80 fraction during the transfer of seawater into the cubitainers. Prior to sampling, each
81 cubitainer was washed with small amounts of seawater before the final transfer of the
82 whole seawater sample. For all depths, 5-L filtered and 5-L unfiltered seawater samples
83 were collected to study the dissolved and particulate ^{230}Th and ^{232}Th . Each sample was
84 acidified with concentrated (9N) ultrapure (Optima-grade) HCl to a pH of approximately
85 1.5 and sent to Texas A&M University for radionuclide chemical analyses (Singh et al.,
86 2013).

87 **2.2 Radionuclide Analyses**

88

89 Using the revised protocol of the GEOTRACES program (Anderson et al., 2012)
90 radionuclide measurements of total and dissolved ^{230}Th and ^{232}Th were evaluated. Water
91 samples were weighed and spiked with ^{229}Th for isotope dilution analysis. To induce
92 $\text{Fe}(\text{OH})_3$ precipitation, a purified Fe-carrier solution (FeCl_3) was added. The iron
93 chloride was purified by back extraction with isopropyl ether. After addition of the ^{229}Th
94 spike and the FeCl_3 to the seawater sample, it was left to equilibrate for a day before the
95 addition of trace metal-grade NH_4OH to bring the pH to 8-8.5. Once at desired pH,
96 removal of all Th and precipitation of iron hydroxide [$\text{Fe}(\text{OH})_3$] within the seawater
97 sample was achieved. The sample was left to equilibrate for 5 days, so the precipitate
98 was able to settle undisturbed at the bottom of the cubitainer. After 5 days, the
99 precipitate was separated from the supernatant solution through siphoning into 1 L acid-
100 washed Nalgene bottles, and left to equilibrate for a full day. The following day, the
101 sample was siphoned into acid-washed 50 mL centrifuge tubes, and placed in a
102 centrifuge at 1700 rpm for 15 min. Samples were centrifuged three times, decanting the
103 supernatant each time before dissolving the precipitate in ultrapure HNO_3 . The resulting
104 solution is then evaporated in an acid-washed 15 mL savillex beaker. Once evaporated
105 the sample was reconstituted in 8N HNO_3 . The sample solution was then processed
106 through anionic ion-exchange columns (AG1X8, 100-200 mesh, chloride form) to purify
107 and extract Th for isotope ratio measurement.

108 During column chemistry, each column was washed and conditioned (8N HNO_3 ,
109 Milli-Q water, and concentrated HCl) before adding the sample. After each sample was
110 added, the column was washed again, and a new acid-washed 15 mL beaker was placed

111 to collect the Th (concentrated HCl is used during this step). Promptly after collecting
112 Th, two drops of HClO_4 and 0.25 mL of concentrated HNO_3 were added and left to
113 evaporate and fume on a hot plate at 180° . Once the solution reached 0.05 mL in
114 volume, 2% HNO_3 /0.1% HF was added and taken up to 1 mL of solution in 1.5 mL acid-
115 washed centrifuge tubes for isotope ratio analysis on the Element XR magnetic sector
116 ICP-MS. During each batch of sample analyses throughout this study (May- November
117 2013) several procedural blanks for both ^{230}Th (8 ± 6 fg/kg) and ^{232}Th (13 ± 8 pg/kg) were
118 run. The average ^{230}Th and ^{232}Th blank corrections are about 12% and 21%, respectively.
119 Error bars for all data points (Tables 1, 2) are largely a function of the uncertainty in
120 these blank corrections.

121

122 **3. RESULTS**

123 **3.1 North Pacific Gyre (8°N)**

124 At 8°N , dissolved ^{230}Th concentrations range from 1.1 fg/kg at 100 m to 30.8
125 fg/kg at 4400 m, while dissolved ^{232}Th concentrations span from 10.2 pg/kg at 100 m to
126 19.7 pg/kg at 4400 m (Table 1, Figures 2 and 3). The pattern of the dissolved ^{230}Th
127 profile at 8°N is essentially linear from the surface to 2000 m. From 2000 m to 3000 m,
128 the dissolved (filtered) ^{230}Th concentrations are constant, and then from 3000 m to the
129 bottom, the profile is roughly linear (Figure 2). At the same site, total ^{230}Th and ^{232}Th
130 concentrations range from 1.1 fg/kg and 5.2 pg/kg, respectively, at 100 m to 55.2 fg/kg,
131 53.9 pg/kg, respectively, at 4600 m. (Table 1). The total (unfiltered) ^{230}Th profile mimics
132 that of the dissolved ^{230}Th concentrations (Figure 2). The particulate fraction (filtered –

unfiltered) of the total seawater ^{230}Th increases from about 0% at the surface to 38% at 4400 m (Table 1). From 0 to 3000 m at 8°N , dissolved ^{232}Th concentrations show little variability ($\sim 20\%$; Figure 3 and Table 1). From 3000 m to 4600 m, dissolved ^{232}Th contents are more variable and show a slight increase toward greater depths. The proportion of ^{232}Th in the particulate fraction of the total seawater sample increases exponentially with depth to a value of 56% in the bottommost sample (Figure 3), evidence for a deep nepheloid layer, perhaps sourced at the Line Islands Ridge to the west. Due to errors in experimental procedure, we did not collect data for approximately nine samples (indicated by *, Table 1).

142

143 **3.2 Equator**

Thorium concentrations from the equator exhibit the same general trend that is exhibited at 8°N (North Pacific gyre). However, ^{230}Th values are not as high as those found at 8°N (Figure 2). Dissolved ^{230}Th concentrations range from not detectable at 25 m to 19.1 fg/kg at 3038 m of depth. The pattern of the dissolved ^{230}Th profile at 0° is essentially linear from the surface to 2000 m. Between 2000 m and 3000 m depth, dissolved ^{230}Th concentrations show little variability; beyond 3000 m depth, dissolved ^{230}Th concentrations increase with depth. Total ^{230}Th concentrations span from 2.2 fg/kg at 25 m to 25.4 fg/kg at 3038 m, while total ^{232}Th concentrations vary from 19.3 pg/kg at 25 m to 325.0 pg/kg at 3038 m (bottommost depth) (Table 2). As at 8°N , the equator profile of the total ^{230}Th concentrations mimics that of the dissolved ^{230}Th concentrations (Figure 2). The particulate fraction of the total seawater ^{230}Th ranges from about

155 virtually all of the ^{230}Th that the surface to 24% at 3038 m (Table 2). There is a factor-of-
156 two variability within the dissolved ^{232}Th concentrations, spanning from 3.9 pg/kg at 25
157 m to 8.1 pg/kg at 3038 m. Overall, however, there is no relationship of ^{232}Th
158 concentrations with depth (Figure 3 and Table 2). The bottommost sample has an
159 anomalously high total ^{232}Th concentration (Figure 3) and may reflect active
160 resuspension of ^{232}Th -enriched fine-grained particles from the seafloor, or, it may be a
161 result of an anomalous blank problem, which has been documented for ^{232}Th analyses in
162 the GEOTRACES literature (Anderson et al., 2012). The latter problem may also have
163 occurred at the 8°N site in the analysis of total ^{232}Th at 3600 m.

164

165 **4. DISCUSSION**

166 **4.1 ^{230}Th Residence Time and Reversible Scavenging**

167 To provide further evidence for the behavior observed in our ^{230}Th profiles, we
168 compare our ^{230}Th residence time values in the Central Equatorial Pacific to residence
169 times calculated in other areas of the Pacific Ocean. We calculate total water column
170 residence times for ^{230}Th at both of our sampling stations (00° 13.166' S, 155° 57.668'
171 W; 8° 19.989' N, 159° 18.000' W). These calculations can be used to estimate the extent
172 of depletion found within our ^{230}Th profiles relative to reversible scavenging. The first
173 station (03CTD) is located at the equator where particle fluxes are greatest due to
174 upwelling and high export particulate production. The second station (12CTD) is at 8°N
175 and is within the oligotrophic North Pacific gyre where particulate fluxes are lower than

176 those at the equator. Assuming no lateral transport, scavenging residence times of ^{230}Th
 177 were calculated using:

$$\frac{\int_{z_0}^{z_f} {}^{230}\text{Th}(z) dz}{P_d z_f} \quad (1)$$

178 where the numerator represents the integrated inventory of total water column dissolved
 179 ^{230}Th , and the terms in the denominator, P_d and z_f , represent the production of ^{230}Th from
 180 ^{234}U decay in the water column ($0.0267 \text{ dpm m}^{-3} \text{ y}^{-1}$), and the depth of the water column
 181 for each station, respectively. Traditionally, units of activity (dpm or disintegration per
 182 minute) have been used to measure ^{230}Th . We use such activity and weight units
 183 interchangeably here so that our results can be compared with earlier work in the
 184 literature. To convert activities to weight concentrations, one must divide by the ^{230}Th
 185 decay constant and Avogadro's number, and then multiply by the atomic weight of
 186 ^{230}Th .

187 The scavenging residence times of ^{230}Th for our equator and North Pacific sites
 188 are 17 and 31 years, respectively, within the range of oceanic ^{230}Th residence times
 189 recorded in the Pacific Ocean (10 - 41 years, as summarized in Okubo et al., 2012). The
 190 ^{230}Th residence time at the equator is lower by about a factor of 2 compared to the North
 191 Pacific site. Most of this difference is due to the greater water depth and, hence, larger
 192 inventory of ^{230}Th at the 8°N site.

193 Throughout many of the world's ocean basins, ^{230}Th water column profiles are
 194 linear with depth suggesting reversible scavenging between dissolved and particulate

195 ^{230}Th (Bacon and Anderson, 1982; Francois et al., 2004; Krishnaswami et al., 1976;
 196 Nozaki et al., 1981; Roy-Barman et al., 1996). Although linear increases of ^{230}Th
 197 concentrations with depth are observed at both of our sites from about 0-2000 m, this
 198 relationship breaks down at depths greater than 2000 m, as was also observed in the
 199 eastern Pacific by Singh et al. (2013). The deviation from reversible scavenging below
 200 2000 m is significant. To quantify the extent of deviation from reversible scavenging we
 201 construct a vertical profile using a one-dimensional model in which equilibrium exists
 202 between adsorption/desorption reactions involving ^{230}Th in the dissolved phase and
 203 vertically sinking particles. In this simple model we assume that advection of ^{230}Th does
 204 not occur. The activity of model dissolved ^{230}Th , C_d , is given by (e.g., Krishnaswami et
 205 al., 1976; Anderson and Bacon, 1982; Van der Loeff and Berger, 1993; Moran et al.,
 206 1997; Okubo et al., 2012; Hayes et al., 2013b):

$$C_d = \left(\frac{P_d}{SK} z \right) \quad (2)$$

207 where P_d represents the production of ^{230}Th from ^{234}U decay in the water column; S is the
 208 ^{230}Th particle settling rate (m y^{-1}); K is defined as the distribution coefficient of ^{230}Th
 209 between the particulate (unfiltered-filtered) and dissolved (filtered) phases; and depth, z ,
 210 is measured in m. Particulate and dissolved data were used to determine the distribution
 211 coefficient K , which was calculated using particulate and dissolved ^{230}Th values. To
 212 determine particulate ^{230}Th , dissolved ^{230}Th was subtracted from the total ^{230}Th . For the
 213 North Pacific gyre an average K value of 0.12 ± 0.06 was calculated; whereas for the
 214 equator a slightly higher average K value of 0.23 ± 0.12 was determined. The K values at

the sites are within error of each other and fall within the range of K values found in the Pacific Ocean (0.1 – 0.3; Bacon and Anderson, 1982; Nozaki et al., 1981; Okubo et al., 2012). For equation (2), S can be approximated using measured dissolved ^{230}Th , depth (z), the constant, P_d , and the average K (Krishnaswami et al., 1976). The average settling rate values deduced in this way for the equator ($524 \pm 74 \text{ m y}^{-1}$) and North Pacific gyre ($376 \pm 68 \text{ m y}^{-1}$) are similar to others estimated in a similar fashion for the Pacific Ocean ($380 \text{ m y}^{-1} - 580 \text{ m y}^{-1}$; Bacon and Anderson, 1982; Nozaki et al., 1981; Okubo et al., 2012).

The distribution profiles of dissolved ^{230}Th derived from the reversible scavenging model can be seen in Figure 4 and Figure 5 (solid black lines). The extent of anomalous depletion of ^{230}Th from a linear reversible scavenging profile can be calculated as follows:

$$\text{Depletion \%} = \frac{\int_{z_0}^{z_f} (\text{observed dissolved } ^{230}\text{Th}(z) - C_d(z)) dz}{\int_{z_0}^{z_f} C_d(z) dz} \quad (3)$$

where the numerator is the integrated inventory of the difference between observed dissolved ^{230}Th and the dissolved ^{230}Th predicted by the reversible scavenging model throughout the whole water column, and the denominator is the integrated inventory of dissolved ^{230}Th . The depletion in ^{230}Th compared to reversible scavenging occurs below a depth of 2000 m (Figures 4 and 5). The quantitative extent of this depletion, from

232 about a depth of 2000 m to the bottom of the profile is greater (27%) at the North Pacific
233 Gyre site than that (18%) at the equator site.

234 Depletions in ^{230}Th have been attributed to bottom scavenging (Bacon and
235 Anderson, 1982; Nozaki and Nakanishi, 1985; Okubo et al., 2012; Singh et al., 2013;
236 Hayes et al., 2013b), boundary scavenging (Anderson et al., 1983; Nozaki and
237 Nakanishi, 1985; Hayes et al., 2013b), renewal of basin waters with different dissolved
238 Th concentrations (Francois et al., 2004; Hayes et al., 2014), enhanced scavenging of Th
239 in high particle flux zones (Anderson et al., 1983), recent mixing with depleted or
240 enriched water masses (Van der Loeff and Berger, 1993; Moran et al., 1997; Francois et
241 al., 2004; Okubo et al., 2012; Hayes et al., 2013a, 2014), advection or diffusion of ^{230}Th
242 in the water column (Nozaki et al., 1981; Bacon and Anderson, 1982; Van der Loeff and
243 Berger, 1993; Venchiarutti et al., 2008; 2013), and/or hydrothermal scavenging (Kadko,
244 1996; Hayes et al., 2014). In the CEP, we believe that the cause of the depletions in
245 ^{230}Th that we observe is clearly hydrothermal in origin, and we present supporting
246 evidence in the following sections.

247

248 **4.2 Hydrothermal Plume and Effects on Dissolved ^{230}Th Concentrations below 2000** 249 **m**

250 Hydrothermal activity plays a vital role in the chemistry of the oceans—
251 particularly in the Pacific Ocean, where ridge spreading is fast and voluminous.
252 Emanation and subsequent dispersion of mantle ^3He from the EPR into deep ocean
253 basins of the Pacific has been used to assess ocean circulation and mixing patterns

(Lupton and Craig, 1981; Lupton, 1998). In addition, radioisotopes have been used to quantify and effectively explore rates of chemical processes within hydrothermal plumes and the role these plumes play in the regulation of ocean chemistry (Kadko, 1996). Hydrothermal plumes are not only enriched in ^3He , but also in particulates enriched in metalliferous Mn and Fe. Such particulate plumes can be laterally transported from the ridge crest to the interior of oceans (e.g. Kadko, 1996) and therefore affect the scavenging of trace elements (such as particle-reactive Th) within the water column (Trefry and Metz, 1989; Kadko, 1996). Here, we assess whether such scavenging is the cause of the observed depletions in ^{230}Th below 2000 m at our sites that are more than 5,000 km away from the EPR ($\sim 110^\circ\text{W}$ - 95°W).

To further support the idea that hydrothermal particulate scavenging of ^{230}Th can explain the trends in our Line Islands ^{230}Th profiles, we first need to evaluate evidence for migration of the EPR plume to the CEP. The first evidence for the far-field westward migration of the hydrothermal plume in the Pacific Ocean is the original ^3He concentration data from the southern and northern Pacific Ocean from the Kermadec Trench (NOVA VI-2) and North Pacific (GEOSECS-I) stations, respectively (Craig et al., 1975; Clarke et al., 1970). ^3He profiles far from the plume showed almost identical patterns, with the highest $\delta^3\text{He}$ values between depths of 2000 – 3000 m, and maxima at 2500 m (Clarke and Craig, 1970). Since those first studies, other studies have shown similar profiles westward from the EPR throughout the Pacific Ocean (e.g., Lupton and Craig, 1981; Lupton, 1998; Boyle and Jenkins, 2008; German and Seyfried, 2014). A recent synthesis of the ^3He data (German and Seyfried, 2014: their Figure 19) identifies

276 the existence of two ^3He plumes with a maximum helium concentration at 2500 m, one
277 centered between approximately $8^\circ - 12^\circ\text{N}$ north of the equator, and the other centered
278 between about $15^\circ - 5^\circ\text{S}$. If the ^3He data represent the true migration of the EPR
279 hydrothermal plume, then manganese and iron profiles should be similar to those of
280 helium. Indeed, this is the case (Klinkhammer and Hudson, 1986; Hudson et al., 1986;
281 Wu et al., 2011; Fitzsimmons et al., 2014). Initial Mn work in the South Pacific showed
282 dispersion patterns for hydrothermal Mn that suggests the existence of off-axis Mn
283 anomalies at the same depths as helium anomalies originating from the EPR, and
284 dispersed westward as far as ~ 2000 km (Klinkhammer and Hudson, 1986; Hudson et al.,
285 1986). Complementary initial Fe data near the crest of the EPR at 19°S showed the same
286 patterns in profile as those of Mn (Klinkhammer and Hudson, 1986; Hudson et al.,
287 1986). More recent work with respect to both Fe and Mn has been completed by Wu et
288 al. (2011) and Fitzsimmons et al. (2014). The latter study found a mid-depth Fe maxima
289 (between 2000 and 3000 m, peaking at ~ 2500 m) coincident with a ^3He concentration
290 maxima in nearby tropical Pacific waters of the southern hemisphere (20°S , 170°W),
291 ~ 6000 km (Fitzsimmons et al., 2014) westward of the EPR hydrothermal vents. More
292 importantly our anomalous ^{230}Th depletions are the greatest between these same water
293 depths (Figures 6) suggesting a hydrothermal role as an explanation for Th
294 concentrations on a large Pacific-basin-wide scale.

295 The Fe study closer to our sites (along 158°W ; Wu et al., 2011) also indicate
296 maximum Fe concentrations in mid-depth waters, but over a wider interval in the water
297 column between 1000 and 3000 m. The correlation between our ^{230}Th results and the Fe

298 results from Wu et al. (2011) is not as clearly correlated (Figure 7). Specifically, Wu et
299 al. (2011) observed a lens of elevated dissolved Fe concentrations at depths of slightly
300 less than 1000 m (Figure 7) that is not associated with lower ^{230}Th concentrations
301 observed within our sites, as Fe concentrations are deeper in the water column (>2000
302 m). Wu et al. (2011) attribute their elevated Fe to a dispersal of Fe from the Loihi
303 seamount located approximately 2000 km (Wu et al., 2011). This dispersal might be the
304 cause of the higher concentrations of Fe over a wider range of depths. We do not have
305 sufficient data to ascertain the cause of the discrepancy between Fe and ^{230}Th in the
306 upper water column (<1000 m).

307 Two hypotheses can be formulated to explain the depletion of ^{230}Th in Pacific
308 waters enriched in ^3He and Fe between 2000 and 3000 m: a) Fe-Mn particulates from the
309 EPR hydrothermal plume were transported 5000 km west to our site, after which ^{230}Th
310 scavenging took place *in situ* and the ^{230}Th depleted signatures were formed, or b) ^{230}Th
311 was scavenged by Fe-Mn particulates closer to the EPR, after which ^{230}Th -depleted
312 waters were transported westward to the locations investigated here. Arguing against the
313 first hypothesis is the fact that the particulate Th fraction within the 2000-3000 m
314 interval is quite low (10-20%) at both sites. If *in situ* particulates were the cause of the
315 ^{230}Th depletion, significantly higher particulate ^{230}Th concentrations would be expected.
316 Nevertheless, in each scenario, knowledge of the movement of mid-depth water between
317 2000 to 3000 m, likely Pacific Deep Water (PDW), is necessary.

318 Reid (1997) has calculated geostrophic flow streamlines (black lines with arrows;
319 Figure 1) for Pacific Deep Water, and we reproduce these streamlines at 2500 db (~2550

320 m depth) for the region of the Pacific within about 10-15° of the equator. Waters which
321 bathe the EPR hydrothermal plume travel westward within a band of latitude from about
322 5-15°N and S. These same waters veer towards the equator in the western Pacific in both
323 the southern and northern hemispheres, and then return eastward along the equator
324 within a band of about 5° (N and S). The rates of circulation at this depth have been
325 estimated to be about 0.2-0.5 cm/s (Hautala and Riser, 1993; Fitzsimmons et al., 2014).
326 Consequently, such rates would correspond to travel times of about 30-80 years from the
327 EPR to the 8°N site. The fact that this decadal timeframe is within the range of oceanic
328 residence times for ^{230}Th in the Pacific suggests that lateral transport of waters with a
329 low dissolved- ^{230}Th signature is possible (i.e., retention of a signature derived from close
330 to the hydrothermal plume site with insufficient time for the waters to re-equilibrate with
331 respect to ^{230}Th scavenging). The residence time of ^{230}Th , along with the anomalous
332 non-equilibrium profiles we observe, can provide constraints on the transport time for
333 these mid-depth waters. Typical ^{230}Th residence times of a few decades suggest fast
334 transport times with water velocities closer to the upper end (0.5 cm/s) of that estimated
335 by Hautala and Riser, 1993). An additional factor pointing toward rapid transport times is
336 the fact that mid-depth waters with a low- ^{230}Th signature are still observed at the equator,
337 i.e., equilibrium reversible scavenging profiles have not yet been regained. Mid-depth
338 waters at the equator have traveled a significantly farther distance from the EPR (bottom
339 limb of cells; Figure 1) and should, therefore, lie closer to the dissolved ^{230}Th
340 equilibrium model. Indeed, lower % depletion values (i.e., deviation from ideal profiles) at

341 the equator (18%) compared to those at 8°N (27%) suggest longer dissolved ^{230}Th
342 transport times from the original site of scavenging.

343 If our reasoning is correct, one would expect ^{230}Th depletions in mid-depth
344 waters to be the greatest closest to the EPR, with a reduction in depletion further away
345 and along the flow path of these waters. There is one site (HY-3, Okubo et al., 2012), in
346 addition to the sites studied here, for which a dissolved ^{230}Th profile has been measured
347 along the northern-hemisphere mid-depth water flow paths presented in Figure 1. HY-3
348 is a site in the eastern equatorial Pacific (8°N, 95°W), for which the ^{230}Th profile is
349 variable and significant depletions from reversible scavenging throughout the water
350 column occur. ^{230}Th concentrations range from 0.05 to 0.72 dpm/m³ with the lowest
351 concentrations occurring in mid-depth waters. Okubo et al. (2012) hypothesize the
352 depletions to be due to enhanced scavenging of ^{230}Th by particulates supplied from the
353 continental slope. HY-3, however, is located in the open ocean, approximately 1000 km
354 from the margin. Hence, another mechanism more likely explains the depletion. We
355 believe hydrothermal scavenging to be that mechanism. Below 1400 m, the lowest ^{230}Th
356 concentrations are found between 2000 m (0.30dpm/m³) and 3000 m (0.39dpm/m³) with
357 a peak ^{230}Th depletion occurring at 2500 m (~0.05 dpm/m³), coincident with the depth of
358 maximum ^3He and Fe. Therefore, we propose that the extremely low ^{230}Th
359 concentrations at HY-3 (Okubo et al., 2012) are a function of the proximity of this site to
360 the EPR hydrothermal plume. Additional measurements of dissolved ^{230}Th profiles for
361 waters along a zonal transect from the EPR towards the central Pacific in both the
362 northern and southern hemispheres are needed to further test this hypothesis.

363

364 **4.3 Dissolved ^{230}Th concentrations below 4000 m**

365 For the deeper waters below 4000 m at the North Pacific gyre site, dissolved
366 ^{230}Th concentrations are still depleted, i.e., concentrations that are lower than those
367 predicted from the reversible scavenging model (Figure 5). Unlike the waters
368 between 2000 and 3000 m, however, deep waters contain a particulate fraction that
369 makes up ~30-40% of the total ^{230}Th (Figure 2; Table 1). In fact, there is up to four
370 times as much particulate ^{230}Th in deep waters compared to that in the mid-depth waters
371 between 2000-3000 m. Given the enhanced particulate ^{230}Th concentrations, we interpret
372 the deviations from equilibrium scavenging in the deepest water at 8°N to reflect
373 enhanced bottom scavenging from resuspended sediment and/or a nepheloid layer, as
374 others have suggested in several regions of the Pacific (Okubo et al., 2014; Lyle et al.,
375 2014). Resuspension of sediment could take place as a consequence of interaction
376 between currents and seamounts (Turnewitsch et al., 2013; Lyle et al., 2014), that occur
377 throughout the Line Islands region and close to our 8°N site (Sager and Keating, 1984).

378

379 **4.4 Dust and dissolved Fe flux estimates using ^{232}Th – ^{230}Th Systematics**

380 Dissolution of dust is the main source of dissolved ^{232}Th to the CEP because of
381 its remoteness from the continental margins (i.e., lack of blurring effects from
382 hemipelagic detrital material). The advantage of using the two-isotope thorium system is
383 that the dissolved ^{230}Th concentration data can be used to constrain the flux of the
384 dissolved ^{232}Th (the proxy for dust input). Specifically, the supply of dissolved ^{230}Th in

the water column can be determined by its rate of production through the decay of its parent ^{234}U which, in turn, allows us to calculate residence times for ^{230}Th with respect to its removal throughout the water column by scavenging. Assuming that ^{230}Th and ^{232}Th have the same residence time, the removal rate of dissolved ^{232}Th within an interval of the water column can be calculated by dividing the measured inventory of ^{232}Th by the ^{230}Th residence time within the same interval. This is a reasonable assumption as long as there is not a significant amount of ^{232}Th going into a colloidal phase upon dissolution of dust in seawater (Hayes et al., 2015). The interval of the water column from 0-500 m has been argued by Hayes et al. (2013a) to be the most affected by dust input to the surface ocean, and we follow their methodology here.

Assuming steady state for the top 500 m, the scavenging removal rate of ^{230}Th or ^{232}Th is equal to the inverse of the ^{230}Th residence time, τ_{Th} , and can be calculated using the measured dissolved ^{230}Th concentrations from each station in the CEP, and comparing it to the production of ^{230}Th by ^{234}U decay within the same interval. Once the 500-m ^{230}Th residence time is calculated, it is applied to the integrated inventory of dissolved ^{232}Th in order to approximate the ^{232}Th flux due to mineral dust dissolution from 0 to 500 m (Hayes et al., 2013a). As described by Hsieh et al. (2011) and Hayes et al. (2013a), in order to estimate the lithogenic flux, the concentration of dissolved ^{232}Th , $[\text{}^{232}\text{Th}]_{\text{litho}}$, and its fractional solubility, S_{Th} , are necessary (see equation 4). Note, for the North Pacific Gyre, a linear interpolation was necessary to determine the dissolved ^{230}Th and ^{232}Th concentrations at 500 m. Such an interpolation is justified for ^{230}Th given its reversible behavior to 2000 m. For ^{232}Th , there is little variability

$$\text{Lithogenic Flux (z)} = \frac{\int_{z_0}^{z_f} \text{Dissolved } ^{232}\text{Th dz}}{\tau^{230}\text{Th(z)} * [^{232}\text{Th}]_{\text{litho}} * S_{\text{Th}}} \quad (4)$$

407 with depth.

408 The residence times of the ^{230}Th in the top 500 m at the Equator and North
 409 Pacific Gyre sites are approximately 3.3 and 4.1 years, respectively. These surface-ocean
 410 ^{230}Th residence times are similar to those found throughout the Pacific Ocean and
 411 recently summarized by others (Hayes et al., 2013a; Deng et al., 2014). With regard to
 412 dissolved ^{232}Th fluxes, approximations of the concentrations and fractional solubility of
 413 ^{232}Th in the dust are essential. Most studies suggest normalizing ^{232}Th lithogenic
 414 concentrations to an averaged upper continental crust value of 10.7 ppm (Taylor and
 415 McLennan, 1995; McGee et al., 2007; Winckler et al., 2008; Woodard et al., 2012).
 416 Fractional solubility of Th in dust, S_{Th} , is more difficult to assess. We have utilized an
 417 estimated value of 20% that falls within the range of previous studies (Hsieh et al., 2011;
 418 Hayes et al., 2013a). Calculated removal fluxes for ^{232}Th range from $1.1 - 1.4 \mu\text{g m}^{-2} \text{yr}^{-1}$
 419 ¹ for the upper ocean in the Central Equatorial Pacific (Table 3).

420 Central Equatorial Pacific dust deposition estimates determined via integration of
 421 ^{232}Th inventories from 0 – 500 m are approximately $0.6 \text{ g m}^{-2} \text{yr}^{-1}$ (Table 3), and
 422 complement the values for the subtropical North Pacific estimated by Hayes et al.
 423 (2013a). The low dust fluxes observed are consistent with the remote location of the
 424 CEP. We can also compare our 500-m dissolved ^{232}Th dust flux values to dust
 425 deposition estimates from marine sediments. For the Holocene period, rates of supply of
 426 dust delivered to the equatorial Pacific seafloor range from $0.13 - 0.55 \text{ g m}^{-2} \text{yr}^{-1}$

(McGee et al., 2007; Winckler et al., 2008). Our values are on the high end, but in agreement with, these sedimentary estimates, and are significantly higher than estimates given by dust models for the equatorial Pacific for which dust fluxes span from 0.01 – 0.1 g m⁻² yr⁻¹ (Figure 2 in Mahowald et al., 2005). In order to decrease our thorium-derived estimate of dust flux to levels consistent with dust models would require either *increasing* [²³²Th_{litho}] or S_{Th}, or a combination of both, by at least a factor of 6.

The supply of Fe can be determined by estimating the ratio of the partial dissolution of Fe to Th in the dust (S_{Fe}/S_{Th}), and by assuming an average Fe/Th concentration ratio in the dust that is equal to the concentration ratio in the continental crust (A_{Fe}/A_{Th}). The dissolved Fe supply to the sea surface can then be estimated as follows (Hayes et al., 2013a):

$$\text{Dissolved Fe Supply} = \text{Dissolved } ^{232}\text{Th flux} * (A_{\text{Fe}}/A_{\text{Th}}) * (S_{\text{Fe}}/S_{\text{Th}}) \quad (5)$$

Application of equation (5) results in dissolved Fe fluxes of approximately 4 (North Pacific Gyre) to 5 mg Fe m⁻²yr⁻¹ (Equator) in the CEP (Table 3), using our estimated dissolved ²³²Th fluxes (Table 3). We assume the same values used by Hayes et al. (2014) for A_{Fe}/A_{Th} (3262 g/g) and S_{Fe}/S_{Th} (1). Our dissolved Fe fluxes are lower than those estimated at sites in the North Pacific (15 mg Fe m⁻²yr⁻¹) by Hayes et al. (2013a), which might be expected given the remoteness of our sites from the continents. Estimates of dissolved Fe flux based on models of total Fe deposition to the equatorial Pacific (Mahowald et al., 2009) are lower (<2 mg Fe m⁻²yr⁻¹) by more than a factor of 2 than the estimates we determine here using the thorium systematics.

447 Although Pacific model dust and dissolved Fe fluxes are lower than those
448 estimated empirically by us and Hayes et al. (2013a) using similar thorium systematics,
449 Hsieh et al. (2011) measured dust fluxes in the Atlantic that agree with those modeled
450 by Mahowald et al. (2005). The agreement between Atlantic dust models and
451 measurements may be due to the high supply of dust to the Atlantic that makes
452 measurements easier and models more robust. On the other hand, in the Pacific, the
453 miniscule amounts of dust delivered to the oceans magnify the inherent errors.
454 Sensitivity to the unconstrained values of S_{Th} and S_{Fe}/S_{Th} could also explain the
455 difference between models and measurements in the Pacific. Experimental determination
456 of the solubilities of Th and Fe in aerosol dust needs to take place before applying long-
457 lived Th isotopes as a quantitative tool to estimate the supply of trace elements from
458 lithogenic sources to the upper ocean.

459

460 **5. CONCLUSIONS**

461 Low concentrations of ^{230}Th at the equator and 8°N in CEP mid-depth (2000-
462 3000 m) waters result from far-field hydrothermal scavenging by a hydrothermal plume
463 emanating from the East Pacific Rise approximately 5000 km away from our sites. This
464 is further supported by dissolved Fe, dissolved Mn and 3He studies conducted in the near
465 vicinity of our study area, which suggest that these hydrothermal vent fluids have
466 retained their signature throughout the lifetime of the plume and have traveled ~ 5000
467 km westward to our sites. The depleted ^{230}Th signatures are most likely a product of

468 ^{230}Th scavenging by Fe-Mn particulates close to the EPR, and then transit of these
469 waters to the CEP.

470 The North Pacific gyre ^{230}Th concentrations below 3000 are also lower than
471 those expected from reversible scavenging behavior. These deep waters, especially those
472 lower than 4000 m, are likely depleted of ^{230}Th because of enhanced bottom scavenging
473 from resuspended sediment and/or a nepheloid layer, as others have noted for the
474 equatorial Pacific (Lyle et al., 2014).

475 Finally, our dissolved ^{232}Th concentrations are relatively invariable with depth
476 throughout the water column for both sites. Dissolved ^{232}Th concentrations are likely a
477 result of dissolution of dust in the upper ocean. By combining our measured seawater
478 $^{232}\text{Th} - ^{230}\text{Th}$ concentrations within the topmost water column, as done in recent studies
479 (Hsieh et al., 2011; Hayes et al., 2013; Deng et al., 2014), we have been able to determine
480 lithogenic fluxes (based on dissolved ^{232}Th) to be between $0.5 - 0.6 \text{ g m}^{-2} \text{ yr}^{-1}$ in the
481 CEP. Such values are significantly higher than those determined in models (e.g.,
482 Mahowald et al., 2005; Jickells et al., 2005).

483

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492 (<http://www.ncdc.noaa.gov/data-access/paleoclimatology-data>) and the Biological and
493 Chemical Oceanography Data Management Office (<http://www.bco-dmo.org/>).

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692

FIGURE CAPTIONS

Figure 1. Location map of study sites (solid circles) near the Line Islands, Central Equatorial Pacific: ML1208-03CTD ($00^{\circ} 13.166' \text{ S}$, $155^{\circ} 57.668' \text{ W}$) and ML1208-12CTD ($8^{\circ} 19.989' \text{ N}$, $159^{\circ} 18.000' \text{ W}$). HY-3 (solid square), a site studied by Okubo et al. (2012), is also discussed in the text. Approximate geostrophic flow streamlines (black lines with arrows) for Pacific Deep Water at 2500 db shown only for the region of the Pacific within about $10\text{--}15^{\circ}$ of the equator (from Reid, 1997). Location of the East Pacific Rise (EPR) in the Northern and Southern Pacific is also shown.

Figure 2. Filtered (closed symbol) and unfiltered (open symbol) ^{230}Th concentrations in the Central Equatorial Pacific (0° -black circle; 8°N -red circle). ^{230}Th error bars are smaller than symbol size (see Tables 1, 2 and text for discussion).

Figure 3. Filtered (closed symbol) and unfiltered (open symbol) ^{232}Th concentrations in the Central Equatorial Pacific (0° -purple; 8°N -maroon). ^{232}Th error bars are smaller than symbol size (see Tables 1, 2 and text for discussion).

Figure 4. Dissolved ^{230}Th concentration profile at 8°N (red) and modeled dissolved ^{230}Th from reversible scavenging model (black line). ^{230}Th error bars are smaller than symbol size (see Tables 1, 2 and text for discussion).

Figure 5. Dissolved ^{230}Th concentration profile at 0° (red) and modeled dissolved ^{230}Th from reversible scavenging model (black line). ^{230}Th error bars are smaller than symbol size (see Tables 1, 2 and text for discussion).

Figure 6. Excess ^3He (blue) and dissolved Fe (green) concentrations in the South Pacific replotted from Fitzsimmons et al. (2014) (20°S , 170°W). Dissolved ^{230}Th ($d^{230}\text{Th}$) concentrations (red and black) at our sites.

Figure 7. Dissolved Fe concentrations in the CEP replotted from Wu et al. (2011) (8°N , 158°W —dark green; 0°N , 158°W —light green). Dissolved ^{230}Th ($d^{230}\text{Th}$) concentrations (red and black) at our sites.

Figure 1

736 **Figure 2**
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739 **Figure 3**
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743 **Figure 4**

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754 **Figure 7**

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757 **LIST OF TABLES**

758

759 **Table 1.**Data for 8°N. Shaded columns represent unfiltered Th; non-shaded columns
760 represent filtered Th concentrations. Particulate ^{230}Th % = $100 \times (\text{unfiltered} -$
761 $\text{filtered}) / (\text{unfiltered})$. (* = no data)

762

763 **Table 2.**Data for 0°. Shaded columns represent unfiltered Th; non-shaded columns
764 represent filtered Th concentrations. Particulate ^{230}Th % = $100 \times (\text{unfiltered} -$
765 $\text{filtered}) / (\text{unfiltered})$. (* = no data)

766

767 **Table 3.** Dust and dissolved Fe flux estimates using dissolved ^{230}Th - ^{232}Th systematics in
768 the upper 500 m of the water column at both sites.

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Table 1.

Water Cast ID:	Depth (m)	Filtered	Filtered	Unfiltered	Unfiltered	Particulate
8°N		²³⁰ Th (fg/kg)	²³² Th (pg/kg)	²³⁰ Th (fg/kg)	²³² Th (pg/kg)	²³⁰ Th %
ML1208-12CTD	100	1.1±0.1	10.2±1.0	1.1±0.1	5.2±0.5	-
ML1208-12CTD	900	10.3±0.6	8.1±0.8	10.5±0.5	10.8±1.1	2
ML1208-12CTD	1500	18.9±1.1	11.7±1.2	18.5±0.9	13.6±1.4	-
ML1208-12CTD	1800	18.6±1.1	8.4±0.8	22.5±1.1	10.3±1.0	17
ML1208-12CTD	2100	22.4±1.3	*	24.8±1.2	11.3±1.1	10
ML1208-12CTD	2400	22.6±1.4	8.8±0.9	25.7±1.3	10.7±1.1	11
ML1208-12CTD	2701	22.0±1.3	10.6±1.1	24.7±1.2	12.6±1.3	11
ML1208-12CTD	3001	22.0±1.3	19.5±2.0	*	*	*
ML1208-12CTD	3300	26.4±1.6	15.0±1.5	29.0±1.5	18.0±1.8	9
ML1208-12CTD	3600	*	*	32.3±1.6	139±1.4	*
ML1208-12CTD	3900	*	*	37.1±1.9	27.6±2.8	*
ML1208-12CTD	4201	29.0±1.7	16.7±1.7	41.2±2.1	32.3±3.2	29
ML1208-12CTD	4400	30.8±1.8	19.7±2.0	49.9±2.5	45.1±4.5	38
ML1208-12CTD	4601	*	*	55.2±2.8	53.9±5.4	*

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Table 2.

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Water Cast ID: 0°N	Depth (m)	Filtered ²³⁰ Th (fg/kg)	Filtered ²³² Th (pg/kg)	Unfiltered ²³⁰ Th (fg/kg)	Unfiltered ²³² Th (pg/kg)	Particulate ²³⁰ Th %
ML1208-03CTD	25.4	Below detection limit	3.9±0.3	2.2±0.1	19.3±1.9	100
ML1208-03CTD	147.6	2.3±0.1	12.8±1.3	4.3±0.3	29.7±3.0	48
ML1208-03CTD	499.5	4.1±0.2	6.2±0.6	5.8±0.3	17.5±1.8	29
ML1208-03CTD	801.1	5.9±0.3	6.1±0.6	8.5±0.5	30.2±3.0	31
ML1208-03CTD	1001.8	7.9±0.4	12.3±1.2	10.2±0.6	15.6±1.6	23
ML1208-03CTD	1200.3	8.7±0.4	5.8±0.6	9.6±0.6	16.6±1.7	10
ML1208-03CTD	1399.9	11.3±0.6	10.2±1.0	15.2±0.9	11.5±1.2	26
ML1208-03CTD	1600.5	12.1±0.6	6.0±0.6	14.4±0.9	19.4±1.9	16
ML1208-03CTD	1801	15.9±0.8	14.0±1.4	16.5±1.0	17.4±1.7	4
ML1208-03CTD	2001	14.7±0.7	12.9±1.3	17.7±1.0	25.3±2.5	17
ML1208-03CTD	2200	14.4±0.7	8.2±0.8	16.8±1.0	14.5±1.5	14
ML1208-03CTD	2401	14.1±0.7	6.9±0.7	17.5±1.1	14.3±1.4	20
ML1208-03CTD	2602	14.2±0.7	5.4±0.5	17.3±1.0	15.5±1.6	18
ML1208-03CTD	2749	17.0±0.9	7.1±0.7	20.2±1.2	24.2±2.4	16
ML1208-03CTD	2900	16.8±0.8	6.6±0.7	19.9±1.2	18.4±1.8	1
ML1208-03CTD	3038	19.0±1.0	8.1±0.8	25.4±1.5	325±33	25

776 **Table 3.**
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Station	$\tau^{230}\text{Th}$ (500 m, years)	Dissolved ^{232}Th flux ($\mu\text{g m}^{-2} \text{ yr}^{-1}$)	^{232}Th -derived Dust Flux ($\text{g m}^{-2} \text{ yr}^{-1}$) $S_{\text{Th}} = 20\%$	Dissolved Fe Flux ($\text{mg Fe m}^{-2} \text{ yr}^{-1}$)
ML1208-03CTD: Equator	3.3	1.4	0.6	5
ML1208-12CTD: N. Pacific Gyre (8°N)	4.1	1.1	0.5	4