Geophysical Research Letters

RESEARCH LETTER

10.1029/2020GL087266

Key Points:

- The Southern Ocean contains an inventory of hydrothermally-sourced ^3He
- If we divide this inventory by the global hydrothermal flux that must support this inventory, we obtain an estimate of an upwelling time scale for deep waters in the Southern Ocean

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Citation:

Jenkins, W. J. (2020). Using excess ^3He to estimate Southern Ocean upwelling time scales. *Geophysical Research Letters*, 47, e2020GL087266. <https://doi.org/10.1029/2020GL087266>

Received 28 JAN 2020

Accepted 28 JUL 2020

Accepted article online 4 AUG 2020

Using Excess ^3He to Estimate Southern Ocean Upwelling Time Scales

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Abstract Using a recently compiled global marine data set of dissolved helium isotopes and helium and neon concentrations, we make an estimate of the inventory of hydrothermal ^3He in the Southern Ocean to be $4.9 \pm 0.6 \times 10^4$ moles. Under the assumption that the bulk of the hydrothermally sourced ^3He is upwelled there, we use recent estimates of the global hydrothermal ^3He flux to determine an e -folding residence time of 99 ± 18 years, depending on assumptions of water mass and upwelling boundaries. Our estimate is within the broad range of values obtained from recent Southern Ocean circulation models.

Plain Language Summary Seafloor hydrothermal systems inject significant amounts of dissolved iron, but it remains unknown what fraction of that trace element survives in situ chemical scavenging to upwell in the Southern Ocean to arrive at the sea surface. Since iron is a necessary trace-level nutrient for marine biological productivity, knowledge of the time it takes for deep water to rise to the surface is important. Along with dissolved iron, helium-3, a rare, inert isotope, is also injected into the deep water. Since the global flux of this isotope has been independently determined by large-scale ocean models, we can estimate the upwelling time scale by dividing it into the inventory of this isotope in the Southern Ocean.

1. Introduction

Recent observations of the large-scale distributions of dissolved iron and manganese in the abyssal South Pacific suggest that ocean ridge crest hydrothermal activity produces a significant flux of a dissolved phase for these metals that persists for many thousands of kilometers away from their injection sites (Fitzsimmons et al., 2014, 2017; Resing et al., 2015). These large-scale horizontal plumes are also rich in hydrothermal ^3He (Jenkins et al., 2018; Resing et al., 2015). ^3He is an inert, unambiguous tracer of hydrothermal activity that has well-established global flux of $548 \pm 52 \text{ mol a}^{-1}$ (the weighted mean of estimates by Bianchi et al., 2010; DeVries & Holzer, 2019; Schlitzer, 2016). This allows the use of ^3He as a “flux gauge” to calculate the global hydrothermal fluxes for these trace metals by determining their concentrations relative to ^3He (Resing et al., 2015). The question of how this abyssal source of trace metals may eventually contribute to and influence global oceanic biological productivity pivots on what fraction of these metal fluxes survive deep water in situ scavenging processes (Bergquist & Boyle, 2006; Bruland et al., 1994; Fitzsimmons et al., 2014, 2017). A first step in addressing this question involves establishing the time it takes for deep-water fluid parcels to upwell in the Southern Ocean.

We begin by describing the distribution in terms of the most commonly used measure of nonatmospheric helium in the ocean, the helium isotope ratio anomaly

$$\delta^3\text{He} = \left(\frac{\left(\frac{^3\text{He}}{^4\text{He}} \right)_{\text{X}}}{\left(\frac{^3\text{He}}{^4\text{He}} \right)_{\text{Atm}}} - 1 \right) \times 100\% \quad (1)$$

Hydrothermal helium introduced into the ocean is highly enriched in ^3He (Jenkins et al., 1978; Lupton & Craig, 1981), the rarer isotope of helium (Aldrich & Nier, 1948). We use a recently compiled global oceanic data set of helium isotope and dissolved helium and neon concentration measurements retrieved from a variety of sources (Jenkins, Doney, et al., 2019), and it is publicly available online (<https://www.nodc.noaa.gov/ocads/data/0176626.xml>). This data set presently contains approximately 63,000 $\delta^3\text{He}$, 57,000

Figure 1. The distribution of $\delta^3\text{He}$ (in %) on a key neutral surface ($\sigma_N = 28.0 \text{ kg m}^{-3}$). This density horizon corresponds to the core of the South Pacific hydrothermal ^3He plume (Jenkins et al., 2018). The main features include the dominance of the Pacific deep water, a weaker but significant contribution from the Indian Ocean, and the ^3He -impoverished North Atlantic deep water in the Atlantic sector. Note also the low ^3He waters surrounding the Antarctic continent. The gray dots denote station locations. The white lines (partially obscured by station marks) correspond to the sections shown in the next figure.

dissolved helium concentration, and 34,000 dissolved neon concentration measurements. Although some of these measurements were made by the author, the bulk were made by other investigators, who are identified in the database along with other metadata including quality control flags (Jenkins, Doney, et al., 2019). We initially describe only the $\delta^3\text{He}$ distribution from the perspective of the Southern Ocean, while the concentration data will become necessary for subsequent isotopic flux calculations later in this discussion.

2. The Distribution of $\delta^3\text{He}$ in the Southern Ocean

Figure 1 is a map of $\delta^3\text{He}$ interpolated onto the 28.0 kg m^{-3} neutral density anomaly (computed according to Jackett & McDougall, 1997) surface. This level corresponds approximately to the core of the large ^3He plume observed to emanate from the East Pacific Rise at 15°S (Jenkins et al., 2018; Lupton & Craig, 1981). It appears to track the major trajectory of the ^3He -laden Pacific deep water, which moves southward along the eastern boundary of the South Pacific and ultimately around the southern tip of South America (e.g., Well et al., 2003). By far the dominant feature of the global distribution on this horizon is the contrast between very high values (red trending to pink and white) in the Pacific and the relatively ^3He -impoverished waters (aqua to blue) of the Atlantic sector, with the Indian Ocean being intermediate (green to yellow). This contrast arises in large part from the direction of the planetary-scale meridional overturning circulation (the MOC; e.g., Kuhlbrodt et al., 2007). Newly ventilated waters move southward from the subarctic North Atlantic to join the Antarctic Circumpolar Current (ACC) after it passes through the Drake Passage. From there, some of the Antarctic Circumpolar Deep Water (ACDW) mixture “peels off” to enter the abyssal Indian and then Pacific Oceans along their respective western boundaries. Within those basins, generalized upwelling feeds a returning

middepth flow toward the Southern Ocean, where it ultimately upwells to the surface (Lumpkin & Speer, 2007; Talley, 2013) or forms intermediate waters. On these travels, the water accumulates hydrothermal ^3He (along with hydrothermal iron and manganese) from seafloor injection sites as it ages. The global contrast is accentuated by differences in mid-ocean ridge spreading rates between the Pacific and the other basins—the fastest spreading ridges tend to occur in the Pacific Ocean sector, while the Atlantic and Indian oceans typically have slow spreading ridges. The flux of a primitive volatile like ^3He should be related to seafloor spreading rates. Thus, the deep water in the Indian Ocean appears intermediate in $\delta^3\text{He}$ both because of its slower spreading ridges and its place in the MOC path.

The southward propagation of Pacific deep water ^3He appears most pronounced in the southeast Pacific basin near the southern tip of South America. On the opposite side, between New Zealand and Australia, there is a northward bowing of $\delta^3\text{He}$ isopleths associated with the incursion of less $\delta^3\text{He}$ -enriched circumpolar waters into the Pacific (Stommel & Aarons, 1960; Warren, 1971). There is also evidence of a similar feature in the southwest Indian Ocean, particularly around Madagascar expected from the deep western boundary current observed there (Warren, 1974). The Southern Ocean/ACC regime appears as a pale blue band encircling the Antarctic continent, seemingly pushed northward by ^3He -impoverished (here depicted as darker blue) waters emanating from the Antarctic continent. This is especially evident in the Weddell Sea and to a lesser extent west of the Ross Sea and the Antarctic continental slope in between.

The three white lines in Figure 1 that extend radially from the coast of Antarctica around 30°E , 145°E , and 60°W mark hydrographic sections at three points along the Circumpolar Current that are shown in Figure 2. Note the color map used for these sections differs from that used in Figure 1 to match the observed range of values in those sections and to make the vertical and horizontal structure observed in the sections more

Figure 2. The depth distributions of $\delta^3\text{He}$ for the three sections denoted by the solid white lines in Figure 1. Note the different $\delta^3\text{He}$ color scale. Gray dots signify sample locations. The dashed white line corresponds to $\delta^3\text{He} = 9\%$. The black contours are neutral density with an interval of 0.1 kg m^{-3} . Antarctica is on the left for all three sections taken (a) from South Africa to Antarctic, (b) from Australia to Antarctica, and (c) across the Drake Passage. The map (d) shows the section locations overlaying austral winter sea surface temperature obtained from the World Ocean Atlas 2013 (Locarnini et al., 2013).

visible. For comparison purposes, the white dashed line contour marks the $\delta^3\text{He} = 9\%$ contour in each section. The black contour lines indicate neutral density anomaly levels (0.10 kg m^{-3} contour interval).

2.1. Antarctic to South Africa

There are three subsurface $\delta^3\text{He}$ maxima associated with the ACCW in this sector of the Southern Ocean (Figure 2a). The dominant feature, characterized by $\delta^3\text{He}$ values exceeding 9%, follows the 28.0 kg m^{-3} neutral surface, shoaling from $\sim 2,200 \text{ m}$ at about 45°S to $\sim 500 \text{ m}$ at about 60°S . This maximum is linked to the eastward Pacific deep water tongue in the Drake Passage that pivots around the tip of South America (Figure 2c). The northern extent of this tongue (around $54\text{--}55^\circ\text{S}$) has a significant northward decrease in nutrients along its isoneutral surface, marking a boundary between ACDW waters to the south and North Atlantic Deep Waters. To the north of this front is a weaker, less dense maximum centered on about 27.85 kg m^{-3} that extends to the South African coast. This corresponds to a silicate maximum that has been observed to extend throughout the South Atlantic (e.g., see Figure 4 of Jenkins & Clarke, 1976) and appears

in a much-diluted form in the subtropical North Atlantic. The deepest $\delta^3\text{He}$ maximum lies on a somewhat deeper neutral density anomaly horizon (28.2 kg m^{-3}) and extends further northward to around 43°S . As there appears to be no clear link with other hydrographic properties, it remains unclear whether this feature represents a unique water mass or is just an artifact of mixing between eastward flowing ACDW and northward moving Antarctic Bottom Water. Indeed, there is an appearance of this tongue spilling over the bottom topography. The ACDW complex is underlain by lower $\delta^3\text{He}$ Antarctic Bottom Water, with values around 5–6% in the south (between 51°S and the base of the Antarctic continental slope). To the north, a broad $\delta^3\text{He}$ minimum appears centered on 28.1 kg m^{-3} , the signature of lower nutrient, higher salinity North Atlantic Deep Waters being entrained into the eastward flowing ACDW, perhaps to partly turn northward into the western Indian Ocean.

2.2. Antarctic to Australia

The middepth ACDW maximum in $\delta^3\text{He}$ extends over most of this section (Figure 2b) and appears on a slightly denser neutral surface than the South African section, ranging from 28.05 kg m^{-3} in the south to 28.1 kg m^{-3} north of the Southeast Indian Ridge (SEIR). It approximately corresponds to a middepth salinity maximum as well as deep nitrate and phosphate secondary minima, although is somewhat shallower than these features south of 62°S and on average about 300 m deeper to the north. The pinching of the 9% $\delta^3\text{He}$ contour around 56°S corresponds to the minimum in eastward velocity south of the high velocity core of the ACC (Yaremchuck et al., 2001) and is suggestive of a layered structure (two maxima at 1,300 and 2,300 m, separated by a minimum in between). This appears to correspond to the northward deflection of transport streamlines as the ACC crosses the SEIR (McCartney & Donohue, 2007; Rintoul & Sokolov, 2001) and may reflect a meridional interleaving of the middepth water masses associated with the crossing of rough topography. The middepth $\delta^3\text{He}$ maximum is more intense, approaching 14%, to the north of the SEIR. It appears to highlight a deep anticyclonic recirculation exceeding $2 \times 10^7 \text{ m}^3 \text{ s}^{-1}$ identified by others (McCartney & Donohue, 2007; Rintoul & Bullister, 1999; Rintoul & Sokolov, 2001; Yaremchuck et al., 2001) with the southern part of the feature eastbound along the northern flank of the SEIR and the northern part westbound into the South Australia and Perth Basins. Examination of other WOCE helium/hydrographic sections to the west (I9S at $\sim 120^\circ\text{E}$, I8S at $\sim 85\text{--}95^\circ\text{E}$, and I7S at $\sim 60^\circ\text{E}$) within the context of the circulation scheme described by McCartney and Donohue (2007) reveals that the feature first appears at 90°E . Based on this, we conclude that the ^3He was acquired as the ACC passed through the Kerguelen-St. Paul Island Passage.

A distinct $\delta^3\text{He}$ maximum appears at intermediate depths (1,300–1,700 m) in the northernmost station (supported by several data points). Values here approach 13.6%, higher than anywhere else in this section at this neutral density anomaly (27.7 kg m^{-3}). This must arise from the southwestward intrusion of waters from the Tasman Sea, indicative of a deeper penetration Tasman limb of the southern hemispheric oceanic supergyre, which is more polar intensified on AAIW strata (Ridgway & Dunn, 2007), but suggesting that this “Tasman leakage” extends deeper than previously recognized. Thus, some Pacific-derived volcanic helium is leaking into the Indian Ocean at these depths.

2.3. Antarctic to South America (Drake Passage)

The Drake Passage (Figure 2c) represents the narrowest choke point for the ACC, where the current is forced to flow between the southern tip of South America and the northward extending West Antarctic Peninsula. At the southern side, there is a residual $\delta^3\text{He}$ maximum centered on the 28.1 kg m^{-3} surface but extending from 28.0 to 28.2 kg m^{-3} . The dominant feature of this section is the intense maximum along 28.0 kg m^{-3} in the north, a signature of ^3He -rich, deep Pacific waters leaking around the southern tip of South America (Well et al., 2003) into the eastward flowing ACC. Examination of three WOCE/CLIVAR meridional sections (P16S at $\sim 150^\circ\text{W}$, P18S at $\sim 105^\circ\text{W}$, and P19S at $\sim 88^\circ\text{W}$) reveals a progressive southward extension of a broad, middepth $\delta^3\text{He}$ maximum spanning a range of neutral surfaces from 27.8 to 28.1 kg m^{-3} , that culminates in the “leakage” around the southwest Chilean coast into the Passage. Within the Drake Passage, the anomaly appears to extend to the bottom with waters slightly greater than 28.2 kg m^{-3} . This is likely a result of intense vertical mixing associated with passage through the choke point. Interestingly, we observe a shallower maximum at about 1,500 m depth (27.8 kg m^{-3}) that occurs in only one station and was not observed by Well et al. (2003), which may indicate a spurious data point or a genuine feature that was

missed due to station spacing. Above the 27.8 and below the 28.2 kg m⁻³ isoneutral surfaces, the $\delta^3\text{He}$ and density contours appear to track within errors. South of the ACC, one finds relatively low $\delta^3\text{He}$ water near the bottom with neutral density anomalies greater than 28.3 kg m⁻³, which are associated with Antarctic Bottom Water originating from the Weddell Sea.

2.4. The General Distribution

The overall picture is one of the ACDW having a $\delta^3\text{He}$ of ~10% that is intermediate between the hydrothermally enriched deep Pacific (~30%) and the relatively ^3He -poor ventilated end-members from the North Atlantic and Antarctic slope and embayments. The deep Indian Ocean appears slightly elevated ($\delta^3\text{He}$ ~ 15%) compared to the ACDW. A simple mass transport calculation gives a rather similar ^3He transport across these three sections of the order of ~1,000 mol a⁻¹ (Jenkins et al., 2007), which is about twice the global hydrothermal ^3He flux (Bianchi et al., 2010; DeVries & Holzer, 2019; Schlitzer, 2016). In a sense, the ACDW is a “geochemical flywheel” that holds a hydrothermal ^3He inventory that passes through the Southern Ocean as it transits from the many seafloor injection points through its gradual and convoluted rise toward the ocean surface and its ultimate release to the atmosphere. Determining the magnitude of this inventory and scaling it against the known flux should yield a crude estimate of the upward transit time scale.

3. Determining the Southern Ocean Volcanic ^3He Inventories

3.1. Computing the Hydrothermal ^3He Contribution

Following earlier methods (Jenkins et al., 2018), we express the concentration of hydrothermal (“XS”) ^3He using the semiempirical relationship

$$C_{XS}(^3\text{He}) = R_A [(1 + \delta^3\text{He}/100)C_M(^4\text{He}) - \alpha_E C_S(^4\text{He}) - 0.22(C_M(\text{Ne}) - C_S(\text{Ne}))] \quad (2)$$

where subscripts *S* and *M* signify solubility equilibrium and measured concentrations, R_A the atmospheric helium ($^3\text{He}/^4\text{He}$) isotope ratio (Clarke et al., 1976), and α_E the helium solubility equilibrium isotope effect (Benson & Krause, 1980). The last term involving neon concentrations is an empirical correction for air injection by bubble dissolution near the sea surface and is generally a small but significant term. Solubility concentrations are temperature and salinity dependent (Jenkins, Lott, & Cahill, 2019) and are further corrected for a slight depth-dependent enhancement (~0.2 and ~0.3% for He and Ne, respectively) arising from diapycnal mixing effects (see Jenkins et al., 2018).

3.2. Calculating Water Column Inventories

While the bulk of the global deep water upwelling occurs within the Southern Ocean (Marshall & Speer, 2012), some fraction of the hydrothermal helium entering the Southern Ocean must mix downward into the bottom waters to be returned to other basins. Since the bottom waters are impoverished in ^3He , it is logical to draw the inventory boundary between the two water masses. We argue that the lower limit of the Circumpolar Deep Waters inventory is essentially the boundary between the ^3He -enriched returning deep waters and those bottom water components that are formed around Antarctica and ultimately flow northward into the interiors of the subtropical ocean basins. The choice of the neutral density of that surface has a significant effect on the computed inventory and could range between 28.11 (Orsi et al., 2002) and 28.18 kg m⁻³ (Orsi et al., 1999), or 28.20 kg m⁻³ (Speer et al., 2000), to as deep as 28.26 (Garabato et al., 2002) to 28.27 kg m⁻³ (Jullion et al., 2014). However, we choose a slightly denser horizon of 28.30 kg m⁻³ to include as much of the Southern Ocean hydrothermal ^3He inventory as possible. The 28.30 kg m⁻³ contour seen in the three “choke point” sections shown in Figure 2 delineates the downward penetration of the bulk of the hydrothermal ^3He mass, exemplified by the $\delta^3\text{He}$ = 9% contour. Winckler et al. (2010) reported a potential hydrothermal source in the Pacific sector of the Southern Ocean centered at the 28.2 kg m⁻³ neutral surface, which may contribute to the deeper tongue at the northern end of the Antarctic-South Africa section. Our choice thus includes that contribution.

The data set (Jenkins, Doney, et al., 2019) was filtered such that only data with corresponding neon measurements exhibited saturation anomalies between -5% and +20% to avoid obviously contaminated or compromised measurements. A station's hydrothermal ^3He inventory was computed using a tensioned smoothing

Figure 3. (a) The depth of the 28.00 kg m^{-3} neutral density anomaly surface as a function of latitude from the World Ocean 2013 Atlas. The red, green, and blue points correspond to the Pacific, Indian, and Atlantic Ocean sectors, respectively. (b) A map of the 28.00 kg m^{-3} neutral density anomaly surface 2,200 m depth contour. The southernmost contour is used to define the northern boundary of the integration volume.

spline (to avoid spurious peaks) and only accepted if there were at least four discrete measurements made within and spanning the density range of interest (between the surface and 28.30 kg m^{-3}). Profiles were visually scanned to ensure anomalous results were avoided. In all, inventories at 109 locations were computed within the geographic range of interest (see black dots in Figure 4).

3.3. Laterally Summing Water Column Inventories

The individual station inventories were gridded as a function of latitude and longitude onto a 1° resolution mesh using a GMT minimum curvature gridding routine (“surface,” using a tension of 0.25; Smith & Wessel, 1990; Wessel & Smith, 2008). The objective was to minimize the effects of sampling bias by smoothly extrapolating into areas of no/low coverage, avoiding generation of spurious extrema, and generally producing a reasonable looking distribution based on knowledge of other property distributions. Although other gridding methods will produce different results, variations in inventory estimates driven by uncertainties associated with the bounds of integration (both vertical and horizontal) make such refinements moot.

In order to establish the inventory, we define the geographical bounds of the ACDW ^3He distribution in the deep Southern Ocean and, more specifically, the northern extent of the region where upwelling may occur. Considering the large lateral volcanic ^3He plumes in the Pacific where the 28.00 kg m^{-3} neutral density surface begins to rise up toward the surface in its southward journey is an appropriate northern boundary. Figure 3a plots the depth of the 28.00 kg m^{-3} neutral density surface as a function of latitude for each ocean using the WOA13 (Locarnini et al., 2013; Zweng et al., 2013). We select the 2,200 m depth as the horizon where this upwelling nominally commences. Figure 3b is a map of the 2,200 m depth contour for the 28.00 kg m^{-3} surface (note we use the southernmost circum-Antarctic contour is the ACDW northern boundary). Figure 4 is a map of the resultant water column ^3He inventories in nmol m^{-2} above 28.3 kg m^{-3} horizon. The black dots signify the location of the stations used to construct the map. At each longitude, the gridded inventory distribution is integrated southward from this contour using area (latitude) weighting until either the 28.30 kg m^{-3} horizon outcrops or the 150 m isobath is encountered.

The computed inventories are mapped in Figure 4a, which shows the influence of the ventilation pathways exuding from with Antarctic Bottom Water formation regions but also highlighting the influence of the deep hydrothermal feature reported by Winckler et al. (2010) around 120°W . The relationship between the computed inventories and neutral density shows a hint of specific “modes” near 28.1 , 28.2 , and 28.3 kg m^{-3} . The flattening of the cumulative inventory curve (Figure 4b) and the drop-off in the inventory density (Figure 4c) above 28.3 kg m^{-3} support our choice of that horizon as a bound for our inventory calculation. We thus estimate the inventory of ^3He in the Southern Ocean to be $4.9 \times 10^4 \text{ mol}$. Adjusting the integration volume

Figure 4. (a) A map of the computed ^3He water column inventories in nmol m^{-2} above the 28.30 kg m^{-3} horizon (or seafloor) and the relationship between (b) cumulative ^3He inventory and (c) ^3He inventory by density bin as a function of neutral density.

between the 28.25 and 28.35 kg m^{-3} horizons changes this number from 4.15 to $5.30 \times 10^4 \text{ mol}$, suggesting an uncertainty of $0.6 \times 10^4 \text{ mol}$.

4. Discussion

The global marine hydrothermal ^3He flux is $548 \pm 52 \text{ mol a}^{-1}$ (using a weighted mean of the estimates by Bianchi et al., 2010; DeVries & Holzer, 2019; Schlitzer, 2016). Assuming that $90 \pm 10\%$ of this flux exits through the Southern Ocean, we obtain a displacement time scale of 99 ± 18 years for this inventory. Given that the Pacific deep water overturning cell (where the bulk of the hydrothermal injection takes place) occurs below the intermediate waters ($\sim 1,000 \text{ m}$; Wijffels et al., 1996, 2001), this is a reasonable assumption. Moreover, it has recently been shown that the modest upwelling of ^3He seen in the North Pacific amounts to only $\sim 2\%$ of the global flux (Jenkins et al., 2020), making the relative uncertainty in the fraction exiting through the Southern Ocean an upper bound.

The time scale obtained here represents a single metric of the spectrum of the transit times for deep water return to the surface and should therefore be interpreted with caution. It represents the approximate transit time from where the 28.00 kg m^{-3} neutral density surface rises above $2,200 \text{ m}$ (see Figure 3b) to the surface. It is important to note that there is a model dependency to this estimate in that the global flux calculations depend on a modeled ventilation time scale in the Southern Ocean. That is, the various estimates of the global hydrothermal ^3He flux (Bianchi et al., 2010; DeVries & Holzer, 2019; Holzer et al., 2017; Schlitzer, 2016) have an assumed Southern Ocean ventilation time scale (based on a variety of different physical models and tracer distributions), so that this estimate is not independent of those assumptions. If, for example, the model-estimated fluxes were to halve, this time scale estimate would double. Conversely, if it were shown that this estimate indicated a much faster Southern Ocean ventilation time scale than originally intrinsic to the models, the model fluxes would correspondingly increase. Indeed, the “mean reexposure time” obtained from one model estimate (DeVries & Holzer, 2019, see their Figure 6a; e.g., near 50° south and 2 km depth) appears considerably greater than this number, suggesting further investigation is merited. Although such a comparison is made difficult by the differing nature of the time scale definitions, such a large discrepancy may signal deeper challenges.

By comparison, using a combination of hydrographic analysis and particle trajectories from high-resolution Southern Ocean model simulations, Tamsitt et al. (2017) estimate an upwelling half-time of $60\text{--}90$ years,

which appears broadly consistent our analysis. A precise comparison would require knowledge of the true transit time distribution for this system, which we currently lack. Recent work has indicated that simple spectral functions such as the inverse Gaussian do not apply well to Southern Ocean upwelling (Trossman et al., 2014). In another study, Drake et al. (2018) use a hierarchy of models exhibiting Southern Ocean Lagrangian upwelling time scales that decline from 87 to 17 years with increasing resolution. The estimate of transit times for the rise of deep waters to the Southern Ocean surface that we obtain here thus appears similar to the hydrographic model presented by Tamsitt et al. (2017) and intermediate between the Drake et al. (2018) estimate of a few decades and that by DeVries and Holzer (2019) of a few centuries.

This calculation was in part motivated by interest in the potential role of abyssal marine hydrothermal iron emissions in fueling new primary production in the Southern Ocean. Since it is known that significant fractions of this iron persists in the dissolved phase (Bennett et al., 2008; Boyle & Jenkins, 2008; Jenkins et al., 2020; Resing et al., 2015; Tagliabue et al., 2010) but given that scavenging processes also remove this iron from the dissolved phase (Bergquist & Boyle, 2006; Fitzsimmons et al., 2015, 2017; Hayes et al., 2015), consideration of time scales becomes important. Dissolved iron scavenging is a complicated biogeochemical process, and the determination of these time scales and contributing factors is currently an extremely active research area (Tagliabue et al., 2016; Tagliabue & Resing, 2016). In general, one would expect short iron residence times in regions of high biological activity, contact with sediments, or active hydrothermal vent environments. There exists a broad range of time scale estimates (Hayes et al., 2015; Table 2 is a useful summary, see also Tagliabue et al., 2016) ranging from a few years to multiple centuries. Although our estimate falls within this range, extrapolating to an estimate of what fraction of, for example, East Pacific Rise derived iron surfaces in the Southern Ocean would be a very long reach. In fact, a recent modeling effort suggests that only a small percentage may make it to the surface of the Southern Ocean (Roshan et al., 2020). Nonetheless, the amounts are still significant and bear consideration.

Data Availability Statement

All of the data used in this publication are openly available from the U.S. National Oceanic and Atmospheric Administration's National Centers for Environmental Information (<https://www.nodc.noaa.gov/ocads/data/0176626.xml>). The data set's DOI is 10.25921/c1sn-9631.

Acknowledgments

The author is grateful to all those dedicated analysts and seagoing support groups that have toiled to create a global data set of marine helium isotope measurements. I also thank Bob Anderson for inspiring this endeavor. This work was funded under the auspices of the U.S. National Science Foundation's Grant OCE-1756138. The author is also appreciative of numerous constructive critical comments by the editor and three anonymous reviewers that improved this manuscript.

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