210Po-210Pb Disequilibrium in the Western North Pacific Ocean: Particle Cycling and POC Export

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Estimating the particulate organic carbon (POC) export flux from the upper ocean is fundamental for understanding the efficiency of the biological carbon pump driven by sinking particles in the oceans. The downward POC flux from the surface ocean based on 210Po-210Pb disequilibria in seawater samples from the western North Pacific Ocean (w-NPO) was measured in the early summer (May-June) of 2018. All the profiles showed a large 210Po deficiency relative to 210Pb in the euphotic zone (0–150 m), while this 210Po deficiency vanished below ~500 m (with 210Po/210Pb ~1 or > 1).

A one-dimensional steady-state irreversible scavenging model was used to quantify the scavenging and removal fluxes of 210Po and 210Pb in the euphotic zone of the w-NPO. In the upper ocean (0–150 m), dissolved 210Po (D-Po) was scavenged into particles with a residence time of 0.6–5.5 year, and the 210Po export flux out of the euphotic zone was estimated as (0.33–3.49) × 10^4 dpm/m^2/year, resulting in a wide range of particulate 210Po (P-Po) residence times (83–921 days). However, in the deep ocean (150–1,000 m), 210Po was transferred from the particulate phase to the dissolved phase. Using an integrated POC inventory and the P-Po residence times (Eppley model) in the w-NPO euphotic zone, the POC export fluxes (mmol C/m^2/d) varied from 0.6 ± 0.2 to 8.8 ± 0.4. In comparison, applying the POC/210Po ratio of all (>0.45 µm) particles to 210Po export flux (Buesseler model), the obtained POC export fluxes (mmol C/m^2/d) ranged from 0.7 ± 0.1 to 8.6 ± 0.8. Both Buesseler and Eppley methods showed enhanced POC export fluxes at stations near the continental shelf (i.e., Luzon Strait and the Oyashio-Kuroshio mixing region). The Eppley model-based 210Po-derived POC fluxes agreed well with the Buesseler model-based fluxes, indicating that both models are suitable for assessing POC fluxes in the w-NPO. The POC export efficiency was < 15%, suggesting a moderate biological carbon pump efficiency in the w-NPO. These low export efficiencies may be associated with the dominance of smaller particles and the processes of degradation and subsequent remineralization of these small particles in the euphotic zone of oligotrophic regions in the w-NPO.

Keywords: 210Po deficiency, POC export flux, marine biological carbon pump, Western North Pacific Ocean, euphotic zone
INTRODUCTION

As the Earth's largest carbon reservoir (IPCC in Climate Change, 2013), the marine ecosystem plays a fundamental role in regulating the atmospheric CO$_2$ concentration and buffering the effects of global climate change (Sabine, 2004) by assimilating carbon from the atmosphere via dissolution and photosynthesis in the upper ocean. Even small changes in the magnitude of the downward transport of carbon in the upper ocean can have serious impacts on oceanic carbon sequestration and atmospheric CO$_2$ concentration (Kwon et al., 2009) and thus on global climate change (including global warming). It has been predicted that without the marine biological carbon pump (BCP), the atmospheric CO$_2$ concentration could be approximately 50% higher than its current value (Sanders et al., 2014).

To fully understand how the oceans regulate the atmospheric CO$_2$ concentration, detailed knowledge of the marine carbon cycle is needed. The well-known BCP is one of three major mechanisms by which the ocean takes up atmospheric CO$_2$ (Liu et al., 2018). The BCP is described by these major processes: Phytoplankton convert CO$_2$ into fixed carbon, e.g., carbohydrates or calcium carbonate through photosynthesis in the euphotic zone; Part of the CO$_2$ fixed is transferred to the ocean interior of the ocean, mainly by gravitational sinking of particulate organic carbon (POC); Diffusion, advection and vertical mixing of dissolved organic carbon (DOC) and active bio-transport of organic and inorganic carbon by diel vertical migrating zooplankton are also important carbon removal pathways; Finally very few percentage of fixed carbon will be sequestered in the deep ocean (Buesseler et al., 2007). All these biologically mediated processes constitute the BCP. Thus, sinking particles play an important role in driving the BCP (Falkowski et al., 1998; Ducklow et al., 2001; Wei et al., 2011; Liu et al., 2018).

The ratio of export production (or the POC export flux) to total primary production, known as the “f-ratio” is defined as the export efficiency of the BCP and is used to quantify the strength of the pump (Buesseler and Boyd, 2009). Hence, an investigation of the sinking POC export flux from the upper ocean would provide fundamental parameters for predicting future changes in the marine carbon cycle. To constrain the POC export flux from the upper ocean, two traditional approaches are frequently applied: sediment traps (Honjo et al., 2008; Hayes et al., 2018) and the disequilibrium of the natural radionuclide pair, $^{234}$Th-$^{238}$U (Charette et al., 1999; Benitez-Nelson et al., 2001; Buesseler et al., 2009, 2020; Zhou et al., 2020).

In recent years, another pair of natural radionuclide tracers, $^{210}$Po ($T_{1/2} = 138.4$ days) and its progenitor $^{210}$Pb ($T_{1/2} = 22.4$ year), has been shown to be effective at predicting the POC export flux (Murray et al., 2005; Stewart et al., 2007, 2010; Verdeny et al., 2009; Wei et al., 2011; Le Moigne et al., 2013; Roca-Martí et al., 2016; Hayes et al., 2018; Subha Anand et al., 2018; Tang et al., 2019; Horowitz et al., 2020); however, this tracer pair has not been applied frequently. The $^{210}$Po/$^{210}$Pb ratio exhibits different behavior from the $^{234}$Th/$^{238}$U ratio in terms of particle-binding properties in the ocean. $^{234}$Th is particle reactive and is scavenged from surface waters when particles sink into the deep sea, while naturally occurring $^{238}$U is conserved and remains dissolved in well-oxygenated seawater. In contrast, both $^{210}$Po and $^{210}$Pb are particle reactive. In addition, $^{210}$Pb and $^{234}$Th are only adsorbed onto the particle surface, whereas $^{210}$Po is both adsorbed onto surfaces and biologically assimilated into cells by some species of phytoplankton and bacteria (Fisher et al., 1983; Cherrier et al., 1995; LaRock et al., 1996; Stewart and Fisher, 2003a,b). As a result, phytoplankton cells accumulating $^{210}$Po are further ingested by zooplankton, while sinking particles excreted as fecal pellets and biogenic detritus are depleted in $^{210}$Po (HeyrAUD et al., 1976). This biogeochemical behavior leads to the distribution coefficient of $^{210}$Po being higher than that of $^{210}$Pb (Tang et al., 2017; Zhong et al., 2019). Thus, when particles sink from the upper ocean, the large difference in particle affinity leads to secular disequilibrium between $^{210}$Po and $^{210}$Pb, which can be used to estimate particular material export in a manner similar to the use of $^{234}$Th-$^{238}$U disequilibrium (Friedrich and Rutgers van der Loeff, 2002; Cochrán and Masqué, 2003; Verdeny et al., 2009; Wei et al., 2011; Hayes et al., 2018; Subha Anand et al., 2018; Horowitz et al., 2020). However, relatively few studies have used the $^{210}$Po-$^{210}$Pb disequilibrium method to quantify the POC export flux at a basin wide scale (Ceballos-Romero et al., 2016; Roca-Martí et al., 2016; Tang and Stewart, 2019; Tang et al., 2019).

Oligotrophic open waters compose a major proportion (~75%) of the surface ocean and account for over 30% of global marine carbon fixation (Shih et al., 2015). The NPO is an important atmospheric carbon sink, it can contribute to ~25% of the total ocean CO$_2$ uptake per year (Takahashi et al., 2009). By assessing the carbon export in the NPO, the goal of peak carbon dioxide emissions and carbon neutrality for humanity can be achieved more scientifically, and the pace of global climate change can be gradually showed down. As a part of the "Marine Environment Monitoring and Early Warning System Construction in the West Pacific Ocean (MEMEWSC)" project, we took advantage of a unique opportunity to obtain measurements of $^{210}$Po and $^{210}$Pb activity over a large tract of the w-NPO (from 117°E to 146°E and from 20°N to 40°N). In this study, we describe the activity profiles (1,000 m) of $^{210}$Po and $^{210}$Pb at 7 stations spanning different oceanographic regions with different aeolian inputs, distances from the nearest coast and rates of primary production. The goals of this study are listed as follows: (1) To investigate the spatial variability of particle scavenging and removal processes in the water column of each oceanographic region; and (2) to estimate the magnitude and efficiency of the POC export flux derived from the $^{210}$Po-$^{210}$Pb disequilibrium in the euphotic layer of the w-NPO.

MATERIALS AND METHODS

Sampling and Preparation

The study area located in the western part of the NPO, covering an area of 20–40°N and 118–153°E. The main surface ocean currents in this region are shown in Figure 1. The Kuroshio Current (KC) is the most important western boundary current in the w-NPO, it originates from the North Equatorial Current and subsequently intrudes from the western Philippine Sea into the northern South China Sea (SCS) in the Luzon Strait. The
The main KC enters the East China Sea (ECS) through a channel in east of Taiwan Island and then passes through Tokara Strait at around 30°N. After the current has passed the Shikoku Basin, the northeastward KC merges with the southwestward subarctic Oyashio Current (OC) before leaving the coast of Japan to form the KC Extension in the NPO, resulting in a system known as the Oyashio-Kuroshio boundary region (Figure 1).

The MEMEWSC cruise was carried out in May-June 2018 on board RV Xiangyanghong III from the SCS to the Oyashio-Kuroshio mixing region of the western subarctic NPO (Figure 1). Dissolved and particulate samples for \(^{210}\)Po and \(^{210}\)Pb activity analysis were collected from the water column at 7 stations during the \(~1\) month duration of the cruise. Discrete seawater samples were collected using a stainless steel sampling rosette equipped with 24 \({\times}\) 12 L Niskin bottles and a CTD Seabird sensor package. Seawater samples (~20 L each) were taken from 9 depths from 0 to 1,000 m, with higher resolution in the upper 200 m of the water column. Salinity and temperature were obtained from these sensors (Seabird SBE 9/11plus CTD deployed with two Sea-Bird SBE 3P temperature sensors and two Sea-Bird SBE 4C conductivity sensors). To evaluate the POC export flux, 150 m is chosen as the integration depth according to estimated depth for the base of the euphotic zone on the ship from the Chl-a profiles.

**\(^{210}\)Po and \(^{210}\)Pb Determination**

Analysis for \(^{210}\)Pb and \(^{210}\)Po was described in Zhong et al. (2019). In brief, seawater samples (18–24 L each) were rapidly filtered through 142-mm diameter Nucleopore filters (0.45 \(\mu\)m pore size) to separate the particulate and dissolved \(^{210}\)Po and \(^{210}\)Pb. The filtered (dissolved) seawater was then acidified to pH 1–2 with concentrated HCl immediately after collection and spiked with a known amount of \(^{209}\)Po (No. 7299, Eckert and Ziegler Isotope Products) and stable Pb\(^{2+}\) to quantify any subsequent losses of Po and Pb. After equilibration for 6–12 h, 100 mg of Fe\(^{3+}\) was added to the solution, the pH was adjusted to approximately 8–9, using concentrated NH\(_4\)OH to co-precipitate Po and Pb with Fe(OH)\(_3\) precipitate. After settling for 8–12 h, the precipitate was transferred into a 1.5-L polyethylene bottle, and stored on board for processing upon arrival on land. In the laboratory, the precipitate was centrifuged and dissolved in 6 M HCl solution in a clean Teflon beaker, and the pH was neutralized to 1–2. Both \(^{210}\)Po and \(^{209}\)Po were auto-plated onto a nickel disc (Zhong et al., 2020) after adding 0.3 g of ascorbic acid, 1 mL of 25% sodium citrate, and 1 mL of 20% hydroxylamine hydrochloride to the solution. The particulate filters were spiked with known quantities of \(^{209}\)Po and stable Pb\(^{2+}\) and totally dissolved in a mixture of HF, HNO\(_3\) and HClO\(_4\). After evaporating to nearly dryness several times to ensure that all the radionuclides were in dissolved form, the residue was then picked up in dilute HCl (0.1 M) for plating of Po onto a nickel disc. The plating procedure was similar to that utilized for dissolved samples. \(^{210}\)Po and \(^{209}\)Po activities were determined by alpha spectrometry (Canberra series 7200-08).

For \(^{210}\)Pb determination, any remaining Po isotopes in the plating solution was removed by resuspending another nickel disc, and the purified sample solution was re-spiked with
additional $^{209}$Po and stored for ~12 months to allow ingrowth of
$^{210}$Po from $^{210}$Pb, and then polonium isotopes were again auto-
deposited, and counted. The recoveries of $^{210}$Pb were determined
through the added stable Pb and the measured Pb, using atomic
absorption spectrometry (AAS). The in-situ activities of $^{210}$Po and $^{210}$Pb at the sampling date were determined by correcting
decay, ingrowth, chemical recoveries, detector backgrounds,
and reagent blanks ($^{210}$Pb and $^{210}$Po present in the stable lead)
(Church et al., 2012; Baskaran et al., 2013; Rigaud et al., 2013).

**Particulate Organic Carbon Analysis**

POC samples were obtained by filtering 4–6 L seawater through
pre-combusted and pre-weighted 25 mm QMA filters (with a
pore size of 0.45 µm, Whatman) immediately after sampling,
and the filters were then washed 3 times with Milli-Q water to
remove salt before frozen storage prior to laboratory analysis.
Before the POC content was measured, the sample was oven-
dried at 55°C to a constant weight. Then, the inorganic carbon
was removed from the filter by using acid vapor (concentrated
HCl) in desiccators for 48 h. The POC was then determined
using a Vario ELIII CHNOS Elemental Analyzer. The analytical
precision of the method was estimated to be around ± 4%, using
triplicate measurement of the same sample.

**RESULTS**

**Hydrographic Characteristics and**
**Particulate Organic Carbon**
**Concentrations during the MEMEWSC**
**Cruise**

Vertical profiles of the temperature and salinity in the w-NPO
are shown in Figure 2. Among the 7 stations, the sea surface
temperature (SST) ranged from 15.5 to 28.9°C. The SST was
high (27.6–28.9°C) at the low-latitude (21°N) stations (XTW1-3,
XTW1-11 and XTJ1-1) but decreased gradually with increasing
latitude from station XTJ1-1 (28.4°C) to station XTJ1-22 (15.5°C).
Influenced by the OC, station XTJ1-22 showed the lowest SST.
Furthermore, the water temperature decreased with increasing
depth and did not differ appreciably between stations at 1,000 m
(Figure 2A).

The salinity profiles spatially differed among the 7 stations
(Figure 2A). At station XTJ1-22, the salinity decreased rapidly
with increasing depth and exhibited a subsurface minimum
(33.82) at 150 m, displaying the features of low-salinity subarctic
waters (OC), and then the salinity increased gradually to 1,000 m.
In contrast to station XTJ1-22, the other six stations exhibited
similar salinity distributions, with the maximum values occurring
at depths of approximately 150–200 m (Figure 2A), indicating
the existence of stratification in the water column at each station.
However, the salinity in the top 150 m of the water
column at station XTW1-3 (located in Luzon Strait) was lower
(from 33.80 to 34.66) than that at the station in the NPO
basin (from 34.55 to 35.05) (Figure 2A). From the temperature-
salinity-depth diagram (Figure 2B), the water masses can be
clearly discriminated for the study area. Station XTJ1-22 showed
a signature of OC, and station XTW1-3 might represent the
South China Sea water. From the Figure 2B, the remaining five
stations (XTJ1-1, XTW1-1, XTJ1-8, XTJ2-7, and XTJ1-13) could
be divided into two groups with different hydrographic features
for the surficial water mass (upper 200 m). One group includes
stations XTJ1-1 and XTW1-11, and the other group contains
stations XTJ1-8, XTJ1-13, and XTJ2-7, showing a much narrower
ranges of salinity and temperature (Figure 2B). In 1,000 m, all
the stations displayed a similar temperature-salinity feature, with
low temperatures of < 5°C and a salinity range of 34.3–34.5.
Based on the salinity and temperature profiles, the thermocline
was observed between 100 and 200 m. To determine the euphotic
zone depth, we use information for the base of the euphotic zone
according to the subsurface chlorophyll-a maximums (SCMs)
features. The profiles of chlorophyll-a concentrations showed
that SCMs could occur between 120 and 150 m for most of the
stations (33 out of 43 stations) during the cruise (unpublished
data, personal communication with Dr. Jianhua Kang). Hence the
euphotic zone could be delimited from 0 to 150 m. Overall, the
water column could be divided into two boxes: 0–150 m box (the
euphotic zone) and 150–1,000 m box (the mesopelagic zone).

All POC concentration data can be found in Supplementary
Table 1. In total, the POC concentrations ranged from 1.85 to
14.35 µmol C/L, with an average of 4.15 ± 2.15 µmol C/L in
the w-NPO (Figure 3). Station XTJ1-22 at the Kuroshio Current-
Oyashio Current boundary had the highest POC concentration
(mean: 6.23 ± 4.12 µmol C/L) in the upper 150 m, followed
by station XTW1-3 (mean: 5.44 ± 1.28 µmol C/L) in the
Luzon Strait, while the other five stations had similar POC
concentrations with a mean value of 3.12–3.81 µmol C/L (Table 1).
Almost all POC profiles showed a decrease with increasing
water column depth from 0 to 500 m, and remained
relatively constant from 500 to 1,000 m except at two stations
(XTJ1-8 and XTJ1-13) (Figure 3).

$^{210}$Po and $^{210}$Pb Profiles

Profiles of dissolved $^{210}$Po (D-Po), particulate $^{210}$Po (P-Po),
total $^{210}$Po (T-Po), dissolved $^{210}$Pb (D-Pb), particulate $^{210}$Pb
(P-Pb) and total $^{210}$Pb (T-Pb) were displayed in Figures 4, 5,
respectively. From Figures 4, 5, overall, P-Po and P-Pb showed
a similar variation trend, but D-Po showed a little bit of
difference comparing with the D-Pb. The activity concentrations
of particulate $^{210}$Pb (P-Pb) ranged from 1.4 to 10.4 dpm/100
L (Figure 4), which are comparable to the reported values in
the upper Sargasso Sea (> 500 m, 0.5–8.2 dpm/100 L) (Kim and
Church, 2001), and slightly higher than those in the Aleutian
Basin (negligible to 6.0 dpm/100 L) (Hu et al., 2014) and in
the North Atlantic Ocean (0.1–7.0 dpm/100 L) (Horowitz et al.,
2020). All stations showed a systematic decrease in P-Po activity
with depth in the upper 1,000 m, except for station XTW1-3 in
the Luzon Strait, which exhibited a subsurface maximum between
200 and 500 m (Figure 4).

The activities of dissolved $^{210}$Po (D-Po) varied from 2.87
to 14.57 dpm/100 L (Figure 4), which are comparable to
other results previously measured in the w-NPO (5.28–14.02
dpm/100 L) (Nozaki and Tsunogai, 1976), the North Atlantic
Ocean (0.6–13.1 dpm/100 L) (Horowitz et al., 2020) and the
western equatorial Pacific Ocean (4.26–14.82 dpm/100 L) (Peck and Smith, 2000), although these regions might correspond to very different biogeochemical conditions. The D-Po concentration was lowest in surface seawater and gradually increased with depth; the subsurface D-Po peak was observed within approximately 50–150 m at all stations (Figure 4). This increase in the D-Po concentration with depth is consistent with the rapid scavenging of $^{210}$Po near the surface and the remineralization of particulate $^{210}$Po at depth. The profiles of total $^{210}$Po (T-Po) showed vertical variations similar to those of P-Po, with a range of 4.61–20.69 dpm/100 L.

The activity concentrations of dissolved $^{210}$Pb (D-Pb) and particulate $^{210}$Pb (P-Pb) ranged from 1.86 to 24.45 dpm/100 L and from 1.88 to 31.33 dpm/100 L, respectively (Figure 5). The P-Pb activity was highest in the near-surface seawaters at all stations; thus, $^{210}$Pb activity was predominantly in the particulate phase (P-Pb/T-Pb > 50%) in the upper 50 m (Figure 6). Below 50 m, P-Pb showed a systematic downward trend with depth.
FIGURE 3 | Vertical profiles of the POC concentration at 7 stations in the w-NPO during May-June 2018.

TABLE 1 | Model calculations of the scavenging rates (J) and removal fluxes (P) for $^{210}$Po and $^{210}$Pb, together with the particulate and dissolved residence times in the surface and deeper layers in the w-NPO.

| Station | Depth layer | Average POC concentration* | Inventory of $^{210}$Po deficit | $J_{Po}$ | $P_{Po}$ | $\tau_{Po-D}$ | $\tau_{Po-P}$ | $J_{Pb}$ | $P_{Pb}$ | $\tau_{Pb-D}$ | $\tau_{Pb-P}$ |
|---------|-------------|-----------------------------|-------------------------------|--------|--------|---------------|---------------|--------|--------|---------------|---------------|
|         | m           | $\mu$mol C/L                | dpm/m$^2$                     | dpm/m$^2$/yr | yr | d | dpm/m$^2$/yr | yr |
| XTW1-3  | 0–150       | 5.44 ± 1.28                 | 8,733 ± 694                  | 1,693  | 17,242 | 5.5 ± 1.3 | 130 ± 7      | 18,697 | 18,244 | 0.51 ± 0.02 | 0.80 ± 0.03  |
| XTW1-11 | 0–150       | 3.66 ± 0.64                 | 6,144 ± 1,058                | 3,464  | 12,509 | 4.2 ± 0.9  | 308 ± 28     | 18,504 | 18,024 | 0.85 ± 0.03 | 0.86 ± 0.04  |
| XTJ1-1  | 0–150       | 3.81 ± 1.29                 | 1,121 ± 821                  | 7,683  | 3,327  | 1.6 ± 0.2  | 921 ± 229    | 18,504 | 18,317 | 0.85 ± 0.03 | 0.33 ± 0.02  |
| XTJ1-8  | 0–150       | 3.12 ± 0.83                 | 4,344 ± 857                  | 4,214  | 9,219  | 3.2 ± 0.5  | 275 ± 27     | 18,522 | 18,222 | 0.82 ± 0.03 | 0.53 ± 0.02  |
| XTJ1-13 | 0–150       | 3.36 ± 0.58                 | 18,415 ± 1820                | 17,644 | 34,941 | 0.6 ± 0.0  | 83 ± 5       | 18,373 | 17,835 | 1.09 ± 0.05 | 0.98 ± 0.09  |
| XTJ1-22 | 0–150       | 6.23 ± 4.12                 | 7,389 ± 457                  | 5,921  | 14,785 | 1.1 ± 0.1  | 93 ± 4       | 18,720 | 18,454 | 0.47 ± 0.02 | 0.47 ± 0.01  |
| XTJ2-7  | 0–150       | 3.62 ± 1.24                 | 13,507 ± 874                 | 9,400  | 25,969 | 1.2 ± 0.1  | 84 ± 4       | 18,517 | 18,053 | 0.83 ± 0.04 | 0.83 ± 0.02  |
| XTW1-3  | 150–1,000   | 5.59 ± 2.65                 | 2,5310 ± 7261               | –6,023 | 63,509 | /            | 331 ± 40     | 3,132  | 15,575 | 22.9 ± 5.7  | 5.5 ± 0.3    |
| XTW1-11 | 150–1,000   | 3.08 ± 0.27                 | –3,910 ± 6,341               | –86,511| 5,362  | /            | /            | 3,800  | 15,713 | 13.1 ± 2.6  | 4.8 ± 0.2    |
| XTJ1-1  | 150–1,000   | 2.42 ± 0.41                 | –34,615 ± 6270              | –64,036| –59,950| /            | /            | 3,527  | 17,608 | 16.0 ± 3.8  | 1.3 ± 0.1    |
| XTJ1-8  | 150–1,000   | 2.65 ± 0.74                 | –16,670 ± 5421              | –26,022| –21,254| /            | /            | 3,297  | 17,649 | 20.3 ± 4.8  | 1.1 ± 0.1    |
| XTJ1-13 | 150–1,000   | 3.63 ± 0.75                 | 33,430 ± 4475               | 7,294  | 96,052 | 8.1 ± 4.1  | 85 ± 6       | 3,308  | 16,231 | 18.7 ± 3.2  | 3.2 ± 0.1    |
| XTJ1-22 | 150–1,000   | 3.63 ± 1.64                 | 14,370 ± 4187               | –7,166 | 41,054 | /            | 155 ± 18     | 3,753  | 17,350 | 13.9 ± 2.2  | 2.1 ± 0.1    |
| XTJ2-7  | 150–1,000   | 2.82 ± 0.03                 | 54,550 ± 5731               | 32,338 | 125,888| /            | 103 ± 7      | 3,103  | 15,810 | 22.8 ± 4.8  | 4.6 ± 0.2    |

/*" denotes an "invalid calculation."

*Represents the average POC concentration at 0–150 m box and 150–1,000 m box.
FIGURE 4 | Vertical distributions of particulate (P), dissolved (D), and total (T) $^{210}$Po in the water column at each station in the w-NPO.
FIGURE 5 | Vertical distributions of particulate (P), dissolved (D), and total (T) $^{210}$Pb activities in the water column at each station in the w-NPO.
However, the activity of D-Pb was lowest in the surface seawater; then, D-Pb increased with depth from 50 to 200 m and finally decreased below 200 m. All stations had the highest total $^{210}\text{Pb}$ (T-Pb) concentration at the surface, and T-Pb decreased with depth from the surface to 1,000 m, suggesting a surface source of $^{210}\text{Pb}$ due to atmospheric input, as indicated by the $^{210}\text{Pb}$ activity (7.22–38.61 dpm/100 L) being greater than the $^{226}\text{Ra}$ activity (6.0–14.4 dpm/100 L, Kawakami and Kusakabe, 2008) in the upper water column.

At all stations, the percentages of P-Po in the w-NPO surface seawater (0–50 m) were higher than 50%, showing that P-Po often exceeded D-Po and dominated T-Po. The high fraction of particulate $^{210}\text{Pb}$ indicated a signature of atmospheric input. Moreover, the P-Po/T-Po percentages decreased slowly to constant values of 20–30% as the water depth increased up to 500 m (Figure 6). P-Pb was generally 20–50% of the T-Pb at depths below the mixed layer, demonstrating that T-Pb was dominated by D-Pb in most layers of the water column at all stations in the w-NPO (Figure 6). Overall, these findings reveal that both $^{210}\text{Po}$ and $^{210}\text{Pb}$ were predominantly present in the dissolved phase below 50 m at all stations (Figure 6), as is commonly the case in other oceans.

**$^{210}\text{Po}/^{210}\text{Pb}$ Activity Ratio**

Vertical profiles of the $^{210}\text{Po}/^{210}\text{Pb}$ activity ratios in the particulate, dissolved and total fractions are plotted in Figure 7. Particulate matter was enriched in $^{210}\text{Po}$ (particulate $^{210}\text{Po}/^{210}\text{Pb}$ activity ratios > 1) at the subsurface (50–100 m) at all stations, especially in the particulate sample from a depth of 75 m at station XTJ1-1, where the P-Po/P-Pb ratio reached 3.04 ± 0.73 (Figure 7). These data indicates that there was an excess of P-Po relative to P-Pb. The dissolved $^{210}\text{Po}/^{210}\text{Pb}$ activity ratios (D-Po/D-Pb) increased with depth, with values < 1 at the surface and increasing toward 1 or even > 1 below the euphotic zone, which complemented the elevated P-Po/P-Pb ratio. All stations showed similar vertical distributions of T-Po/T-Pb with a significant deficiency of $^{210}\text{Po}$ relative to $^{210}\text{Pb}$ from the surface to 200 m. The deficit of $^{210}\text{Po}$ relative to $^{210}\text{Pb}$ declined with depth and then approached equilibrium or shifted to an excess ($^{210}\text{Po}/^{210}\text{Pb}$ > 1) at different depths for different stations, except for stations XTJ1-1 and XTJ1-8, which displayed an excess of T-Po relative to T-Pb (T-Po/T-Pb ratio > 1) at 50–75 m and at 500–1,000 m. Interestingly, $^{210}\text{Po}$ excess (T-Po/T-Pb ratios > 1.2) was observed in seawater at 500 and 1,000 m of station XTJ1-1 and in seawater at 1,000 m of station XTJ1-8 (Figure 7). From Figure 7, we can find that this $^{210}\text{Po}$ excess is caused by the high dissolved $^{210}\text{Po}$ activity concentration in the deep layers (500–1,000 m), because D-Po/D-Pb ratios at deep layers (500–1,000 m) were much higher than unity (Figure 7). And this high dissolved $^{210}\text{Po}$ activity concentration implied the occurrence of a strong $^{210}\text{Po}$ dissolution in the deep ocean.

**DISCUSSION**

**Enhanced Particle Scavenging of $^{210}\text{Po}$ and $^{210}\text{Pb}$ and Deficiency of $^{210}\text{Po}$ in the Western North Pacific Ocean**

Both Pb and Po are particle-reactive elements that can be scavenged effectively from the water column by lithogenic and biogenic particles. Figure 4 reveals that both the P-Po and the
P-Pb activity concentrations were highest in the near-surface (0–30 m) layers at all stations; additionally, at most stations, the highest POC concentrations were similarly detected in the upper water column (>100 m), as shown in Figure 3, which resulted in higher percentages of P-Po/T-Po and P-Pb/T-Pb ratios (>50%) in the near-surface (30–50 m) seawater. The high fraction (30–60%) of particulate $^{210}$Pb for most stations was related to atmospheric deposition of $^{210}$Pb. In the open ocean, the most important input of $^{210}$Pb is the atmospheric deposition, and the $^{210}$Pb is mainly in dissolved phase in the atmospheric deposition samples. $^{210}$Pb is particle-reactive, therefore, in the upper layer with higher primary productivity, some $^{210}$Pb sourced from the atmosphere would be absorbed and enriched by the biogenic particles. In addition, our study area is under the cover of dust deposition from the East Asia (Hayes et al., 2013); during every spring (especially in April-May), the prevailing dust storms from the East Asia would possibly increase the particulate $^{210}$Pb activity.

Furthermore, the P-Po/P-Pb ratios were nearly 1 or higher than 1 within 30–75 m at all stations (Figure 7), supporting the previously findings that $^{210}$Po was preferentially scavenged over $^{210}$Pb (Bacon et al., 1976; Murray et al., 2005; Teng and Stewart, 2019). In addition, on all profiles, both P-Po and P-Pb decreased with depth in the upper 100 m, which is consistent with the increases in D-Po and D-Pb (Figures 4, 5). The highest P-Po activity concentrations were presented at the surface for all stations (Figure 4), which could be related to particle concentration, since particle concentration is usually higher than in the twilight zone or deeper layers. D-Po increased from the surface to 1,000 m (Figure 4), especially at stations XTW1-11 and XTJ1-8, which showed subsurface D-Po maxima at 150 and 50 m, respectively. This phenomenon implies that sinking particulate matter is re-mineralized at depth in the w-NPO, which is supported by an excess of D-Po relative to $^{210}$Pb, with the D-Po/D-Pb activity ratios often > 1 below 200 m (Figure 7). Specifically, at station XTW1-11, the feature of D-Po/D-Pb > 1 appeared between 150 and 1,000 m, while at station XTJ1-1, D-Po/D-Pb > 1 only appeared between 500 and 1,000 m. Except for station XTJ2-7, the feature of D-Po/D-Pb > 1 appeared at 1,000 m for all stations (Figure 7). Based on the above discussion, it can be concluded that enhanced particle scavenging of both $^{210}$Po and $^{210}$Pb occurred in the upper ocean and that $^{210}$Po adsorbed onto particles from the upper ocean was released in the deeper ocean due to the remineralization of particulate organic matter below the mixed layer.

The minimum activity concentration of $^{210}$Po appeared at the 100 m layer at XTW1-3, XTJ1-8 and XTJ2-7, while the minimum $^{210}$Po activity concentration appeared at the 200 m layer at station XTW1-11, XTJ1-1, XTJ1-13, and XTJ1-22 (Figure 4). As shown in Figure 7, T-Po was largely deficient relative to T-Pb in the upper 200 m of the w-NPO, with the T-Po/T-Pb ratios as low as 0.4. This phenomenon is similar to the general pattern in which T-Po is usually deficient with respect to T-Pb (T-Po/T-Pb ratio = 0.5) in the euphotic zone due to the rapid removal of Po caused by the sinking of biogenic particles (Bacon et al., 1976; Nozaki et al., 1998; Stewart et al., 2010; Roca-Martí et al., 2016; Subha Anand et al., 2018). In contrast, $^{210}$Po gradually reached secular equilibrium with $^{210}$Pb or even an excess relative to $^{210}$Pb (Figure 7) at the bottom of the euphotic zone (~150 m) or in the mesopelagic waters (500 or 1,000 m) due to the regeneration of $^{210}$Po from sinking particles during particle remineralization. For example, at stations XTJ1-1 and XTJ1-8, T-Po/T-Pb ratios were high up to 1.61–2.02 at 500–1,000 m layers. From Figures 4, 5, we can clearly find that the D-Po activity concentration increased significantly, but at the same layers, both D-Pb and P-Pb activity concentration decreased significantly. The explanation for this increased D-Po activity at the deep layer may be related to strong particle export events. We believe that a strong particle export event occurred in the upper ocean before our sampling, and subsequently sinking particles dissolved at the deep layers. This ultimately increased the activity concentration of dissolved $^{210}$Po in the mesopelagic zone. Similar research cases have also been reported in a review paper published by Verdeny et al. (2009). Such disequilibrium between $^{210}$Po and $^{210}$Pb in the water column could help depict the scavenging rates of $^{210}$Po and
Scavenging and Removal Fluxes of $^{210}$Po and $^{210}$Pb in the Western North Pacific Ocean

$^{210}$Pb and $^{210}$Po in the upper ocean come from both atmospheric deposition and in situ production via parent radionuclide decay. Generally, residence times of $^{210}$Po and $^{210}$Pb in the atmosphere are only several days to several weeks. Hence, $^{210}$Po deposition fluxes to the surface ocean are only approximately 10–20% of those of $^{210}$Pb (Masqué et al., 2002; Baskaran, 2011). Many previous researchers have applied the $^{210}$Po/$^{210}$Pb disequilibrium method to trace the POC export fluxes in the open ocean, polar sea or remote sea areas, but the $^{210}$Po and $^{210}$Pb deposition fluxes have consistently been neglected due to a lack of relevant data (Friedrich and Rutgers van der Loeff, 2002; Murray et al., 2005; Le Moigne et al., 2013; Roca-Martí et al., 2016; Subba Anand et al., 2018; Tang et al., 2019; Horowitz et al., 2020). According to a compilation of global $^{210}$Pb fallout data by Du (2019), among all the observation stations worldwide, Taiwan, Shanghai and Tatsunokuchi recorded the highest annual $^{210}$Pb deposition fluxes, indicating that East Asia exhibits the highest $^{210}$Pb deposition flux in the world. Thus, when applying $^{210}$Po-$^{210}$Pb disequilibrium to study particle dynamics and to estimate the POC export fluxes in the seas of East Asia, the atmospheric inputs of $^{210}$Po and $^{210}$Pb and their impacts on the POC export flux should be considered.

The $^{210}$Po and $^{210}$Pb activities in the ocean are the result of a balance among atmospheric inputs, continuous production from the decay of mother nuclides ($^{210}$Po or $^{226}$Ra) in seawater, the radioactive decay of $^{210}$Po and $^{210}$Pb, removal onto sinking particles, and transport into or out of the system by advection and diffusion. Generally, the general form of the mass balance equation in the upper ocean for $^{210}$Po and $^{210}$Pb between sources and sinks can be described as follows:

$$\frac{\partial \text{Pb}}{\partial t} = F_{\text{Pb}} + \lambda_{\text{Pb}} I_{\text{Ra}} - \lambda_{\text{Pb}} I_{\text{Pb}} - k_{\text{Pb}} I_{\text{Pb}} + V,$$

$$\frac{\partial \text{Po}}{\partial t} = F_{\text{Po}} + \lambda_{\text{Po}} I_{\text{Pb}} - \lambda_{\text{Po}} I_{\text{Po}} - k_{\text{Po}} I_{\text{Po}} + V,$$

where $\partial \text{Pb}/\partial t$ and $\partial \text{Po}/\partial t$ are the changes in the $^{210}$Pb and $^{210}$Po activities with time, respectively; $F_{\text{Pb}}$ and $F_{\text{Po}}$ (dpm/m²/d) are the atmospheric deposition fluxes of $^{210}$Pb and $^{210}$Po to the sea surface, respectively; $\lambda_{\text{Pb}}$ and $\lambda_{\text{Po}}$ are the decay constants of $^{210}$Pb and $^{210}$Po (d⁻¹), respectively; $I_{\text{Ra}}$, $I_{\text{Pb}}$ and $I_{\text{Po}}$ (dpm/m²/year) are the inventories of $^{228}$Ra, $^{210}$Pb, and $^{210}$Po, respectively; $k_{\text{Pb}}$ and $k_{\text{Po}}$ are the scavenging (from dissolved to particulate) rate constants of $^{210}$Pb and $^{210}$Po, respectively; and $V$ (dpm/m²/d) is the sum of the advection and diffusion fluxes.

The advection-diffusion term is relatively important only in the case of algal blooms, mesoscale eddies, and upwelling regions (Tang et al., 2019; Horowitz et al., 2020). Generally, a steady-state model can be adopted by ignoring advection and diffusion ($V \approx 0$). Figure 8 displays the one-dimensional irreversible scavenging conceptual model for $^{210}$Po and $^{210}$Pb in surface and deep waters. Thus, Eqs. (1) and (2) can be rearranged as follows. For the upper ocean (0–150 m), we obtain:

$$\lambda_{\text{Pb}} I_{\text{Ra}} + F_{\text{Pb}} = \lambda_{\text{Pb}} I_{\text{Pb}} - D + k_{\text{Pb}} I_{\text{Pb}} + V \text{(}= I_{\text{Pb}}(0–150 \text{ m})),$$

$$F_{\text{Pb}}(0–150 \text{ m}) = \lambda_{\text{Pb}} I_{\text{Pb}} - D + P_{\text{Pb}}(0–150 \text{ m}),$$

$$\lambda_{\text{Po}} I_{\text{Pb}} - D + F_{\text{Po}} = \lambda_{\text{Po}} I_{\text{Po}} - D + k_{\text{Po}} I_{\text{Po}} - D \text{(}= I_{\text{Po}}(0–150 \text{ m})),$$

$$F_{\text{Po}}(0–150 \text{ m}) = \lambda_{\text{Po}} I_{\text{Po}} - D + P_{\text{Po}}(0–150 \text{ m}),$$

where Eqs. (3) and (4) are mass-balance equations for D-Pb and P-Pb, and Eqs (5) and (6) are mass-balance equations for D-Po and P-Po.

The residence times of $^{210}$Po and $^{210}$Pb in the upper ocean can therefore be written as:

$$\tau_{\text{Po}}(0–150 \text{ m}) = I_{\text{Po}}(0–150 \text{ m}) / I_{\text{Po}}(0–150 \text{ m}),$$

$$\tau_{\text{Po}}(0–150 \text{ m}) = I_{\text{Po}}(0–150 \text{ m}) / P_{\text{Po}}(0–150 \text{ m}),$$

$$\tau_{\text{Po}}(0–150 \text{ m}) = I_{\text{Po}}(0–150 \text{ m}) / P_{\text{Po}}(0–150 \text{ m}).$$

Similarly, for the deeper ocean (150–1,000 m), we obtain:

$$\lambda_{\text{Po}} I_{\text{Ra}} = \lambda_{\text{Po}} I_{\text{Pb}} - D + k_{\text{Po}} I_{\text{Pb}} + V \text{(}= I_{\text{Pb}}(150–1000 \text{ m})),$$

$$P_{\text{Po}}(0–150 \text{ m}) = \lambda_{\text{Po}} I_{\text{Pb}} - P_{\text{Po}}(0–150 \text{ m}),$$

$$\lambda_{\text{Po}} I_{\text{Pb}} = \lambda_{\text{Po}} I_{\text{Po}} - D + k_{\text{Po}} I_{\text{Po}} - D \text{(}= I_{\text{Po}}(150–1000 \text{ m})),$$

$$P_{\text{Po}}(0–150 \text{ m}) = \lambda_{\text{Po}} I_{\text{Pb}} - P_{\text{Po}}(150–1000 \text{ m}).$$

The residence times of $^{210}$Po and $^{210}$Pb in the deeper ocean can therefore be written as:

$$\tau_{\text{Po}}(150–1000 \text{ m}) = I_{\text{Po}}(150–1000 \text{ m}) / I_{\text{Po}}(150–1000 \text{ m}),$$

$$\tau_{\text{Po}}(150–1000 \text{ m}) = I_{\text{Po}}(150–1000 \text{ m}) / P_{\text{Po}}(150–1000 \text{ m}),$$

$$\tau_{\text{Po}}(150–1000 \text{ m}) = I_{\text{Pb}}(150–1000 \text{ m}) / I_{\text{Pb}}(150–1000 \text{ m}).$$

In the above equations, $I_{\text{Pb}}$ and $I_{\text{Po}}$ are the inventories of $^{210}$Pb and $^{210}$Po, respectively, in the different boxes; $P_{\text{Pb}}$ and $P_{\text{Po}}$ are the removal fluxes of P-Pb and P-Po, respectively, by particles.
sinking out of the box; $J_{\text{Pb}}$ and $J_{\text{Po}}$ are the rates of scavenging of $^{210}\text{Pb}$ and $^{210}\text{Po}$, respectively (if a radionuclide is transferred from solution to the particulate phase by particle scavenging, then $J$ is positive, while $J$ is negative for opposite); and $\tau_P$ and $\tau_D$ are the residence times of the radionuclides in the particulate and dissolved phases, respectively.
The average annual atmospheric $^{210}$Pb deposition flux ($F_{Pb}$) in East Asian is 18540 dpm/m$^2$/yr ($n = 25$, Zhong, 2020), while the atmospheric flux of $^{210}$Po ($F_{Po}$) is usually assumed to be only 10% of $F_{Pb}$ (Turekian et al., 1977; Masqué et al., 2002; Baskaran, 2011). Table 1 shows the scavenging rates, removal fluxes and residence times of $^{210}$Po and $^{210}$Pb in the near-surface and deeper ocean. Inventories of $^{226}$Ra were calculated based on published data from station KH-71-3-S (which is very close to station XTJ1-8) in the w-NPO (Nozaki and Tsunogai, 1976), the $^{226}$Ra inventories in the 0–150 m and 150–1,000 m boxes were calculated to be 14,625 dpm/m$^2$ and 172,530 dpm/m$^2$, respectively (see Table 2).

Within the 0–150 m surface layer, $J_{Po}$ ranged from 1,693 to 17,644 dpm/m$^2$/yr; these positive values confirm that $^{210}$Po was transferred from the dissolved phase to the particulate phase due to the scavenging of sinking particles in the upper ocean. Correspondingly, the removal flux of $^{210}$Po ranged from 3,327 to 34,941 dpm/m$^2$/year, showing a large spatial variation in the w-NPO. The residence times of D-Po ($\tau_{Po,D}$) and P-Po ($\tau_{Po,P}$) in the upper 0–150 m layer ranged from 0.6 to 5.5 years and from 83 to 921 days, respectively. Interestingly, the values of $\tau_{Po,D}$ (0.6–1.2 year) and $\tau_{Po,P}$ (83–93 days) at the three more northerly stations (XTJ1-13, XTJ1-22, and XTJ2-7) were much lower than the values of $\tau_{Po,P}$ (1.6–5.5 year) and $\tau_{Po,P}$ (130–921 days) of the more southerly stations (XTW1-3, XTW1-11, XTJ1-1, and XTJ1-8). Generally, the residence time of 0.6 years was common for dissolved $^{210}$Po under more biologically productive conditions in the ocean (Shimmield et al., 1995). The $\tau_{Po,D}$ at stations XTJ1-13, XTJ1-22, and XTJ2-7 were much shorter than the other four stations, and correspondently, the average POC concentration at 0–150 m box were higher than the other four stations (Table 1), which supports that the residence time of dissolved $^{210}$Po was shorter when the biological activity was higher. In addition, longer particulate residence times are related to inefficient vertical removal processes (low export fluxes), which can be confirmed from the removal flux of $^{210}$Po ($F_{Po}$). As seen from Table 1, the $F_{Po}$ values of the three more northerly stations (3,327–17,242 dpm/m$^2$/year) were much lower than those of the four more southerly stations (14,785–34,941 dpm/m$^2$/year). The above phenomena suggest that the upper waters of the southern part of the w-NPO are characterized by inefficient vertical removal processes and high turnover rates, while strong particle export occurred in the northern part of the w-NPO.

At greater depths (150–1,000 m), the $F_{Po}$ values were negative at all stations except stations XTJ1-13 and XTJ2-7 (Table 1), further demonstrating the occurrence of $^{210}$Po remineralization in the deeper ocean. However, due to the release of $^{210}$Po, $\tau_{Po,D}$ could not be calculated at almost all stations in the deeper layer (150–1,000 m) except at station XTJ1-13. The $F_{Po}$ values were negative at stations XTJ1-1 and XTJ1-8, implying that $^{210}$Po could not be removed from the 150–1,000 m layer to the much deeper (>1,000 m) ocean at these two sampling sites. In addition, the $\tau_{Po,P}$ in the 150–1,000 m layer ranged from 85 to 331 days at the remaining four stations (XTW1-3, XTJ1-13, XTJ1-22, and XTJ2-7).

In the upper 0–150 m, the scavenging rates of $^{210}$Pb ($J_{Pb}$) varied between 18,373 and 18,720 dpm/m$^2$/yr and the removal fluxes of $^{210}$Pb ($F_{Pb}$) ranged from 17,835 to 18,454 dpm/m$^2$/year (Table 1). As atmospheric input is the dominant source term for $^{210}$Pb in the upper ocean, the removal flux of $^{210}$Pb was close to the constant atmospheric $^{210}$Pb deposition flux of 18,540 dpm/m$^2$/year. The values of $\tau_{Pb,D}$ and $\tau_{Pb,P}$ in the upper ocean (0–150 m) were nearly on the same level, ranging from 0.47 to 1.09 year and from 0.33 to 0.98 year, respectively. However, in the deeper ocean (150–1,000 m), the scavenging rate of $^{210}$Pb decreased significantly to a low level of 3,103–3,800 dpm/m$^2$/year, whereas the removal flux of $^{210}$Pb remained at a high level (15,575–17,649 dpm/m$^2$/year). Moreover, the residence times of D-Pb and P-Pb below 150 m were calculated to be 13.1–22.9 and 1.1–5.5 years, respectively.

### Particulate Organic Carbon Export Flux Estimated From $^{210}$Po Deficiency

The scavenging of particle-reactive nuclides produces a deficiency of the daughter nuclides with respect to their parents in the water column. From these daughter deficiencies, Buesseler et al. (1992) developed a method to estimate the POC export flux ($F_{POC}$) from the upper ocean by using $^{234}$Th-$^{238}$U disequilibrium, and this method has been applied extensively. $F_{POC}$ can be derived by multiplying the daughter nuclide removal flux from the euphotic zone by the ratio of the POC concentration to the daughter nuclide activity of the total particulate material. Similarly, based on the $^{210}$Po deficiency, the POC export flux is expressed as follows:

$$F_{POC}(\text{Buesseler}) = \frac{POC}{A_{Po}^P} \times P_{Po}, \quad (19)$$

where $POC/A_{Po}^P$ denotes the measured POC/$^{210}$Po ratio in particulate matter, and $P_{Po}$ denotes the removal flux of $^{210}$Po at the output interface of 150 m.

Similarly, Coale and Bruland (1987) showed that the profile of $^{234}$Th was closely linked to the profiles of nutrients and chlorophyll, implying that the cycling of $^{234}$Th was linked to the cycling of organic matter. This led Eppley (1989) to assume that, if the residence times of POC and $^{234}$Th in surface waters are similar, the $^{234}$Th flux can be used to calculate $F_{POC}$. Because

### Table 2: $^{226}$Ra inventories integrated from 0 to 150 m and from 150 to 1,000 m in the w-NPO.

| Station  | Longitude | Latitude | Date     | Integrated depth (m) | $^{226}$Ra inventory (dpm/m$^2$) | Reference            |
|----------|-----------|----------|----------|----------------------|----------------------------------|----------------------|
| KH-71-3-S | 28°29′N   | 145°06′E | 29-Jun   | 0–150                | 14,625                           | Nozaki and Tsunogai, 1976 |
|          |           |          |          | 150–1,000            | 172,530                          |                      |

Dissolved $^{226}$Ra data is cited from Nozaki and Tsunogai (1976).
$^{210}$Po has stronger particle affinity in organic matter than $^{234}$Th, $^{210}$Po could be superior to $^{234}$Th as a tracer for POC; hence, based on Eppley’s assumption, the $^{210}$Po-based POC flux can also be derived as follows:

$$I_{POC} = \int_0^z (POC) \, dz,$$

(20)

$$F_{POC}(\text{Eppley}) = \frac{I_{POC}}{\tau_{Po-p}},$$

(21)

where $I_{POC}$ stands for the integral inventory of POC in the upper ocean and $\tau_{Po-p}$ is the residence time of $^{210}$Po in the particulate phase.

The POC fluxes estimated by the Eppley and Buesseler models in the upper layer of the w-NPO (0–150 m) are displayed in Table 3 and Figure 9A. From Figure 9A, we can divide these seven stations into two groups. One group (Group I) includes the station XTW1-3 in the Luzon Strait, station XTJ2-7 in the northern Shikoku Basin, and station XTJ1-22 located in the Oyashio Current -Kuroshio Current boundary near the main island of Japan, which showed a higher estimated $F_{POC}$ level of 6–9 mmol C/m$^2$/d (see Table 3 and Figure 9A). The common characteristic of Group I was that all of them were located near the continental shelf of East Asia. However, station XTJ1-22 was also affected by the cold and nutrient-rich Oyashio Current, which appears to promote high primary production and subsequently high POC export flux. Therefore, the FPOC of station XTJ1-22 was the highest among all stations (Figure 9A).

By contrast, the other four stations (XTJ1-1, XTW1-11, XTJ1-8, and XTJ1-13 station) can form another group (Group II), which denoted a lower POC export flux level in the w-NPO (Figure 9A). Concretely, the lowest POC export ($<1$ mmol C/m$^2$/day) was found at station XTJ1-1 in the Marianas Trench, at which the negligible sinking flux of $^{210}$Po (3,327 dpm/m$^2$/year, Table 1) was the most obvious feature. The relatively lower POC export fluxes (1–2 mmol C/m$^2$/day) at stations XTW1-11 and XTJ1-8 were derived from the relatively low sinking fluxes of $^{210}$Po (Table 1) and relatively low particulate POC/$^{210}$Po ratios (Table 3). Geographically, these four stations of Group II (XTW1-11, XTJ1-1, XTJ1-8, and XTJ1-13) were located in the central gyre of the w-NPO, which had the lower particulate POC/$^{210}$Po ratios in total suspended matters comparing with values of the stations (XTW1-3, XTJ1-22, and XTJ2-7) in Group I close to the East Asian continent (Table 3). $F_{POC}$ of stations in Group II were much lower than those of Group I, which may be related to the dominated phytoplankton. The central gyre of the w-NPO (stations of Group II) is an oligotrophic area where the phytoplankton community is dominated by picoplankton (such as smaller diatoms or coccolithophores), and these small size phytoplankton produce a low POC flux due to the oligotrophic living conditions in this region. Interestingly, in these areas, stations in Group II had much higher particulate POC/$^{210}$Po ratios than those of stations in the central gyre of the w-NPO (Group I). For example, the stations XTW1-3, XTJ1-22, and XTJ2-7, located at the margins of the Chinese mainland and the main island of Japan, have higher particulate POC/$^{210}$Po ratios of 181 $\pm$ 16, 184 $\pm$ 16, and 108 $\pm$ 10 $\mu$mol/dpm, respectively. And the other four stations XTW1-11, XTJ1-1, XTJ1-8, and XTJ1-13, located in the central gyre of the w-NPO, have much lower particulate POC/$^{210}$Po ratios of 73 $\pm$ 5, 76 $\pm$ 6, 84 $\pm$ 7, and 69 $\pm$ 5 $\mu$mol/dpm, respectively (Table 2). Generally, smaller phytoplankton cells can scavenge more $^{210}$Po (smaller particles dominate higher particulate $^{210}$Po activity relative to larger particles in the non-oligotrophic area) due to their larger surface area per unit of volume, lowering their particulate POC/$^{210}$Po ratios (Tang et al., 2019).

Overall, the POC fluxes calculated by the Eppley method and Buesseler method ranged from 0.6 to 8.8 mmol C/m$^2$/d and from 0.7 to 8.6 mmol C/m$^2$/d, respectively (Figure 9A). These values are in good agreement with the estimated POC fluxes derived via the steady-state $^{210}$Po-$^{210}$Pb method in other open oceans with similar latitudes (negligible to 8.5 mmol C/m$^2$/day) (Kim and Church, 2001; Stewart et al., 2007; Verdeny et al., 2009; Roca-Martí et al., 2016; Tang et al., 2019; Horowitz et al., 2020). From Figure 9A, although there are some differences, Eppley model and Buesseler model-derived $F_{POC}$ are very similar within uncertainties. Buesseler model estimated $F_{POC}$ were 24 and 43% higher than that of Eppley model for stations XTW1-3 and XTJ2-7, respectively. However, Eppley model-derived $F_{POC}$ was 17% higher than that of Buesseler model.

### Table 3 | POC fluxes and e-ratios derived from $^{210}$Po tracer by the Eppley and Buesseler models.

| Station | Integrated depth | $^{210}$Po-derived $F_{POC}$ (Eppley model) | e-ratio* (Eppley model) | Particulate POC/$^{210}$Po at 150 m | $^{210}$Po-derived $F_{POC}$ (Buesseler model) | e-ratio* (Buesseler model) |
|---------|-----------------|--------------------------------|------------------------|-----------------------------------|--------------------------------|------------------------|
| XTW1-3  | 0–150           | 6.0 $\pm$ 0.3               | 9.7                    | 181 $\pm$ 16                      | 8.6 $\pm$ 0.8                | 13.9                   |
| XTW1-11 | 0–150           | 1.7 $\pm$ 0.2               | 2.7                    | 73 $\pm$ 5                        | 2.5 $\pm$ 0.2                | 4.0                    |
| XTJ1-1  | 0–150           | 0.6 $\pm$ 0.2               | 1.0                    | 76 $\pm$ 6                        | 0.7 $\pm$ 0.1                | 1.1                    |
| XTJ1-8  | 0–150           | 1.6 $\pm$ 0.2               | 2.6                    | 84 $\pm$ 7                        | 2.1 $\pm$ 0.2                | 3.4                    |
| XTJ1-13 | 0–150           | 5.9 $\pm$ 0.4               | 9.5                    | 69 $\pm$ 5                        | 6.6 $\pm$ 0.5                | 10.6                   |
| XTJ1-22 | 0–150           | 8.8 $\pm$ 0.4               | 14.2                   | 184 $\pm$ 16                      | 7.5 $\pm$ 0.6                | 12.1                   |
| XTJ2-7  | 0–150           | 6.2 $\pm$ 0.3               | 10.0                   | 108 $\pm$ 10                      | 7.7 $\pm$ 0.7                | 12.4                   |

The uncertainties in the $^{210}$Po-derived POC flux were estimated based on the propagation of error. The export interface is defined at 150 m. *The primary production for calculating the e-ratio values was 62 $\pm$ 19 mmol C/m$^2$/d (cited from Palevsky et al., 2016).
for station XTJ1-22. The difference of $F_{\text{POC}}$ results given by different models may be related to the assumptions of the models. For example, the Eppley model's assumption that the residence times of POC and particulate daughter radionuclides ($^{234}\text{Th}$ or $^{210}\text{Po}$) should be the same within the upper ocean is possibly too demanding in the real ocean and therefore is often difficult to achieve. Murray et al. (1989) compared the residence times of $^{234}\text{Th}$ and POC in the euphotic zone of the eastern equatorial Pacific and found that the residence time of $^{234}\text{Th}$ was approximately three times less than that of POC. In contrast, the Buesseler model needs to assume only that the same particulate matter serves as the carrier of POC and $^{234}\text{Th}$ (or $^{210}\text{Po}$), which is relatively easy to achieve in the actual sea. For some specific stations, the difference between two models may be significant, hence, more care should be taken when selecting the model. But on a larger spatial scale in our study area, Eppley model can also provide reasonable estimation results. For example, Figure 9B shows that Eppley model-derived $F_{\text{POC}}$ was well correlated with the Buesseler model-derived $F_{\text{POC}}$ in the w-NPO ($n = 7$, slope = 1.06, $R^2 = 0.92$, $p = 0.003$); additionally, the $^{210}\text{Po}$-POC export fluxes based on the Eppley and Buesseler models had very similar spatial distribution features (Figure 9A). These evidences suggest that the hypothesis of the Eppley model is also valid in the w-NPO and this agreement of Eppley model-derived $F_{\text{POC}}$ and Buesseler model-derived $F_{\text{POC}}$ in the w-NPO probably suggest the residence times of $^{210}\text{Po}$ and POC were similar in the upper ocean.
Comparison of Particulate Organic Carbon Export Fluxes Among Different Regions of the Western North Pacific Ocean and Its Surrounding Marginal Seas

The $^{210}$Po-derived POC export fluxes from the upper ocean (5.9–8.8 mmol C/m$^2$/day) at the stations along the edges of the w-NPO (close to the East Asian land masses) were significantly higher (by a factor of 2–14) than the fluxes determined from the interior (oligotrophic) seawater in the w-NPO (i.e., XTW1-11, XTJ1-1, and XTJ1-8: 0.6–2.5 mmol C/m$^2$/d) (Table 3 and Figure 9). Comparing our POC fluxes with other fluxes in different regions of the Pacific Ocean and its surrounding marginal seas (Table 4, in which the POC fluxes were estimated either by sediment traps or by the $^{234}$Th-$^{238}$U and $^{210}$Po-$^{210}$Pb disequilibrium methods), reveals systematic higher POC fluxes occurring in the marginal seas surrounding the w-NPO (Figure 10). For example, the POC export fluxes in the Pacific Ocean were much smaller than those reported for the SCS (2.0–16.7 mmol C/m$^2$/day, $^{234}$Th-$^{238}$U disequilibrium, Zhou et al., 2020), inner shelf of the ECS (2.3 mmol C/m$^2$/day, Zhou et al., 2018), station ALOHA in the North Pacific Subtropical Gyre (0.4–4.0 mmol C/m$^2$/d, from April 1999 to March 2000, Benitez-Nelson et al., 2001; negligible-7.0 mmol C/m$^2$/day, in June/July 2004 with no bloom, Buesseler et al., 2009), station K2 in the western subarctic North Pacific (1.9–10.1 mmol C/m$^2$/d, in August/September 2015 with no bloom, Umhau et al., 2019).

The NPO is a region of particular interest for quantifying the rate and efficiency of the BCP, as this region is a major sink for atmospheric CO$_2$, absorbing 0.5 Pg C/year, ~25% of the total ocean CO$_2$ uptake (Takahashi et al., 2009). As an important part of the NPO, the subtropical oligotrophic w-NPO is a typical nitrate-deficient water region and is characterized by low chlorophyll-a (Chl a) concentrations (Chen et al., 2013), and this region occurs a band of strong CO$_2$ uptake between 30 and 40°N, overlapping with the transition zone between the subtropical and subarctic gyres (Shih et al., 2015). Researchers stated the importance of the contributions of biological carbon export to the NPO sink (Ayers and Lozier, 2012). However, the characteristics of the POC export flux in the w-NPO region are not well understood (Kawakami et al., 2015; Shih et al., 2015). Using the information available, based on the summary of the published POC export fluxes in the North Pacific Ocean (Table 4 and Figure 10), we can obtain a preliminary general view for the distribution feature of POC export flux. Although there were spatial differences, the overall mean level of POC export flux was below 10 mmol C/m$^2$/day. Specifically, the POC fluxes in our study area of the w-NPO (0.6–8.8 mmol C/m$^2$/d) are comparable to the results of other published studies obtained with the $^{234}$Th-$^{238}$U method in other Pacific Ocean regions under general production condition (Figure 10A), such as the subarctic northeastern Pacific Ocean (2.8–7.6 mmol C/m$^2$/d, from February 1996 to February 1997, Charette et al., 1999; 2.0 ± 0.56 (n = 61) mmol C/m$^2$/d in August/September 2018 with no bloom, Buesseler et al., 2020), station ALOHA in the North Pacific Subtropical Gyre (0.4–4.0 mmol C/m$^2$/d, from April 1999 to March 2000, Benitez-Nelson et al., 2001; negligible-7.0 mmol C/m$^2$/day, in June/July 2004 with no bloom, Buesseler et al., 2009), station K2 in the western subarctic North Pacific (1.9–10.1 mmol C/m$^2$/d, in July/August 2005 with no bloom, Buesseler et al., 2009) and the North Pacific tropical and subtropical gyre (0.1–6.3 mmol C/m$^2$/d, in August/September 2015 with no bloom, Umhau et al., 2019).

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**Table 4** | Comparison of the POC export fluxes estimated in different regions throughout the w-NPO and its surrounding marginal seas.

| Number | Locations | Export layer (m) | POC flux (mmol C/m$^2$/d) | e-ratio (%) | Method | References |
|--------|-----------|-----------------|--------------------------|------------|--------|-----------|
| 1      | NE Pacific | 25–100          | 2.8–7.6                  | 5.8 ± 2.6 (n = 3) | 6–13   | $^{234}$Th-$^{238}$U | Charette et al., 1999 |
| 2      | NE Pacific | 120             | 1.0–3.3                  | 2.0 ± 0.6 (n = 61) | 13 ± 5 | $^{234}$Th-$^{238}$U | Buesseler et al., 2020 |
| 3      | North Pacific (ALOHA) | 150         | 1.2–10.4                | 4.0 ± 2.3 (n = 9) | 8.8        | $^{234}$Th-$^{238}$U | Benitez-Nelson et al., 2001 |
| 4      | North Pacific (ALOHA) | 150         | 0–6.9                   | 2.1 ± 2.0 (n = 19) | 12       | $^{234}$Th-$^{238}$U | Buesseler et al., 2009 |
| 5      | Western subarctic North Pacific (K2) | 150        | 1.9–10.1                | 6.3 ± 2.5 (n = 26) | 14        | $^{234}$Th-$^{238}$U | Buesseler et al., 2009 |
| 6      | North Pacific tropical and subtropical gyre | 150        | 0.1–6.3                 | 2.3 ± 2.5 (n = 5) | NA       | $^{234}$Th-$^{238}$U | Umhau et al., 2019 |
| 7      | Western subtropical North Pacific (K2) | 100        | 1.2–23.7                | 8.0 ± 7.9 (n = 6) | 3–46     | $^{234}$Th-$^{238}$U | Kawakami et al., 2015 |
| 8      | SOS (SEATS) | 100             | 2.0–16.7                 | 5.2 ± 4.2 (n = 11) | 9–34    | $^{234}$Th-$^{238}$U | Zhou et al., 2020 |
| 9      | Sea of Japan | 100            | 3.2–26.3                | 12.0 ± 8.3 (n = 12) | 7–56    | $^{234}$Th-$^{238}$U | Kim et al., 2011 |
| 10     | Central ECS | 55–80           | 1.4–2.7                 | 2.1 ± 0.6 (n = 5) | NA       | $^{210}$Po-$^{210}$Pb | Su et al., 2017 |
| 11     | Southern ECS | 70             | 18.7–46.0               | 26.8 ± 13.0 (n = 4) | 13–28  | Trap | Shih et al., 2013 |
| 12     | ECS (inner shelf) | 20–30       | 5.7–66.4                | 32.4 ± 22.2 (n = 5) | 9–91   | Trap | Hung et al., 2013 |
| 13     | ECS (outer shelf) | 100–120     | 4.8–5.3                 | 5.0 ± 0.3 (n = 2) | 5–14   | Trap | Hung and Gong, 2007 |
| 14     | ECS (Kuroshio) | 120          | 2.3                     | 2.3 ± 0.3 (n = 1) | 6       | Trap and $^{234}$Th-$^{238}$U | this study |
| 15     | w-NPO       | 150            | 0.6–8.8*                | 4.4 ± 3.1 (n = 7)* | 1.0–14.2 | $^{210}$Po-$^{210}$Pb | this study |

SEATS (18°N 116°E); the South-East Asian Time-series Study site; ALOHA (22.75°N, 158°W); the Hawaii Ocean Time-series (HOT) station; K2 (47°N 160°E): A key time-series mooring station operated by the Japan Agency for Marine Earth Science Technology.

*POC fluxes estimated by the Eppley model.

$^{2}$POC fluxes estimated by the Buesseler model.
Export efficiency can be defined as the ratio of POC export to primary production (the e-ratio, Buesseler, 1998). Although primary productivity levels were not available for the studied sampling stations, Palevsky et al. (2016) summarized the annual and seasonal net primary productivity (NPP) estimates for the western North Pacific. Since our sampling activities took place from late May to late June 2018, we preliminarily evaluated the e-ratios of the BCP in the w-NPO by using the summer NPP estimates for the regions of the North Pacific (62 ± 19 mmol C/m²/day), hence, the e-ratio values were calculated to be in the range of 1.0–14.2% (Table 3), which are similar to the previously reported values in the North Pacific (Table 4 and Figure 10B; Charette et al., 1999; Benitez-Nelson et al., 2001; Buesseler et al., 2009, 2020). As can be seen from Figure 10B, the e-ratios of biological carbon pump were more variable in the marginal seas than that in the Pacific Ocean. For example, the e-ratios were very consistent (in a level of < 20%) in the NE Pacific, North Pacific, and w-NPO. However, in the marginal seas, the POC fluxes and e-ratios changed over a very wide range (Figure 10B). One extreme case reported in the inner shelf of ECS showed that the e-ratios ranged from 10 to 90% (Hung et al., 2013). The e-ratios in our study were below 15%, suggesting a moderate BCP efficiency in the w-NPO. By comparison, it can be found that the BCP strengths in some marginal seas of the Pacific Ocean [for
example, e-ratio = 9–91% in the inner shelf of ECS (Huang et al., 2013), 13–28% in the Southern ECS (Shih et al., 2013), and 7–56% in the Sea of Japan (Kim et al., 2011) is stronger than that in the interior of the Pacific Ocean (for example, e-ratio = ~10% at station XTJ2-7 in the Shikoku Basin, 3–4% at station XTW1-11 in the Philippine Sea, and ~1% at station XTJ1-1 in the Mariana Trench).

We compiled a table for comparing the export efficiencies of BCP in other open seas, as shown in the Table 5. It is worth noting that the e-ratios are similar in such different areas. For example, the e-ratios < 15% observed in our study are similar to those found in the North Atlantic (Iberian Basin, 0.5–2.5%; Irminger Basin, 3 ± 3%; western European Basin, 5 ± 5%; Iceland Basin, 6 ± 6%; and Labrador Basin, 10 ± 3%, see Tang et al., 2019). The similarity between our study and Tang et al. (2019) is that both sampling activities were carried out in May–June, however, the difference is that our study area was in an oligotrophic subtropical region (medium-low latitudes) and the study area of Tang et al. (2019) was in a temperate system (medium-high latitudes). In addition, Tang et al. (2019) started sampling when the bloom was starting, hence, the POC fluxes and e-ratios were still low. Ceballos-Romero et al. (2016) pointed that export efficiency was significantly discrepant in different stages of the bloom and the export efficiency also showed a strong seasonal variability in the North Atlantic. By comparison with study case of Ceballos-Romero et al. (2016) in the North Atlantic Ocean, we can find that algal blooming has a significant influence on BCP efficiency. Ceballos-Romero et al. (2016) reported two sampling campaigns: one sampling was carried out in the pre-bloom stage (April–May 2010), and the export efficiency was high at that moment (~50%); another sampling was carried out in the decline of the bloom and post-bloom stage (July–August 2010), and the export efficiency was very low (3%). Obviously, our study displayed results from medium-low latitudes with no bloom, therefore, the low e-ratios of 1.0–14.2% were reasonable.

In addition, the e-ratios in the w-NPO (1.0–14.2%, this study) are close to or slightly higher than that in the low-latitude tropical seas and oceans, for example, the equatorial Pacific (1–10%, Buesseler, 1998) and the Arabian Sea (1–10%, Buesseler, 1998; 0.2–11.7%, Subha Anand et al., 2017). This low e-ratios of < 10% might represent the values of BCP efficiency under general condition for tropical ocean, considering the weak seasonal variation of tropical oceans and the oligotrophic features. However, the e-ratios in the w-NPO are much lower than those reported at high-latitude sites (>25%), such as the Eurasian Basin of the Arctic Ocean (60 ± 40%, Roca-Martí et al., 2016), the Bellinghausen Sea (37% at station K, Shimmield et al., 1995) and the Weddell Sea and the Antarctic Polar Front (16–100%, Rutgers van der Loeff et al., 1997), since most of these study cases in the high-latitude seas were related to the bloom situation or the increase of chlorophyll-a, POC and plankton biomass concentrations (Shimmield et al., 1995; Rutgers van der Loeff et al., 1997). The lower export efficiencies observed in the w-NPO may be consistent with the predominance of smaller phytoplankton in the oligotrophic seawater. The oligotrophic feature might weaken the export strength of the BCP in the w-NPO. Indeed, small particles are usually slow-sinking particles that are more likely to be degraded during their descent (Villa-Alfageme et al., 2016), leading to lower export efficiencies. In our study, we also observed negative scavenging rates (J) of 210Po (negative J values mean that radionuclides are transferred from particles to the dissolved phase) below 150 m in the w-NPO (Table 1), which supports the occurrence of particle degradation, leading to a low export efficiency. For example, the lowest Jp values in the 150–1,000 m segment were observed at stations XTW1-11 (Jp = –66,511 dpm/m²/year), XTJ1-1 (Jp = –64,036 dpm/m²/year) and XTJ1-8 (Jp = –62,022 dpm/m²/year), and correspondingly, these three stations have the lowest e-ratios (based on Eppley model: 1.0–2.7%; based on Buesseler model: 1.1–4.0% Table 3).

**Table 5** | Summary of carbon export efficiencies (e-ratios) based upon 210Po or 234Th approach in global open oceans.

| Ocean       | Site                   | Longitude | Latitude | Observation time       | Method               | e-ratio            | References                  |
|-------------|------------------------|-----------|----------|------------------------|----------------------|--------------------|-----------------------------|
| Atlantic Ocean | Iberian basin          | 15°W-9°W  | 39°-42°N | May–June 2014          | 210Po-210Pb method   | 0.5–2.5% (n = 2)     | Tang et al., 2019           |
|             | Irminger basin         | 43°-35°W  | 59°-61°N |                        | 210Po-210Pb method   | 3 ± 3% (n = 2)       |                            |
|             | Western European basin | 24°-19°W  | 46°-51°N |                        | 210Po-210Pb method   | 5 ± 5% (n = 3)       |                            |
|             | Iceland basin          | 32°-25°W  | 55°-60°N |                        | 234Th-238U method    | 6 ± 6% (n = 2)       | Buesseler, 1998             |
|             | Labrador basin         | 52°-45°W  | 52°-59°N |                        | 234Th-238U method    | 10 ± 3% (N = 3)      |                            |
| Pacific Ocean | Equatorial Pacific     | 95°-170°E | 12°-12°S | Spring and fall 1992   | 234Th-238U method    | 1–10% (n = 14)       | Subha Anand et al., 2017    |
| Indian Ocean | Arabian Sea             | 57°-65°E  | 10°-18°N | Aug.–Sept. 1995        | 234Th-238U method    | 1–10% (n = 19)        |                             |
| Indian Ocean | Arabian Sea             | 70°-87°E  | 25°S-19°N| March–April 2014       | 234Th-238U method    | 0.2–11.7% (n = 13)    |                             |
| Pacific Ocean | w-NPO                  | 118°-153°E| 20°-40°N | May–June 2018          | 210Po-210Pb method   | 1.0–14.2% (n = 7)     | This study                  |
| Arctic Ocean | Eurasian Basin         | 17°-131°E | 81°-88.8°N| Aug.–Sept. 2012        | 210Po-210Pb and 234Th-238U method | 60 ± 40% (n = 7)     | Roca-Martí et al., 2016     |
| Southern Ocean | Bellinghausen Sea       | 84°-56°W  | 67°-36°S | Nov.–Dec. 1992         | 210Po-210Pb and 234Th-238U method | 37% (n = 1)          | Shimmield et al., 1995      |
| Southern Ocean | Weddell Sea/Polar Front | 49°-35°W  | 47°-57°S | Oct.–Dec. 1992         | 234Th-238U method    | 16–100% (n = 20)       | Rutgers van der Loeff et al., 1997; Buesseler, 1998 |
CONCLUSION

In this study, we reported the vertical distributions of D-Po, P-Po, D-Pb, and P-Pb activities in the w-NPO during late spring and early summer to constrain the particle dynamics (scavenging rates and removal fluxes of $^{210}$Po and $^{210}$Pb), and to estimate the carbon export production (POC export flux). More than 50% of the radionuclides were found in the dissolved phase below a depth of 50 m, while a small proportion was associated with the particulate phase. However, the percentages of P-Po and P-Pb were higher than 50% (even up to 90%) in the surface seawater (0–50 m). $^{210}$Po deficits relative to $^{210}$Pb were observed in the upper 150 m at all the stations, however such $^{210}$Po deficiencies can even extend to deeper ocean (such as 500 m) at some stations. In the interior of the w-NPO, the excess $^{210}$Po (total $^{210}$Po/$^{210}$Pb > 1) activities in the deeper ocean (500–1,000 m) at stations XJTJ-1 and XJTJ-8 were attributed to the release of $^{210}$Po due to biogenic particle release during sinking. Based on a conceptual one-dimensional irreversible scavenging model, the residence times of $^{210}$Po and $^{210}$Pb in the w-NPO were obtained. In the upper ocean (0–150 m), the residence times of D-Po ($\tau_{P-Po}-D$, 0.6–5.5 year) were much higher than those of P-Po ($\tau_{P-Po}-P$, 83–921 days), while in the deep ocean (below 150 m), the residence times of D-Pb were calculated to be 13.1–22.9 year, 2–10 times longer than those of P-Pb (1.1–5.5 year).

Based on $^{210}$Po-$^{210}$Pb disequilibrium, we found that the Eppley model-derived POC export fluxes agreed well with those derived from the Buesseler model in the upper ocean of the w-NPO, suggesting that particulate $^{210}$Po and POC may have similar residence time in the water column. Overall, the $^{210}$Po-derived POC fluxes varied spatially, ranging from 0.6 to < 9 mmol C/m²/day, with the highest export fluxes at stations close to the East Asian continental shelf and the lowest export fluxes at stations in the ocean basin. We concluded that POC export fluxes tended to increase with decreasing distance from the continental margins. The POC export efficiencies also showed regional differences even within the same basin, with e-ratio values ranging from 1.0 to 14.2%, suggesting a moderate BCP efficiency in the w-NPO. The low export efficiencies may be associated with the dominance of smaller particles and particle release below the euphotic zone. The negative scavenging rates of $^{210}$Po (-$J_{P-Po}$) below 150 m at stations in the central basin of the w-NPO supports the occurrence of particle dissolution. Comparing with the reported e-ratios in other open seas, relatively higher carbon export efficiencies occurred at high-latitude sites.

DATA AVAILABILITY STATEMENT

The original contributions presented in the study are included in the article/Supplementary Material, further inquiries can be directed to the corresponding author/s.

AUTHOR CONTRIBUTIONS

QZ was responsible for conceptualization, methodology, sample analysis, data processing, writing original draft, writing–review, and revising. DH, TY, and JD provided a financial support and helped in article reviewing. JL, JJ, and JN contributed to the review of the manuscript. All authors contributed to the article and approved the submitted version.

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SUPPLEMENTARY MATERIAL

The Supplementary Material for this article can be found online at: https://www.frontiersin.org/articles/10.3389/fmars.2021.700524/full#supplementary-material

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