

The climatology of the deep particle flux in the oligotrophic western North Atlantic gyre, 1978–2022

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ABSTRACT

The oceanic particle flux controls, in part, ocean biogeochemical cycles and long-term carbon sequestration. The Oceanic Flux Program (OFP), the longest running time series of its kind, has continuously measured the deep particle flux in the oligotrophic Sargasso Sea southeast of Bermuda since 1978. This paper describes the deep flux climatology at the OFP site over the 1978–2022 time period. Mass flux at 500 m, 1500 m and 3200 m depths has averaged 27.2, 34.8 and 36.8 mg m⁻² d⁻¹, respectively. Carbonates comprise ~ 60 % of the flux, with lesser amounts of organic matter, opal and lithogenics. Flux magnitude and composition vary seasonally with large interannual variability, particularly in the winter/spring flux maximum. Flux frequency distributions are strongly skewed, especially at 500 m depth where flux magnitude and compositional variability are highest. Flux seasonality, skewness and compositional heterogeneity decrease markedly with depth. A significant component of the deep flux is sourced from large particle production in the deep water column (e.g. suspended material repackaging) rather than directly from the overlying export flux. Lithogenic flux increases five-fold between 500 m and 3200 m depths, underscoring the importance of deep lateral advection and lithogenic particle removal via particle cycling processes. Multidecadal averages in deep carbon fluxes are compared with concurrent monthly data on overlying net primary production (PP) and surface export flux (EF, measured by drifting traps at 150 m depth) at the nearby Bermuda Atlantic Time Series (BATS) site. Carbon fluxes are temporally coherent throughout the water column (within the sample resolution) and lag primary production by ~ one month. Approximately 0.6 %, 0.5 % and 0.4 % of PP reaches the 500, 1500 and 3200 m depth horizons, respectively, with the highest depth penetration occurring during the Jan-Mar period of peak primary production. Annually, 7.6 % and 6.2 % of the EF reaches the 1500 and 3200 m depth horizons, respectively, with the highest transfer efficiency (9.1 % and 7.4 %, respectively at 1500 and 3200 m depths) during the post-bloom (Apr-Jun) period. The OFP flux climatology summarized here provides an important baseline for assessing future consequences of a changing climate on ocean functioning in the oligotrophic North Atlantic gyre.

1. Introduction

Understanding the mean state and natural variability of the present-day ocean is increasingly important to predict the future impacts of a changing climate on ocean functioning (e.g., Brandão et al., 2021; Dunne, 2023; Nowicki et al., 2022; Riebesell et al., 2009; Yamamoto et al., 2018). The immense value of continuous long-term ocean observations to provide fundamental data on systematic patterns of temporal variability of key ocean parameters is widely appreciated (Benway et al., 2019 and references therein; Ducklow et al., 2009; Edwards et al., 2010). Examples of open ocean time-series that have provided sustained, multidecadal and multifaceted records of biogeochemical

variables include the Porcupine Abyssal Plain (PAP) time series (Hartman et al., 2012), the Bermuda Atlantic Time Series (BATS, Lomas et al., 2013), the Carbon Retention in a Colored Ocean (CARIACO) time series (Muller-Karger et al., 2019), the Hawaii Ocean Time Series (HOT) (Church et al., 2013), the Station M time series (Smith et al., 2020), Ocean Station PAPA (Freeland, 2007) and the Southern Ocean Time Series (SOTS, Trull et al., 2019). Multidimensional records generated by these and other time series have been instrumental for assessing how key ocean parameters covary over space and time (e.g., Bates et al., 2014; Couret et al., 2023; Montes et al., 2016; Nodder et al., 2016; Trull et al., 2019). This has enabled functional linkages and causal relationships among processes to be inferred for in-depth short-term process study

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campaigns (e.g. EXPORTS, Siegel et al., 2016) and ocean model explorations (e.g., DeVries and Weber, 2017; Henson et al., 2012; Lima et al., 2014; Weber and Bianchi, 2020).

The particle flux is a fundamental ocean process that controls the downward transport of material and energy through the water column and plays a key role in ocean biogeochemical cycles (De La Rocha and Passow, 2014; Siegel et al., 2023; Ittekkot et al., 1996) and references therein). Sinking particles transport photosynthetic energy in the form of organic carbon to depth, providing the main food source that fuels life within the ocean interior. Incorporation of nutrient elements into biogenic materials during photosynthesis and the depths to which biogenic particles sink and degrade controls nutrient redistribution by ocean currents and, in turn, spatial patterns of ocean productivity. Chemical scavenging of dissolved phases onto sinking particles and aggregation of suspended materials into large, sinking particles removes many elements and dissolved materials from the water column and cleanses the ocean of accumulated debris. The particle flux also factors in long-term sequestration of carbon and other elements as materials exported to depths greater than the maximum wintertime mixing depth are isolated from the surface on decade to century time scales while materials exported to > 1000 m depth may be sequestered on time-scales up to that of ocean turnover (DeVries, 2022; DeVries et al., 2012; Guidi et al., 2016; Usbeck et al., 2003).

Deep-ocean sediment trap time series have contributed immensely to our understanding of the oceanic particle flux. When advances in ocean technology enabled the long-term deployment of sediment traps during the 1970s, it became quickly apparent that the deep ocean environment was not constant and static as commonly believed (Sanders, 1968) but intimately linked to the overlying surface ocean productivity via a “rain of particles” (Deuser et al., 1981; Honjo et al., 2008; Lampitt, 1985). In the ensuing years, sediment trap time-series have documented both the geographical variation of the particle flux across different ocean biogeographical regimes and the influences of myriad biogeochemical and physical processes on both particle flux generation in the surface waters and particle cycling within the water column (e.g., Fischer et al., 2020; Francois et al., 2002; Frigstad et al., 2015; Hartman et al., 2012; Honda, 2020; Honda et al., 2018; Honjo and Manganini, 1993; Lampitt et al., 2023, 2010; Nodder et al., 2016; Rixen et al., 2019; Salter et al., 2023; Timothy et al., 2013; Waniek et al., 2005; Wynn-Edwards et al., 2020). Given the central role that the particle flux plays in long-term carbon sequestration, a substantial research effort has been focused on synthesizing sediment trap data to ascertain the controls on carbon transfer efficiency through the water column (e.g., Antia et al., 2001; Cael et al., 2021; De Melo Viríssimo et al., 2022; Helmke et al., 2010; Henson et al., 2012; Lutz et al., 2002; Marsay et al., 2015; Henson et al., 2022).

The Oceanic Flux Program (OFP), located in the oligotrophic deep waters of the subtropical western North Atlantic gyre southeast of Bermuda, is the longest and most continuous deep ocean particle flux time series in the world, with a > 90 % temporal coverage of the 3200 m particle flux over the last 45+ years. The OFP provided the first direct evidence of seasonality in the deep ocean and its linkage with the seasonal cycle in primary production via the particle flux (Deuser, 1986; Deuser et al., 1981). OFP time-series has since produced an unrivaled record of temporal variability in the “biological pump”, a term loosely defined as material transfer from the surface to the deep ocean due to ecosystem processes. The OFP has documented variability in material fluxes to the deep ocean over time scales ranging from weeks to years (Conte et al., 2019, 2001, 1998; Conte and Weber, 2014). Much of the nonseasonal variability has been linked to short-lived phytoplankton bloom/bust events that are induced by transient nutrient upwelling associated with mesoscale eddy and frontal activity (Conte et al., 2003; Salmon et al., 2015; Shatova et al., 2012) or extreme weather events (Pedrosa-Pàmies et al., 2019, 2018). OFP results have provided clear evidence of the intensity of biological reprocessing of the sinking flux within mesopelagic waters (Chuang et al., 2013; Conte et al., 2019,

2003, 2001; Huang and Conte, 2009; Pedrosa-Pàmies et al., 2018; Shatova et al., 2012) and of the roles that lateral advection and suspended particle aggregation play in deep water flux generation (Conte et al., 2019; Huang and Conte, 2009; Jickells et al., 1990).

This paper focuses on the climatology of the deep particle flux in this region, compiled from over four decades of OFP observations. We synthesize key particle flux attributes such as flux frequency distributions, seasonality and depth modifications, and discuss how these inform our understanding of particle cycling within the ocean interior. We compare the seasonal climatology of the deep organic carbon flux measured by the OFP traps with concurrent *in-situ* measurements of overlying primary production and upper ocean carbon export flux measured by the Bermuda Atlantic Time Series (BATS), and provide direct estimates of carbon flux attenuation and long-term carbon sequestration in the North Atlantic oligotrophic subtropical gyre.

2. Study area

The Oceanic Flux Program (OFP) sediment trap mooring site lies about 75 km southeast of Bermuda in the northern Sargasso Sea of the western North Atlantic, a region characteristic of the vast expanses of oligotrophic open ocean gyres (Joyce and Robbins, 1996; Talley, 1996) (Fig. 1). Currents in the region are weak, with mean velocities of $\sim 10 \text{ cm s}^{-1}$ at 500 m depth and decreasing to $< 1 \text{ cm s}^{-1}$ below 1500 m depth (McKee et al., 1981). Under these hydrographic conditions, deep traps exhibit excellent trapping efficiency (Yu et al., 2001), as demonstrated by the study of Bacon et al. (1985) who estimated an annually-integrated trapping efficiency of $105 \pm 17 \%$ for the 3200 m OFP trap based upon fluxes of ^{230}Th and ^{231}Pa .

The Bermuda region is one of the most extensively studied open ocean sites in the world and hosts several ongoing, complementary time series which anchor numerous short-term research studies (Fig. 1 inset): The Hydrostation S time-series (1954-present) profiles water column hydrographic properties (Michaels and Knap, 1996; Phillips and Joyce, 2007; Steinberg et al., 2001); the OFP moored sediment trap time series (1978-present) measures the deep ocean particle flux (Conte et al., 2019, 2001; Conte and Weber, 2014; Deuser, 1986), and the Bermuda Atlantic Time-Series (BATS, 1988-present) makes monthly ship-based measurements of upper ocean biogeochemical parameters (Lomas et al., 2013; Michaels and Knap, 1996; Steinberg et al., 2001). Between 1994 and 2007, the Bermuda Testbed Mooring (BTM) measured upper ocean physical and bio-optical parameters at a location halfway between the OFP and BATS sampling area (Dickey et al., 2001; McGillicuddy et al., 1998). In addition, atmospheric studies on Bermuda have been conducted since the 1980 s (e.g., Aldhaif et al., 2021; Galloway et al., 1988; Prospero et al., 1997; Savoie et al., 2002; Steen-Larsen et al., 2014). The synergies among the Bermuda research programs provide a uniquely integrated view of the interactions among atmosphere–ocean physics, chemistry and biology that drive ocean biogeochemistry and the particle flux.

The seasonal cycle in production and upper ocean biogeochemistry in the Bermuda region has been described in detail (Lomas et al., 2013; Steinberg et al., 2001). Briefly, the wintertime mixed layer reaches a maximum depth of 250–300 m in late January to early February. With the onset of seasonal stratification in late February–early March, a short spring bloom develops. As stratification increases, the surface mixed layer shallows and becomes nutrient-depleted, leading to the development of a pronounced subsurface chlorophyll maximum at approximately 100 m depth. Intensifying stratification limits vertical mixing and nutrient upwelling, and in turn, primary production during the summer and early autumn. Late autumn brings seasonal cooling and increased storminess that leads to gradual mixed layer deepening.

There is significant inter-annual and < seasonal variability in this general pattern. Particularly during periods of low water column stability in early winter and spring, nutrient upwelling induced by passage of mesoscale eddies and synoptic-scale weather systems (Black and

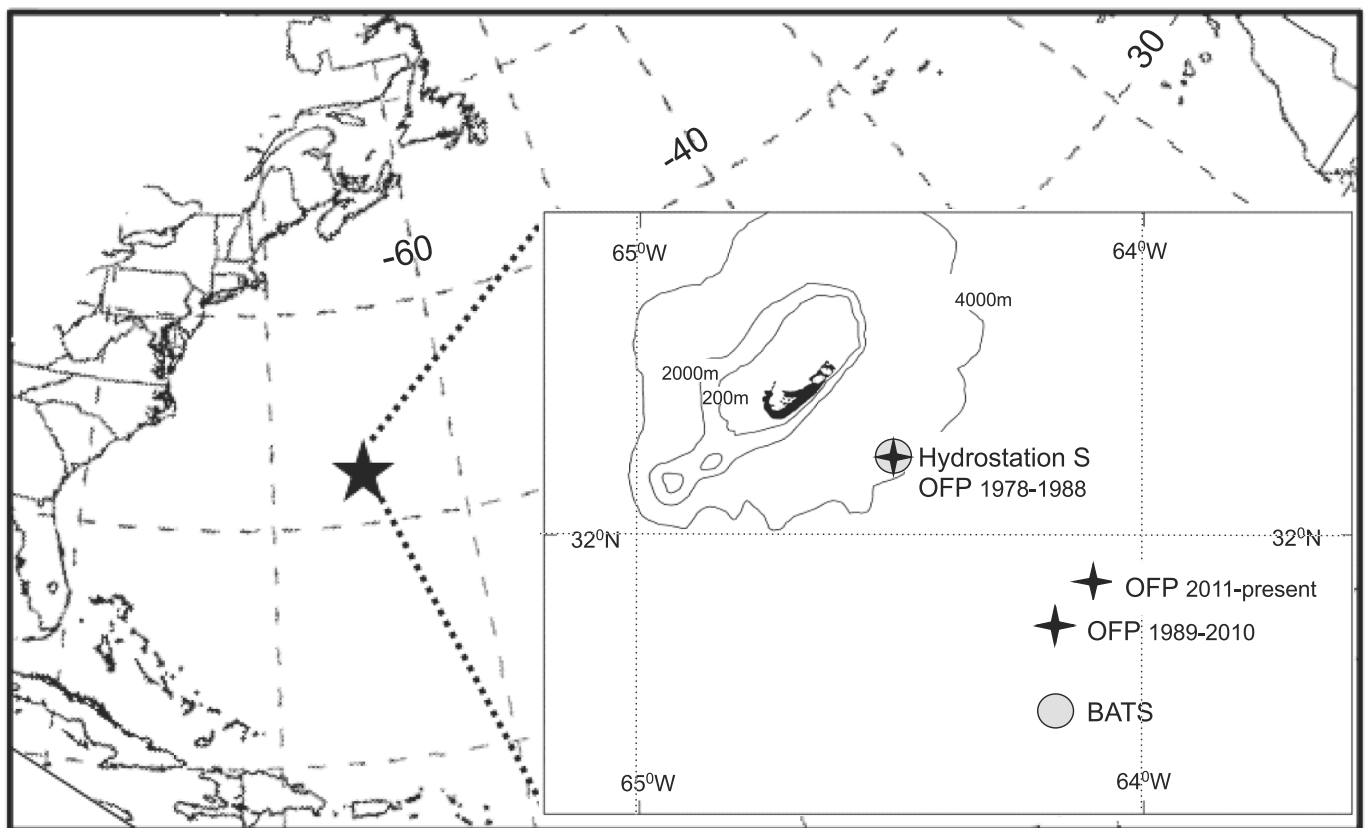


Fig. 1. The Bermuda Time Series site southeast of Bermuda, showing the locations of the Oceanic Flux Program (OFP), the Bermuda Atlantic Time Series (BATS) and the Hydrostation S time-series. The basemap is the NOAA Difax projection map of the North Atlantic. The OFP mooring was located at the Hydrostation S site until 1989 when it was moved further offshore to $31^{\circ}50'N$, $64^{\circ}10'W$. In 2011 the mooring was relocated to its present position of $31^{\circ}55'N$, $64^{\circ}05'W$.

Dickey, 2008; McGillicuddy et al., 1998; McNeil et al., 1999) may result in short bursts of primary production and export flux which, in turn, drives short-term changes in the deep particle flux (Conte et al., 2003, 1998; Lomas et al., 2009; Pedrosa-Pàmies et al., 2018).

3. Methods

3.1. Sample collection

The OFP mooring site is located in the abyssal plain southeast of the Bermuda pedestal (Fig. 1). Initially the mooring was located at the Hydrostation S site and called SCIFF (Seasonal Changes in Isotopes and Flux of Foraminifera, Deuser et al., 1981). The SCIFF mooring consisted of a single 1.54 m^2 conical sediment trap located at 3200 m water depth. In 1984, upon the acquisition of a larger ocean-capable ship, the SCIFF mooring was relocated further offshore to $31^{\circ}50'N$, $64^{\circ}10'W$ (4400 m water depth), additional traps were added at 500 m and 1500 m depths and the time series was renamed the Oceanic Flux Program. Hydrographic conditions at the Hydrostation S site are comparable to conditions farther offshore in the BATS/OFP locale (Phillips and Joyce, 2007), although Hydrostation S's closer proximity to the Bermuda Pedestal may induce slightly higher nutrient upwelling and opal production as compared to the offshore OFP site (Conte et al., 2001). In 2010, the mooring was relocated slightly northeast to its current position of $31^{\circ}55'N$, $64^{\circ}05'W$.

The SCIFF trap had a single collection cup under the trap funnel and the mooring was recovered and redeployed every two months. In 1989, the SCIFF traps were replaced by 0.5 m^2 PARFLUX traps (McLane Labs, Falmouth MA) that introduced a microprocessor controlled rotating sample carousel. This allowed the sample resolution to be increased to a nominal biweekly flux integration period.

In an attempt to keep zooplankton swimmers from entering sample cups, plastic mesh ($\sim 0.5\text{ cm}$ mesh size) was installed over the cup orifices of the 500 m and 1500 m traps in October 1992. This was removed in January 1996 because it interfered with quantitative flux collection. Because traps fitted with mesh under-sampled, the 500 m and 1500 m data collected between October 1992 – January 1996 (black bars in Fig. 2) are omitted from the flux statistics.

The present configuration of the OFP mooring includes McLane 0.5 m^2 Parflux traps at 500, 1500 and 3200 m depths, an upward looking 150 kHz Workhorse ADCP (Teledyne/RDI, Poway CA) at the top of the mooring (300 m depth) to monitor currents and mesoscale physical variability (e.g., passage of eddies) and acoustic backscatter intensity (ABI, a macrozooplankton biomass proxy) in the 0–300 m water column, a Seabird MicroCAT ODO at 500 m depth to measure changes in the lower 18°C mode water properties (T, S, O_2), and a second MicroCAT at 4000 m depth to monitor deep ocean T and S as part of the OceanSITES global initiative. Between 2005 and 2010, prior to the acquisition of the ADCP, a two dimensional acoustical current meter (FSI ACM 2D, Falmouth Scientific, Falmouth MA) was deployed at 750 m depth. Dual Teledyne/Benthos releases are deployed at 4000 m depth for mooring recovery. A MetOcean (Dartmouth, Nova Scotia, Canada) IIRIDIUM satellite transmitter, radio and strobe are mounted on the ADCP buoy to aid in recovery and provide emergency satellite transmission in event of accidental mooring release.

The OFP mooring is designed to minimize hydrodynamic effects. The top of the mooring is at 300 m depth to reduce mooring interaction with upper ocean currents. A section of $\frac{3}{4}$ " plaited nylon above 1500 m trap adsorbs line vibration to reduce mooring line strumming and isolates the lower mooring section. Six glass flotation spheres (152 kg net buoyancy) encircle the top of each trap funnel to maximize buoyancy at the top of the trap to keep the trap upright and reduce trap tilt.

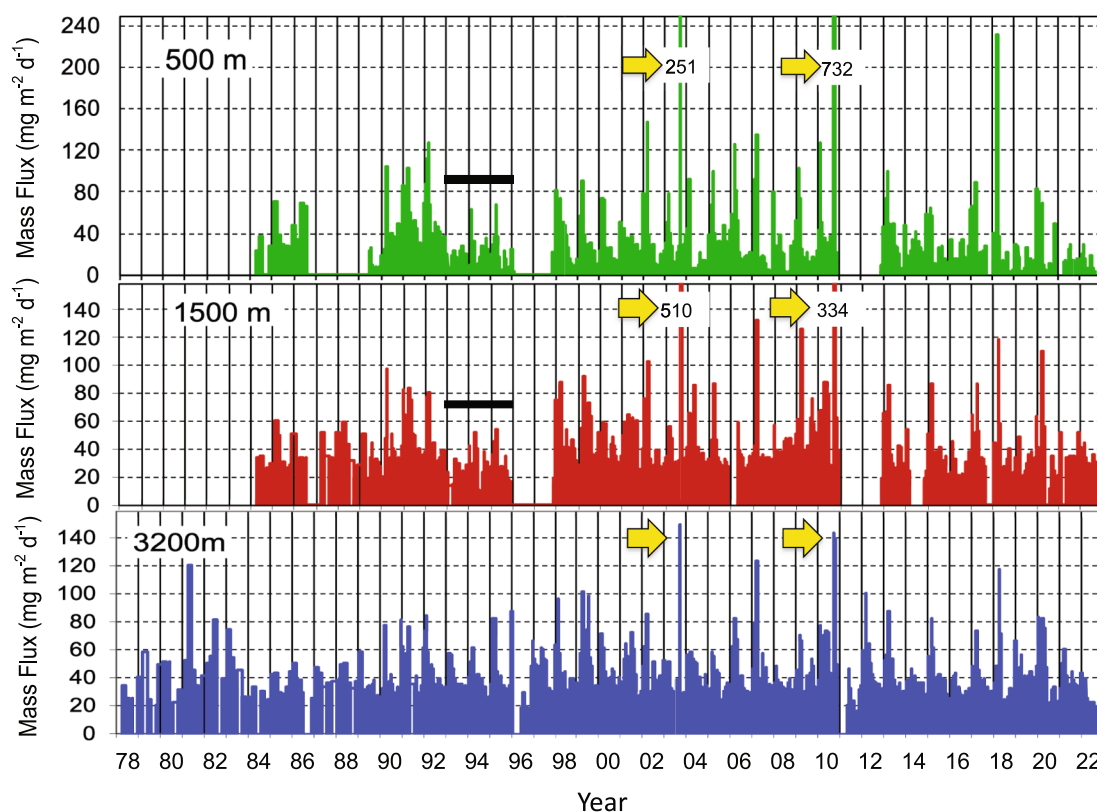


Fig. 2. The OFP record of mass flux, 1978 through 2022. The width of the bars indicates the sample integration period, approximately bimonthly from 1978 to 1989 and biweekly since 1989. The extreme fluxes seen in 2003 and 2010 (yellow arrows) were plumes of resuspended detrital carbonate sediments advected from the Bermuda Pedestal by Hurricanes Fabian and Igor, respectively. Black bars indicate the 1992–1995 period when screens installed at the orificies of the 500 and 1500 m trap cups interfered with quantitative sampling (see text). These samples and those collected during the hurricane plume events have been omitted from flux statistics.

Sample bottles are low density polyethylene (LDPE) and filled with seawater brine of $\sim 40 \text{ g kg}^{-1}$ (40 ‰). Between 1984–1989, the 500 and 1500 m traps were poisoned with sodium azide and the 3200 m trap brine was not poisoned. In 1989, the poison was switched to mercuric chloride. Since 1996, all trap cups have been filled with trace organic/metal clean HgCl_2 poisoned brine made from Sargasso Sea deep water collected at a depth of 3000 m using trace metal clean GoFlo bottles (General Oceanics, Miami FL) and then transferred to trace-metal clean carboys under a laminar flow hood. The seawater is concentrated by freezing and then poisoned with ultra-high purity mercuric chloride (Sigma Aldrich) at a concentration of 0.74 mmol L^{-1} . An intercalibration found no bulk compositional differences between unpoisoned and mercuric chloride poisoned samples (Conte et al., 1998).

Further details on the OFP mooring and earlier sampling methods are described in Deuser et al. (1981), Conte et al. (2001) and Conte et al. (2019). Appendix 1 summarizes the deep water currents measured by the ADCP on the OFP mooring and also mooring tilt data for the period when the 2D ACM current meter was deployed at 750 m depth.

Siegel and Deuser (1997) (see Corrigendum, Siegel and Armstrong, 2002) modelled the “statistical” funnels above the OFP traps for trajectories of particles settling from the surface at different sinking speeds. The model used a similar current regime as that measured by the OFP, which they estimated using a combination of float drifter, mooring and geostrophic data. Model results indicate that the trap collection area is centered over the mooring site at all depths (i.e. no downstream drift) and that the traps integrate over a mesoscale area of surface biogeochemical variability. For example, the 95 % statistical funnels of the 500, 1500 and 3200 m traps for particles sinking at 200 m d^{-1} are estimated have surface collection areas of approximately 31400, 70700, and 94200 km^2 , respectively.

3.2. Analytical

3.2.1. Cleaning procedures

Prior to 1996, trap cups and glassware were cleaned in general laboratory cleaning solutions, soaked in 1 N HCl, and rinsed with distilled water. Since 1996, all glass and Teflon ware for sample processing is cleaned in a NoChromix (Godax Labs)/sulfuric acid oxidizing solution and glassware combusted overnight at 380°C . In 1998, stringent cleaning steps for trace analyses were added (Conte et al., 2019). Trap cups and plastic/Teflonware are soaked in 6 N HNO_3 (6 h), 4 N HCl (12 hrs), and 0.4 N HCl (4–5 d), and then rinsed and filled with MilliQ water and capped with lids fitted with cleaned Teflon liners for storage. Just prior to deployment, cups are rinsed with MilliQ water and filled with poisoned brine under a laminar flow hood.

3.2.2. Sample processing

Recovered samples are refrigerated (4°C) until analysis. From 1978 to 1995, samples were wet sieved through a series of stainless steel sieves (1000, 500, 125 and $37 \mu\text{m}$ mesh) into a beaker. Material on the sieves was rinsed into pre-weighed glass petri dishes using distilled water. The largest two size fractions were examined under a dissecting microscope to remove swimmers, defined as animals which actively enter the trap collection cup alive. All four fractions were then oven dried at 50°C and weighed. The remaining $< 37 \mu\text{m}$ material was concentrated using a flow through centrifuge, dried and weighed.

In 1996, the $37\text{--}125 \mu\text{m}$ and $< 37 \mu\text{m}$ size fractions were combined as a single $< 125 \mu\text{m}$ size fraction and methods were upgraded for trace organic and elemental analyses. Before sample processing, a subsample of supernatant is removed for analysis of dissolved constituents. A second portion of the supernatant is used for rinsing during the sieving

process. The sample is then resuspended and passed through 1000 and 500 μm solvent-rinsed 316 stainless steel sieves. The $> 1000 \mu\text{m}$ fraction is rinsed into a pre-weighed Pyrex petri dish, swimmers removed under a stereomicroscope, and then quantitatively photographed (detailed in Shatova et al., 2012) before oven drying at 55°C . The 500–1000 μm fraction is rinsed into a Teflon petri dish, swimmers removed as above and recombined with the $< 500 \mu\text{m}$ material.

The total $< 1000 \mu\text{m}$ picked flux material is quantitatively split into ten subsamples under a laminar flow hood using a McLane rotary splitter (McLane Laboratory, Falmouth, MA) equipped with a Teflon-coated splitting tray and spigots. Three subsamples are set aside for organic analyses and one for trace elemental which is subsampled for genomic analyses. The remaining splits (60 %) are recombined and fractionated into 500–1000, 125–500, and $< 125 \mu\text{m}$ size fractions. The 500–1000 μm and 125–500 μm fractions are rinsed into pre-weighed Pyrex petri dishes, quantitatively photographed as part of the OFP image archive (Shatova et al., 2012) and then oven dried (55°C). The dried fractions in their petri dishes are weighed using a Mettler balance ($\pm 0.01 \text{ mg}$ accuracy). The $< 125 \mu\text{m}$ fraction, collected in a Teflon beaker, is transferred to a pre-weighed Teflon test tube, concentrated by centrifugation, briefly rinsed with MilliQ water and freeze dried for dry weight determination and bulk component analyses.

3.2.3. Swimmers and fish contamination

Appendix 2 summarizes 20 years of observational data on swimmers and fish contamination in OFP trap samples. Zooplankton swimmers remain intact (non-herniated) in the mercuric chloride trap cup solution and can be easily picked out and archived during sample processing; thus, the primary impact of zooplankton on flux collection is their potential for grazing particles landing on the trap inner surfaces before they reach the collection cup. The most common swimmers are various crustaceans (amphipods, copepods, mysids, euphausiids, shrimps, ostracods, and cyprid larvae). The 500 m trap samples has significantly higher swimmer abundances than traps at 1500 m depth. Swimmer abundances in the 3200 m trap are very low, with $< 5 \%$ of samples having more than 1–2 individuals of any taxa.

Fish mainly affect the 500 m trap, with approximately 10.3 % of samples contaminated by fish debris and omitted from the database. In contrast, $< 1\%$ of 1500 m and 3200 m trap samples are contaminated by fish debris and omitted from the database.

Overall, the data indicate that the 500 m trap may be moderately impacted by swimmer and fish activity, but that swimmers or fish have little to no impact on measured trap fluxes at 1500 m or 3200 m depths.

3.2.4. Carbonate analysis

Before 1995, carbonate was determined manometrically on a vacuum line system (Deuser et al., 1981). Since 1996, carbonate has been analyzed by coulometry (UIC Inc., Joliet, IL). Samples (2–3 mg each) are run in duplicate or triplicate. CaCO_3 and well-characterized reference standards of deep-sea sediment and SCIFF trap material are analyzed with each run ensure data intercalibration. Typical sample c.v. is $\sim 3 \%$. Many pre-1996 archived samples have been reanalyzed using coulometry.

3.2.5. Organic carbon and nitrogen analysis

Prior to 1992, organic carbon (C_{org}) was determined manometrically on the same vacuum line as used for carbonate after combustion of acidified samples (Deuser et al. 1981). Between 1992 and 1995, total carbon and nitrogen were determined on a Perkin Elmer 240 CHN instrument; organic carbon was calculated as the difference between total and carbonate carbon. Between 1995–2004, organic carbon and nitrogen were determined directly on pre-acidified samples using either a Carlo Erba or Perkin Elmer 240 CHN instrument.

Since 2004, organic carbon and nitrogen concentration and isotopic composition have been analyzed by mass spectrometry using either a Europa 20–20 or GV Isoprime mass spectrometer. Roughly 65 % of

archived samples from before 2004 have been reanalyzed using mass spectrometric methods. Prior to analysis, samples are acidified to remove carbonates using a modification of the Verardo et al. (1990) method. Samples are analyzed in duplicate or triplicate. Certified standards and reference standards of sediment-trap material and sediments run with each analysis ensure intercalibration. Typical sample c.v. is $\sim 3 \%$. C_{org} is converted to organic matter (OM) using the equation $\text{OM} = 2.3 \times \text{C}_{\text{org}}$.

3.2.6. Phosphorus analysis

Total phosphorus is reported, i.e. the sum of particulate phosphorus and dissolved phosphorus lost into the supernatant during sample collection. Particulate phosphorus has been analyzed for OFP samples collected between 2001–2015 using a fusion/HR-ICPMS method developed specifically for low level analysis of sediment trap material (Conte et al., 2019; Huang et al., 2007). Data comparison of samples analyzed for particulate phosphorus by the standard Aspilla method (Aspilla et al., 1976) and fusion/HR-ICPMS methods indicate no difference between the two methods (Conte et al., 2019). Typical sample c.v. is $< 5 \%$.

Dissolved phosphorus in the supernatant is analyzed by the phosphomolybdate blue complex reaction (Koroleff and Grasshoff, 1976) using a Lachat autosampler (QuikChem FIA⁺ 8000 series, XYZ Autosampler 500 Series) with artificial seawater brine used as the carrier. Standards are made using brine from dipped trap cup blanks (i.e., cups filled with mercuric chloride brine that are deployed on the traps but do not sample), with dilutions matched to the sample dilutions. Typical sample c.v. is $\sim 2 \%$.

3.2.7. Opal and lithogenic analyses

Total particulate Si and Al have been analyzed in OFP samples collected between 2001–2015 using the fusion/HR-ICPMS method developed by the OFP for small sample analysis (Conte et al., 2019; Huang et al., 2007). The lithogenic fraction of the total Si is calculated from the Al concentration, assuming the average lithogenic composition of deep pelagic North Atlantic sediments (8.4 % Al and 25.0 % Si, Li and Schoonmaker, 2003). Biogenic silica is estimated as the difference between total and lithogenic Si, and converted to opal using an estimated opal composition of $\text{SiO}_2 \cdot 0.4\text{H}_2\text{O}$ (Mortlock and Froelich, 1989). The opal percentage of a subset of 3200 m samples analyzed both by the fusion/HR-ICPMS method and by standard methods based upon biogenic silica dissolution in 2 M Na_2CO_3 (Jickells et al., 1998) is in close agreement (11.3 % and 11.4 %, respectively).

The sum of the masses of the individual bulk components analyzed using the above methods is in close agreement with the total sample mass measured gravimetrically, validating the analytical methods. For samples collected between 2001–2015, the mass balance is $100.0 \pm 9.6 \%$ for 500 m samples, $99.9 \pm 5.1 \%$ for 1500 m samples and $104.4 \pm 5.9 \%$ for 3200 m samples (Conte et al., 2019).

3.2.8. Statistical analyses

The software package XLSTAT (Addinsoft) was used for all statistical analyses and to model data frequency distributions.

The probability density function of the Weibull distribution is

$$f(x) = \frac{\alpha}{\beta^\alpha} x^{\alpha-1} e^{-\left(\frac{x}{\beta}\right)^\alpha} \quad (1)$$

where x , α and β are > 0 and α and β are the shape and scale parameters, respectively. As the Weibull shape parameter α approaches one the distribution approaches an exponential distribution ($\alpha = 1$).

The probability density function of the Gamma distribution is

$$f(x) = \frac{\beta^\alpha}{\Gamma(\alpha)} x^{\alpha-1} e^{-\beta x} \quad (2)$$

where α and β are the shape and scale parameters, respectively, and $\Gamma(\alpha)$ is the Gamma function. The shape parameter α is a measure of the distribution skewness. As α increases, the distribution evolves from a skewed exponential distribution ($\alpha = 1$) to a more symmetric bell-shaped distribution.

The Q-range (Conte et al., 2019) is defined as the difference between the 25 % and 75 % quartiles normalized to the median,

$$Q\text{-range} = \frac{(Q3 - Q1)}{\text{median}} \quad (3)$$

where Q1 and Q3 are the 25 % and 75 % quartiles, respectively. The Q-range is expressed as a percentage.

4. Results

4.1. Mass flux time series

The mass flux time-series at 500 m, 1500 m and 3200 m depths is shown in Fig. 2. The temporal coverage of the flux is 91 % at 3200 m (1978–2022), 82 % at 1500 m (1984–2022), and 77 % at 500 m (1984–2022).

Conte et al. (2019), Conte et al. (2001) and Conte and Weber (2014) have previously summarized the basic features of the OFP mass flux and its composition. Mass flux exhibits strong variability on time scales ranging from weeks to decades. The seasonal cycle exhibits a strong Feb/Mar flux maximum corresponding to the seasonal productivity maximum in overlying waters. A smaller Dec/Jan flux maximum occurs in some, but not all, years. Pronounced flux variability is observed on less than seasonal time scales and episodic, high flux events occur

throughout the record. While there is significant interannual variability, no long-term trends are apparent.

In Sept 2003 and again in Sept 2010, the OFP traps intercepted massive plumes of detrital carbonate sediments resuspended from the Bermuda Pedestal by hurricane-generated internal waves and advected > 50 km offshore to the OFP site (shown by the yellow arrows in Fig. 2). These events, described in Pedrosa-Pàmies et al. (2025), are excluded from the data analyses as they are atypical of the open ocean particle flux.

4.2. Flux frequency distributions

Flux frequency distributions were computed for samples collected at biweekly resolution (1989–2022) (Fig. 3). Flux frequency distributions at all depths are asymmetric with the tail of the distribution more pronounced to the right (right-skewed). Fluxes at 500 m depth exhibit extreme skewness. Flux skewness decreases significantly between 500 m and 1500 m depths and then more modestly between 1500 m and 3200 m depths.

Theoretical probability distributions were fit to the observed flux frequency distributions (shown by the solid curved red lines in Fig. 3). At 500 m depth, the frequency distributions of the mass flux and carbonate flux (i.e. carbonate fraction of the total mass flux) are best modeled by a two-parameter Weibull probability distribution. The Weibull shape parameters of both the mass and carbonate flux models (1.094 and 0.978, respectively) are close to one, indicating that their flux frequency distributions are nearly exponential. The frequency distribution of the C_{org} flux (i.e. the C_{org} fraction of the total mass flux) is best modeled by the closely related Gamma probability distribution. The Gamma shape parameter of the C_{org} flux model is 1.494, reflecting its more

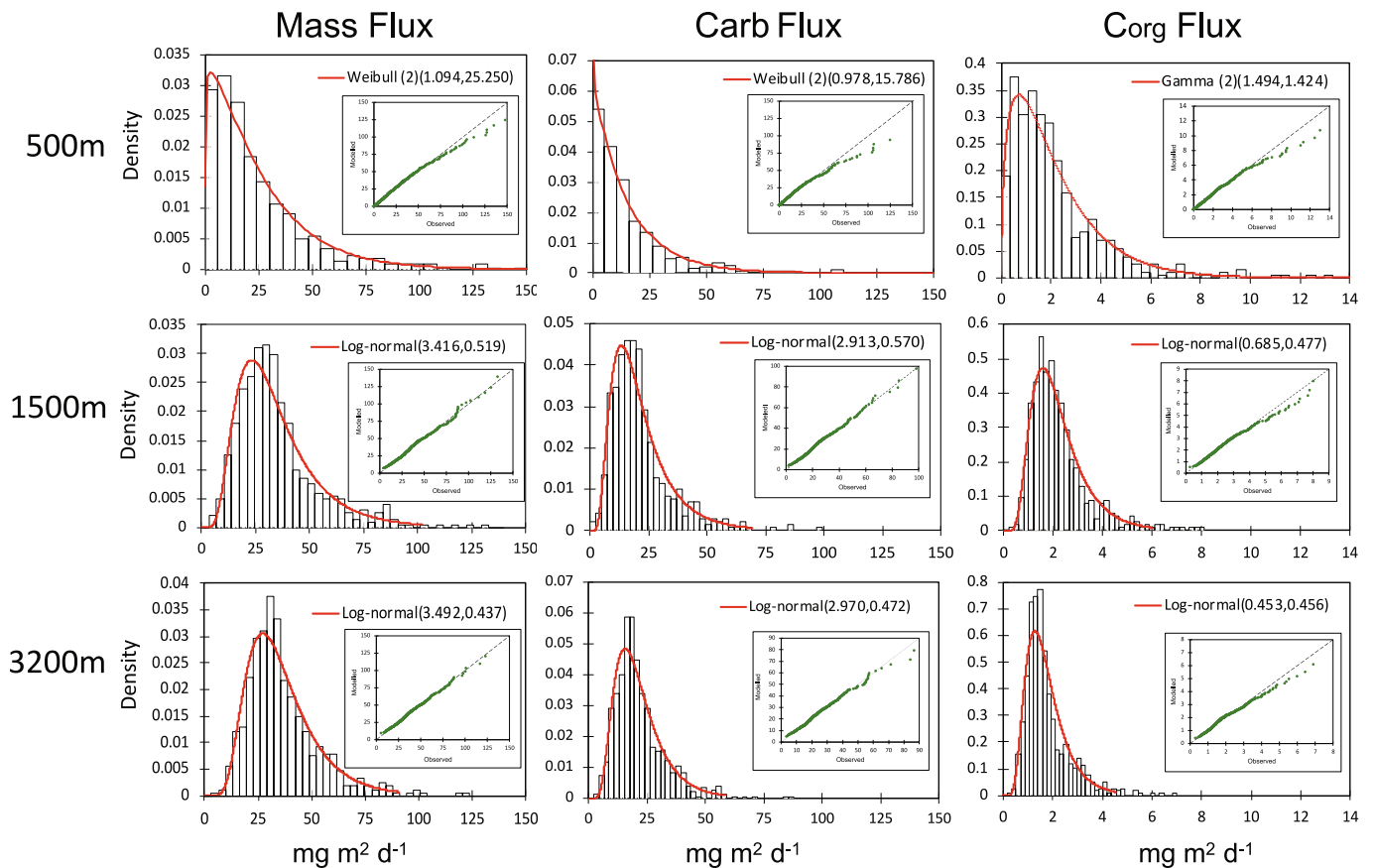


Fig. 3. Frequency distributions for mass, carbonate (Carb) and organic carbon (C_{org}) fluxes at the three trap depths. The empirically modeled distribution is shown by the red line. Model parameters are shown for each plot: lognormal (μ , σ), Gamma (k , β) or Weibull (β , γ). The fit of each model is shown by the probability plot inset. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 1

Summary statistics of total mass flux, fluxes of major components and their percent contribution to the total mass flux, and flux ratios. All fluxes are in $\text{mg m}^{-2} \text{d}^{-1}$. Mass, carbonate and organic matter fluxes are averaged over the entire record; opal and lithogenic fluxes are averaged over the 15-year period (2001–2015) when full elemental data are available. Flux statistics at 500 m and 1500 m depths omit the 1992–1995 period when fluxes were not quantitative (see text). Q-range is the difference between the 25 % and 75 % quartiles normalized to the median and expressed as a percentage.

(a) Mass and bulk component fluxes							
Depth		500 m		1500 m		3200 m	
		Flux	%	Flux	%	Flux	%
Mass Flux	median	16.8		32.1		33.4	
	Mean	27.2		34.8		36.8	
	stdev	24.9		19.2		16.5	
	25 % quartile	9.6		22.1		25.7	
	75 % quartile	37.40		41.1		44.3	
	Q-range	165 %		59 %		56 %	
CaCO_3	median	11.4	62.7	19.5	62.0	19.6	58.7
	Mean	18.1	60.6	21.8	61.0	22.0	59.0
	stdev	19.4	14.5	13.3	7.0	10.8	4.7
	25 % quartile	5.3	50.5	13.1	56.8	15.1	55.9
	75 % quartile	22.8	71.0	25.3	65.5	26.8	62.0
	Q-range	154 %	33 %	63 %	14 %	60 %	10 %
Organic Matter	median	4.1	20.5	4.7	14.4	3.5	10.9
	Mean	5.5	21.6	5.1	15.0	4.1	11.1
	stdev	4.8	7.8	2.7	4.6	2.1	1.6
	25 % quartile	2.3	16.3	3.3	12.9	2.8	10.0
	75 % quartile	7.4	24.8	6.0	16.0	4.8	11.7
	Q-range	124 %	41 %	57 %	22 %	57 %	16 %
Opal	median	1.4	6.5	3.0	9.5	4.6	10.9
	Mean	1.8	6.8	3.9	10.0	5.2	11.3
	stdev	1.7	3.4	2.7	3.5	2.9	3.1
	25 % quartile	0.6	4.4	2.3	7.7	3.5	9.2
	75 % quartile	2.3	8.7	4.8	11.6	6.4	13.1
	Q-range	121 %	66 %	83 %	41 %	63 %	36 %
Lithogenic	median	1.0	5.0	4.2	12.3	6.3	18.2
	Mean	1.3	6.1	4.4	12.6	6.4	18.1
	stdev	1.1	4.1	1.9	3.9	2.2	4.3
	25 % quartile	0.6	2.8	3.0	9.7	4.9	15.3
	75 % quartile	1.6	8.5	5.4	14.9	7.8	21.1
	Q-range	100 %	114 %	57 %	42 %	46 %	32 %
(b) Flux ratios							
Depth		500 m		1500 m		3200 m	
molar C/N	median	9.5		9.0		9.4	
	mean	9.8		9.1		9.5	
	stdev	1.6		1.0		0.9	
	25% quartile	8.8		8.5		9.0	
	75% quartile	10.7		9.6		9.9	
	Q-range	20%		12%		10%	
PIC/POC	median	0.9		1.2		1.5	
	mean	1.0		1.2		1.5	
	stdev	0.4		0.3		0.2	
	25% quartile	0.7		1.0		1.4	
	75% quartile	1.2		1.4		1.6	
	Q-range	56%		33%		13%	

symmetrical flux frequency distribution as compared to the carbonate flux.

The fit of the modeled distributions is shown by the probability plots of the observed versus modeled cumulative data (inset plots within each graph). Both carbonate and C_{org} fluxes show close agreement between the observed and modeled data at lower flux magnitudes. However, both probability models underestimate the frequency of occurrence of the highest values, a pattern commonly termed a “heavy-tailed” distribution (Bryson, 1974).

Between 500 m and 1500 m depths, flux frequency distributions evolve to follow less skewed lognormal distributions. At 3200 m depth,

fluxes continue to follow a lognormal distribution but the flux frequency distribution is more peaked and symmetric than at 1500 m depth. Carbonate fluxes at both 1500 m and 3200 m depths show close agreement with the modeled lognormal probability distribution, whereas the C_{org} flux continues to show a “heavy-tailed” distribution.

4.3. Flux magnitude and composition

The statistics of the flux climatology at the three depths are summarized in Table 1. The average mass flux over this period is $27.2 \pm 24.9 \text{ mg m}^{-2} \text{d}^{-1}$ at 500 m depth, $34.8 \pm 19.2 \text{ mg m}^{-2} \text{d}^{-1}$ at 1500 m

depth, and $36.8 \pm 16.5 \text{ mg m}^{-2} \text{ d}^{-1}$ at 3200 m depth (Table 1a). Carbonate is the most abundant flux component, comprising $\sim 60\%$ of the mass at all depths. The carbonate flux averages $\sim 22.0 \text{ mg m}^{-2} \text{ d}^{-1}$ and is quasi-conservative with depth. Organic matter (OM) comprises $\sim 22\%$, $\sim 15\%$ and $\sim 11\%$ of the mass flux at 500 m, 1500 m and 3200 m depths, respectively. The OM flux decreases from $5.5 \text{ mg m}^{-2} \text{ d}^{-1}$ at 500 m depth to $4.1 \text{ mg m}^{-2} \text{ d}^{-1}$ at 3200 m depth, reflecting the net loss of C_{org} due to water column remineralization. This results in a 50 % increase in the PIC/POC ratio between 500 m (1.0) and 3200 m (1.5) depths (Table 1b).

Mean molar C/N ratio decreases from 9.8 to 9.1 between 500 m and 1500 m depths and then increases to 9.5 in the 3200 m flux. These trends in the molar C/N flux ratio do not indicate a higher net loss of N relative to C_{org} in the deep flux, as would be expected from a higher lability of organic N relative to C within decaying organic matter.

A striking feature of the OFP record is the large increase in the lithogenic flux (i.e. the lithogenic fraction of the total mass flux) with depth (Table 1a). The mean lithogenic flux increases three-fold from $1.3 \text{ mg m}^{-2} \text{ d}^{-1}$ to $4.4 \text{ mg m}^{-2} \text{ d}^{-1}$ between 500 m and 1500 m depths and a further 45 % to $6.4 \text{ mg m}^{-2} \text{ d}^{-1}$ at 3200 m depth, evidencing a significant laterally advected source of lithogenic material in the deep mesopelagic water column. This results in a three-fold increase in the lithogenic percentage of the particle flux, from 6 % at 500 m to 18 % at 3200 m depth.

Opal flux also increases with depth, suggesting that there is also a significant laterally advected source of the opal material in the deep water column. Mean opal flux doubles between 500 m ($1.8 \text{ mg m}^{-2} \text{ d}^{-1}$) and 1500 m ($3.9 \text{ mg m}^{-2} \text{ d}^{-1}$) depths and increases by another 33 % between 1500 m and 3200 m ($5.2 \text{ mg m}^{-2} \text{ d}^{-1}$) depths.

To quantitatively evaluate flux variability, we computed the data Q-range (Section 3.2.7), a simple nonparametric measure of variability that is insensitive to extreme values. The decrease in Q-ranges with depth evidences a large dampening of flux variability within the lower mesopelagic zone, and a smaller dampening in the bathypelagic zone. The Q-range of the total mass flux decreases 3-fold between 500 m (165 %) and 1500 m (59 %) depths, and then slightly between 1500 m and 3200 m (56 %) depths. The Q-range of the carbonate flux decreases by a factor of 2.4 between 500 m (154 %) and 1500 m (63 %) depths, then

slightly between 1500 m and 3200 m (60 %) depths. The Q-ranges of the OM, lithogenic and opal fluxes show similar patterns. Between 500 m and 1500 m depths, the Q-range of the OM flux decreases from 124 % to 57 %, the Q-range of the lithogenic flux decreases from 100 % to 57 %, and the Q-range of the opal flux decreases from 121 % to 83 %. These depth patterns in the Q-ranges mirror the large changes observed in flux frequency distributions between 500 m and 1500 m depths and the much smaller changes in flux frequency distributions between 1500 m and 3200 m depths (Fig. 3).

Compositional heterogeneity decreases as flux variability is dampened with depth. The Q-range of the carbonate fraction decreases by 57 % between 500 m (33 %) and 1500 m (14 %) depths, and the Q-range of the OM fraction decreases by 46 % between 500 m (41 %) and 1500 m (22 %) depths. Q-ranges indicate larger compositional variability in the opal and lithogenic fractions. At 500 m depth, the Q-range of the opal fraction is 66 % and the Q-range of the lithogenic fraction is 114 %, the highest of all bulk components. By 1500 m depth, Q-ranges of the opal and lithogenic fractions decrease to 41 % and 42 %, respectively, but are still 2–3 times higher than the Q-ranges of the OM or carbonate flux components.

Between 1500 m and 3200 m depths, Q-ranges of all components continues to decrease indicating further homogenization of flux composition as particles are recycled within the bathypelagic zone. However, while Q-ranges of carbonate, OM and lithogenic percentages decline $\sim 25\%$ between 1500 m and 3200 m, the Q-range of opal percentage declines by only 12 %.

Homogenization of flux composition with increasing depth is also apparent in the Q-ranges of molar C/N and PIC/POC ratios, which decrease by roughly 50 % between 500 m (20 % and 56 %, respectively) and 1500 m (12 % and 33 %, respectively) depths (Table 1b). While the Q-range of the PIC/POC ratio is $\sim 3\times$ higher than the C/N ratio at 500 m and 1500 m depths, it is of similar magnitude at 3200 m depth.

4.4. Covariation and temporal coherence of flux components

Fluxes of bulk components strongly covary at all depths (Table 2). Flux correlations increase with depth for all components except opal, providing further evidence of flux homogenization with depth. The high correlations between the lithogenic and biogenic components at 1500 m and 3200 m depths indicate a strong covariation of lithogenic flux with the biogenic flux in the deep water column.

Opal correlations exhibit noteworthy differences from other biogenic components. At 500 m depth, opal correlations range from 0.67 (N) to 0.84 (CaCO_3), but at 1500 m depth the correlations are lower, ranging from 0.66 (N) to 0.70 (CaCO_3). This indicates a relative decoupling between opal flux and both carbonate and OM fluxes within the lower mesopelagic. In contrast, the correlation between the opal and lithogenic flux increases between 500 m and 1500 m depths, from 0.56 to 0.70, respectively, and by 3200 m depth, opal is more highly correlated with the lithogenic fraction (0.71) than with any other component. The strong correlation between the opal and lithogenic fluxes in the deep water column is notable considering evidence (Section 4.3) that a significant fraction of both the opal and lithogenic flux in the deep water column is sourced from lateral advection.

In addition to the strong covariation among individual flux components, there is a strong temporal coherence in fluxes at the three trap depths (Fig. 2). To examine this temporal coherence quantitatively, we computed cross-correlations of paired depth records (e.g. 500 m/1500 m, 500 m/3200 m, 1500 m/3200 m) for biweekly resolved flux data. Analyses were run for the case of no sample lag in the deeper flux (i.e., a 0-d ay temporal lag between depths), one sample lag ($\sim$ a 15 day temporal lag) and a two sample lag ($\sim$ a 29 day temporal lag) (Fig. 4). Results provide no evidence of a temporal lag in bulk component fluxes for any depth pair within the two-week sampling resolution of the data. The highest correlation for all depth pairs is the case of no temporal lag, even between the 500 m/3200 m depth pair. Interestingly, the cross-

Table 2
Correlation coefficients of individual flux components at 500, 1500 and 3200 m depths.

(a) 500 m flux						
	Mass	CaCO_3	C_{org}	Nitrogen	Opal	Lithogenic
Mass	1	0.98	0.89	0.83	0.83	0.57
CaCO_3	–	1	0.82	0.76	0.84	0.54
C_{org}	–	–	1	0.92	0.72	0.52
Nitrogen	–	–	–	1	0.67	0.47
Opal	–	–	–	–	1	0.56
Lithogenic	–	–	–	–	–	1
(b) 1500 m flux						
	Mass	CaCO_3	C_{org}	Nitrogen	Opal	Lithogenic
Mass	1	0.98	0.92	0.90	0.77	0.69
CaCO_3	–	1	0.88	0.86	0.70	0.63
C_{org}	–	–	1	0.96	0.73	0.62
Nitrogen	–	–	–	1	0.66	0.57
Opal	–	–	–	–	1	0.70
Lithogenic	–	–	–	–	–	1
(c) 3200 m flux						
	Mass	CaCO_3	C_{org}	Nitrogen	Opal	Lithogenic
Mass	1	0.98	0.96	0.94	0.81	0.70
CaCO_3	–	1	0.94	0.92	0.73	0.60
C_{org}	–	–	1	0.97	0.75	0.62
Nitrogen	–	–	–	1	0.73	0.61
Opal	–	–	–	–	1	0.71
Lithogenic	–	–	–	–	–	1

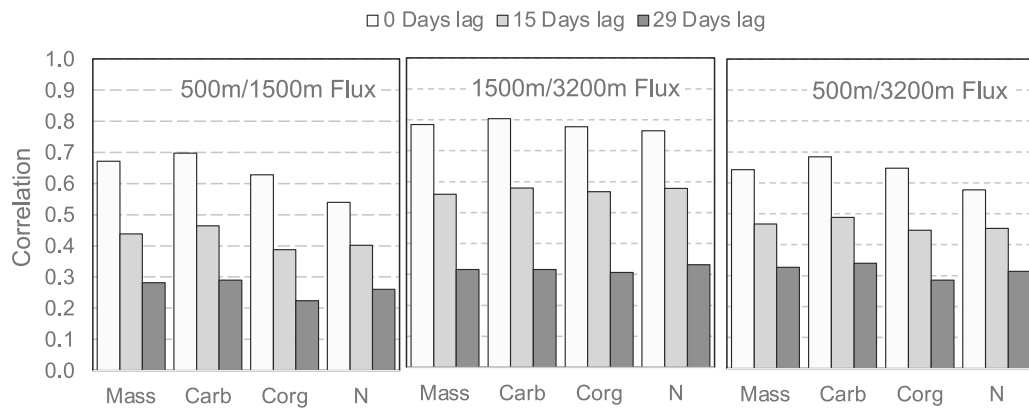


Fig. 4. Cross-correlations of mass, carbonate, C_{org} and nitrogen fluxes for no temporal lag (0 days), one cup lag (~ 15 days) and two cup lag (~ 29 days).

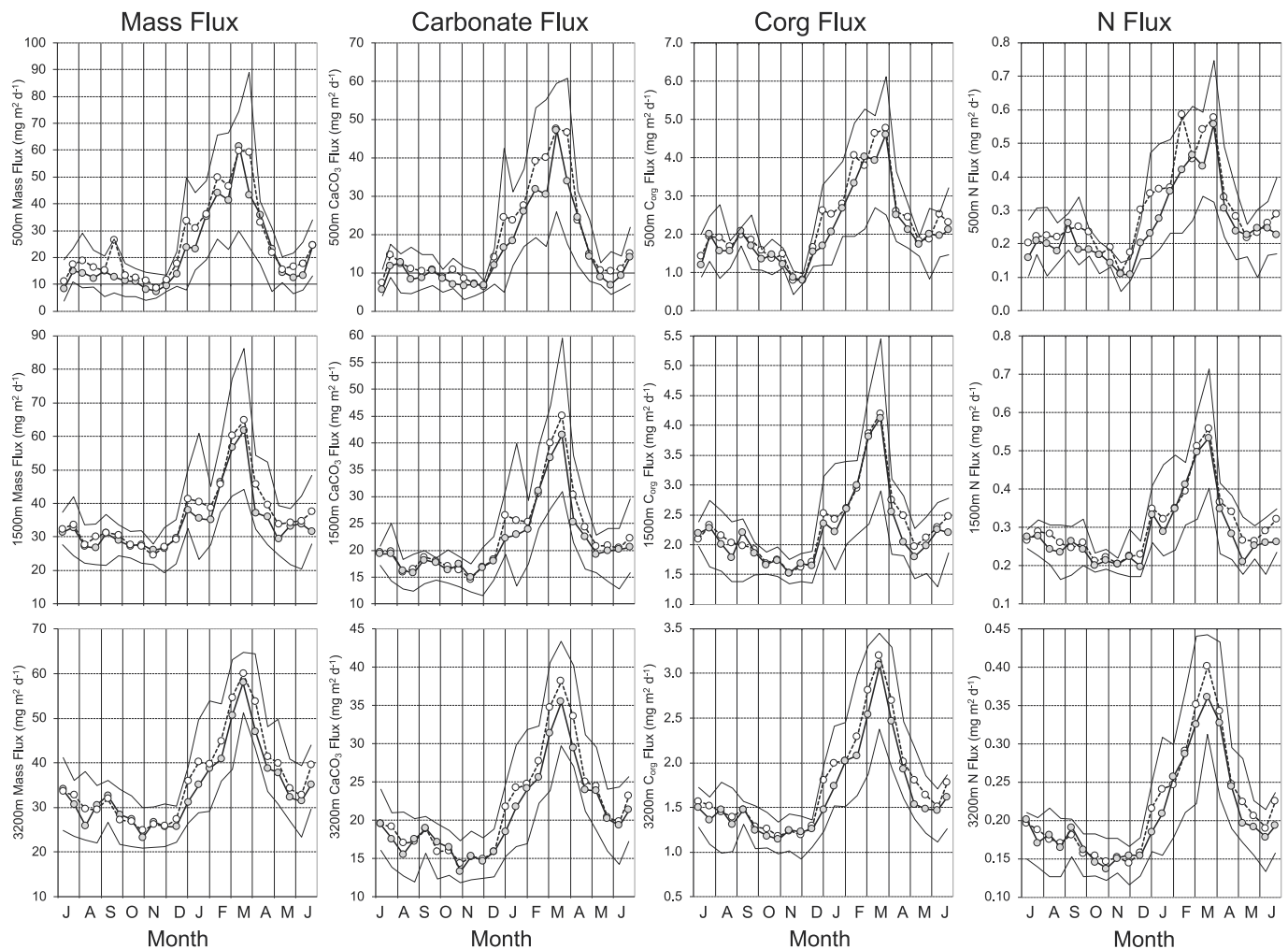


Fig. 5. The seasonal cycle in mass and bulk component fluxes and in molar C/N and PIC/POC ratios. The data are plotted July 1 – June 30, using the midpoint of the sampling interval, so as to better show the winter-spring high flux period. Median values are shown by filled circles, mean values by the open circles, and the upper/lower quartiles by the thin lines. Mass, carbonate, organic carbon and nitrogen fluxes and the molar C/N and PIC/POC ratios are averages of the biweekly resolved flux data from 1989 to 2021; opal and lithogenic flux average data from 2001 to 2015.

correlations for the depth pairs of 500/1500 m and 500/3200 m are comparable, despite the 1700 m difference in depth ranges. The slightly higher correlation for the 1500/3200 m depth pair than the 500/1500 m depth pair indicates that flux temporal variability is more strongly correlated in the deeper water column.

4.5. Flux seasonality

The seasonal cycles of total mass flux and the individual flux components and flux ratios are shown in Fig. 5. (The data are plotted July–July to better show the winter/spring flux peak.) To construct the

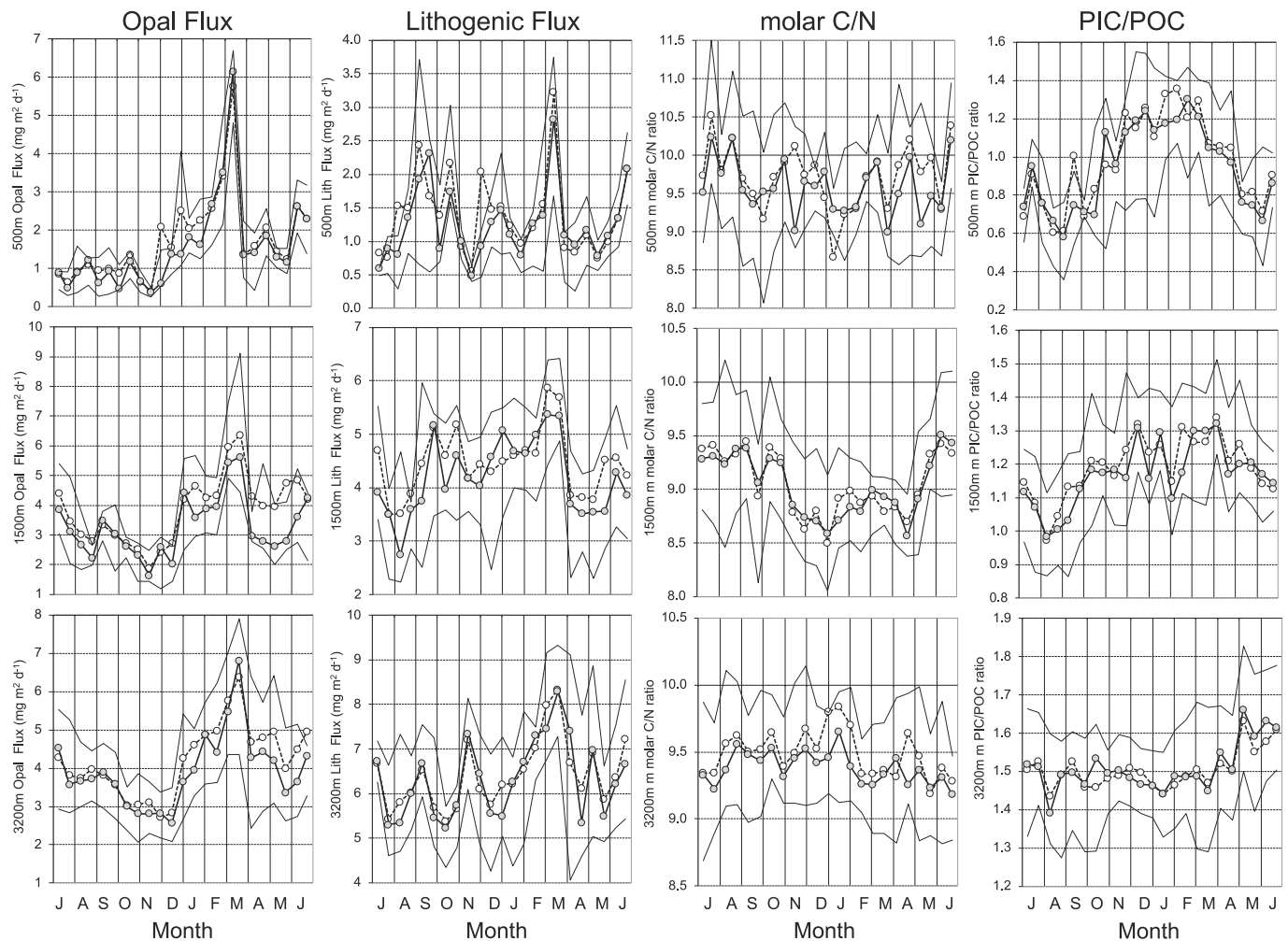


Fig. 5. (continued).

figures, the biweekly resolved flux data were sorted into two-week time bins using the mid sampling date (i.e., day of year) of the sample collection. Flux statistics (mean, median, quartiles) were then calculated for each biweekly bin. Carbonate, organic carbon and nitrogen fluxes are averaged for the 1989–2022 time period, and opal and lithogenic fluxes are averaged for the 2001–2015 time period.

Consistent with the cross-correlation results (Fig. 4), there is no temporal lag with depth in the flux seasonal cycle for any bulk component. Mass flux begins to increase in late December and reaches a maximum in March ($\sim 60 \text{ mg m}^{-2} \text{ d}^{-1}$ at all three trap depths). A secondary flux maximum is observed in late winter (December/January) in some but not all years (Fig. 2), as reflected in the large spread of the 25 % and 75 % quartiles. The mass flux decreases rapidly after the March flux peak and remains low throughout the summer and fall, reaching a seasonal minimum in November.

The seasonal cycle of bulk component fluxes has been previously described (Conte et al., 2001; Conte et al., 2019). Carbonate flux also exhibits a strong flux maximum in March and then decreases to a minimum throughout the summer and fall. A significant observation is that during the minimum flux season in summer and fall the carbonate fluxes at 1500 m and 3200 m depths are twice that at 500 m depth. Assuming mass balance over the annual cycle, this indicates a significant fraction of the deep carbonate flux during low productivity periods is generated via repackaging of suspended carbonate particles in the water column.

C_{org} and nitrogen fluxes show similar seasonality. The magnitude of the March maximum in the C_{org} flux declines by $\sim 40\%$ between 500 m and 3200 m depths, indicating its strong remineralization within the

water column. In contrast to the carbonate flux, C_{org} and N fluxes both show a small increase during the late spring. Subsequently, C_{org} and N fluxes rapidly decline over the summer and fall and reach a seasonal minimum in November.

Opal flux exhibits large seasonal variability. At 500 m depth, there is a large, short-lived opal flux maximum in early March ($\sim 6 \text{ mg m}^{-2} \text{ d}^{-1}$) followed by a rapid decline. As observed in the C_{org} and nitrogen fluxes, there is a small increase in the opal flux in June ($\sim 2.5 \text{ mg m}^{-2} \text{ d}^{-1}$) followed by a two-fold decrease to $\sim 1.0 \text{ mg m}^{-2} \text{ d}^{-1}$, about 15 % of the peak flux, throughout the summer and fall. Opal flux begins to increase in December and a secondary flux maximum is clearly apparent in late December-January. At 1500 m depth, there is a strong opal flux maximum in March and a smaller secondary flux maximum in June ($\sim 4.0 \text{ mg m}^{-2} \text{ d}^{-1}$), although the large quartile spread indicates pronounced interannual variability in the occurrence of the June peak. Unlike at 500 m depth, the opal flux at 1500 m depth declines only slowly during the summer and fall, reaching a minimum in November ($2.0 \text{ mg m}^{-2} \text{ d}^{-1}$). The 1500 m opal flux increases abruptly in late December to a secondary flux maximum in late December-January. At 3200 m depth, the opal flux follows a similar seasonal pattern but the minimum fluxes in summer and fall are 30–50 % higher than at 1500 m depth, indicating a significant fraction of the deep opal flux during low productivity periods is also generated via repackaging of suspended particles in the water column.

Lithogenic fluxes also exhibit a seasonal cycle, but there are notable differences in seasonal patterns at 500 m, 1500 m and 3200 m depths. At 500 m depth, there is a strong, short-lived maximum in the lithogenic

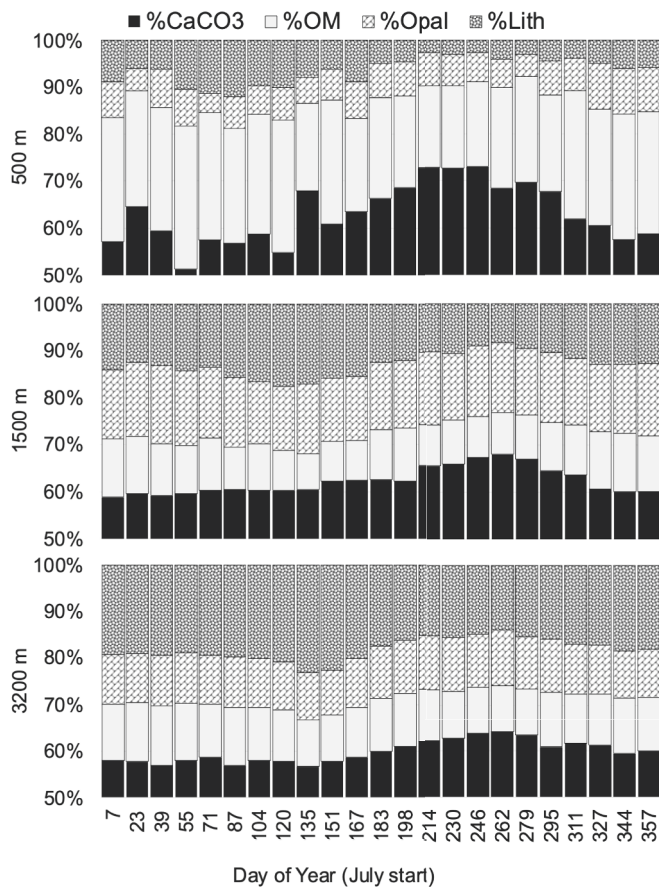


Fig. 6. Seasonal evolution in bulk composition of the mass flux at 500, 1500 and 3200 m depths.

flux in March and evidence of a small flux increase in June, similar to that observed for biogenic components. However, unlike biogenic fluxes, the lithogenic flux at 500 m depth also exhibits a broad secondary flux maximum between August and October, which coincides with the season of Saharan dust deposition (Arimoto et al., 1995). The large quartile spread of the late summer – fall maximum indicates significant interannual variability in its occurrence, which is consistent with the observed interannual variability in Saharan dust transport to the Bermuda region (ibid).

At 1500 m depth, the lithogenic flux gradually increases during the winter to reach a maximum in March, and then declines by about 50 % during April. A secondary summer-fall maximum is also apparent during the late summer and fall, but this peak occurs about one month later than at 500 m depth and is much reduced in magnitude. The large quartile spread indicates significant interannual variability in the overall seasonal flux pattern at 1500 m depth.

The lithogenic flux at 3200 m depth exceeds that at 1500 m depth throughout the year, indicating strong net aggregation of suspended lithogenic particles in the deep water column. Interestingly, the spring

flux maximum is more clearly defined at 3200 m depth than at 1500 m depth. No secondary late summer-fall secondary flux maximum in the lithogenic flux is apparent at 3200 m depth.

Seasonal patterns in molar C/N and PIC/POC ratios evolve with depth. At 500 m depth, molar C/N ranges between 9.0–10.4 and is highly variable, obscuring any seasonal patterns. By 1500 m depth, molar C/N exhibits a clear seasonal pattern, with highest values from June to October (9.0–9.5) and lowest values from November to May (8.6–9.0). At 3200 m depth, no seasonal cycle is apparent except that molar C/N in spring through early summer (9.2–9.3) is about 0.2 lower than during the rest of the year.

In contrast to the molar C/N ratio, the PIC/POC ratio exhibits a clear seasonal pattern at both 500 m and 1500 m depths. At 500 m depth, the PIC/POC ratio is highest during the spring flux maximum (~1.2) and lowest during the summer-fall flux minimum (~0.60–0.65). A similar seasonal pattern is observed at 1500 m depth, but the seasonal amplitude is about 50 % lower (mean range 1.0–1.3). By 3200 m depth, the PIC/POC ratio averages ~1.5 but no seasonal cycle is apparent except that the PIC/POC ratio is ~0.1 higher in the late spring.

The seasonal cycle in bulk flux composition is shown in Fig. 6. Seasonality of the bulk flux composition is primarily driven by the carbonate flux. At 500 m depth, the carbonate fraction of the flux begins to rise in the early winter and reaches a maximum during the Jan – Mar bloom period (days 214–246). As the carbonate fraction declines post-bloom the lithogenic and OM fractions begin to increase. A similar pattern is observed at 1500 m and 3200 m depths but the magnitude of the seasonality becomes increasingly dampened with depth. There is a clear maximum in the percent contribution of opal in the 1500 m flux but no seasonal patterns.

The Flux Stability Index (FSI, Lampitt and Antia, 1997), defined as the number of days required to accumulate 50 % of the total flux over the annual cycle, is a useful statistic of flux seasonal variability. We computed the climatological FSI for the total mass flux and bulk flux components using the biweekly resolved flux data (1989–2022) by sorting the data at each depth by magnitude, and then calculating the fraction of the total number of sampling days needed to accumulate 50

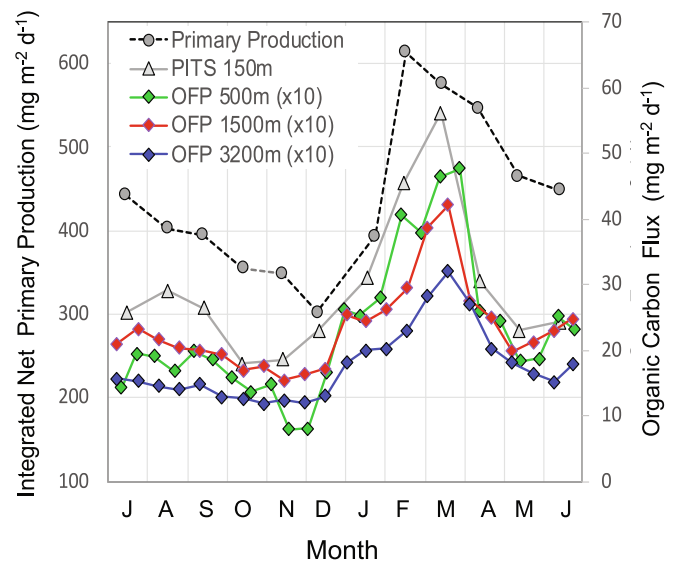


Fig. 7. The seasonal cycle in overlying integrated primary production (intPP) and C_{org} export flux at 150 m depth (EF) at the BATS site compared with the seasonal cycle of deep C_{org} fluxes measured by the OFP moored traps. The data are plotted July 1 – June 30 to better show the flux peak. Both the BATS and OFP data have been averaged over the same thirty-year period (Nov 1989 – Dec 2021). The BATS data are plotted using the average day of year of monthly binned data. The OFP data are plotted using the average midpoint of the sample collection interval of biweekly binned data.

Table 3

Flux Stability Indices (FSI, in days) for the annual cycle. Data are the climatological averages for the 1989–2021 time period.

	500 m	1500 m	3200 m
Mass Flux	68	112	122
Carbonate Flux	59	105	116
C_{org} Flux	60	113	114
N flux	65	108	113
Opal flux	72	99	116
Lithogenic Flux	81	125	133

% of the cumulative flux. We multiplied this fraction by 365 to give the climatological FSI in days (Table 3).

At 500 m depth, FSIs of carbonate, C_{org} and N range from 59 to 65 days, reflecting the high seasonality and pronounced spring flux maximum (Fig. 5). The FSI of opal (72 days) is 7–13 days longer, reflecting the more extended period of elevated opal flux throughout the late spring. The lithogenic flux has the longest FSI (81 days), reflecting both its lower seasonality and the extended secondary flux maximum in summer-fall.

FSIs for all components increase with depth, reflecting the dampening of flux variability and flux seasonality (c.f. Table 1). At 1500 m depth, FSIs of carbonate, C_{org} and N fluxes (105, 113, 108 days, respectively) are 70–90 % longer than at 500 m depth, and FSIs of the opal and lithogenic fluxes (99 and 125 days, respectively) are 40–50 % longer. Between 1500 m and 3200 m depths, there is a further 10–20 % increase in the FSIs of the carbonate, opal and lithogenic fluxes while the FSIs of C_{org} and N fluxes remain constant.

4.6. Linkage of the deep carbon flux with overlying primary production and surface export flux

The seasonal cycle in the deep flux and its linkage to overlying primary production was an early discovery of the OFP (Deuser, 1986). Here we quantify deep-upper ocean linkages over the three decade time period (1990–2021) in which the nearby BATS time series made coincident monthly measurements of integrated net primary production (intPP) and C_{org} export flux at 150 m depth (EF, surface export flux) using 3-day drifting trap deployments (Steinberg et al., 2001). To enable direct comparisons between the data sets, we binned the BATS data into

~ monthly time bins using the day of year of the measurement, and then computed the average day of year, intPP and 150 m C_{org} export flux for each monthly bin.

Fig. 7 compares the 30 year integrated seasonal cycle of the deep C_{org} fluxes measured by the OFP traps with the BATS monthly data of intPP and EF. The intPP rises sharply in January, reaches a seasonal maximum in mid-February ($\sim 610 \text{ mg m}^{-2} \text{ d}^{-1}$) and then slowly declines over the rest of the year to reach a seasonal minimum in early December ($\sim 300 \text{ mg m}^{-2} \text{ d}^{-1}$). The EF rises concurrently with intPP but reaches its seasonal maximum ($\sim 55 \text{ mg m}^{-2} \text{ d}^{-1}$) in March, one month later than intPP.

As seen in Fig. 5, the deep particle flux begins to increase in mid Dec, prior to the Jan increase in intPP observed in the BATS data. This deep flux increase in mid Dec- early Jan appears to be associated with episodic fluxes that are forced by transient upper ocean mixing conditions during the season of mixed layer erosion (e.g. Conte et al., 1998). As synoptic-scale weather systems pass through the area, alternating episodes of negative and positive surface heat loss and wind stress results in alternating episodes of destratification and stratification leading to a sequence of nutrient upwelling during destratification, which supports enhanced phytoplankton production during the next quiescent period of temporary mixed layer stratification. This material is subsequently downmixed during the next period of destratification, resulting in an episodic flux of fresh phytodetrital material to depth.

Within the ~ monthly resolution of the BATS data, the deep C_{org} fluxes are temporally coherent with the EF. The main difference, weakly suggested by the data, is that the seasonal minimum in EF slightly precedes the seasonal minimum in the deep C_{org} fluxes.

Table 4(a) gives the annual and seasonal averages of intPP, EF, the

Table 4

(a) Annual and seasonal averages of depth integrated net primary production (intPP) and the 150 m C_{org} flux (EF) measured by the BATS time-series compared with deep C_{org} fluxes measured by the OFP time series. All data are averaged over the BATS time series record (Nov 1989–Dec 2021) and are in $\text{mg m}^{-2} \text{ d}^{-1}$. The Q-ranges of the intPP and 150 m C_{org} export flux are shown in parentheses. C_{org} export flux efficiency (e-ratio) is defined here as the percentage of intPP exported by sinking particles to 150 m depth. Flux transfer efficiency (TE) is defined as the percentage of the 150 m C_{org} export flux transferred to deep waters, shown here for 1500 m and 3200 m depth horizons.

	Annual	winter (JFM)	spring (AMJ)	summer (JAS)	fall (OND)
Integrated PP	436.3	534.2	472.4	408.8	330.1
(Q-range)	(52 %)	(74 %)	(46 %)	(44 %)	(46 %)
150 m C_{org} flux	29.0	43.2	25.8	27.4	19.4
(Q-range)	(62 %)	(79 %)	(55 %)	(60 %)	(54 %)
OFP C_{org} flux (500 m)	2.4	3.6	2.3	1.8	1.3
OFP C_{org} flux (1500 m)	2.2	3.1	2.3	2.1	1.7
OFP C_{org} flux (3200 m)	1.8	2.4	1.9	1.5	1.2
C_{org} export flux (% of intPP)	6.6 %	8.1 %	5.5 %	6.7 %	5.9 %
% of intPP reaching 500 m	0.6 %	0.7 %	0.5 %	0.4 %	0.4 %
% of intPP reaching 1500 m	0.5 %	0.6 %	0.5 %	0.5 %	0.5 %
% of intPP reaching 3200 m	0.4 %	0.4 %	0.4 %	0.4 %	0.4 %
Transfer efficiency (1500 m)	7.6 %	7.2 %	9.1 %	7.6 %	8.5 %
Transfer efficiency (3200 m)	6.2 %	5.6 %	7.4 %	5.5 %	6.2 %
Difference (1500–3200 m)	1.9 %	1.6 %	1.7 %	2.2 %	2.4 %

Table 4b

(b) As above, but an estimated upper limit for C_{org} advected with deep water transport of lithogenic material has been subtracted from the 1500 and 3200 m fluxes (see text). The 500 m flux has not been adjusted. Data are averages for the 2001–2015 period when lithogenic analyses were conducted.

	Annual	winter (JFM)	spring (AMJ)	summer (JAS)	fall (OND)
IntPP	451.1	553.2	506.5	422.2	322.6
150m C_{org} flux	27.8	46.4	24.3	23.6	16.9
OFP C_{org} flux (500 m)	—	—	—	—	—
OFP C_{org} flux (1500 m)	2.3	3.2	2.3	1.9	1.6
OFP C_{org} flux (3200 m)	1.6	2.3	1.7	1.3	1.1
C_{org} export flux (% of intPP)	6.2%	8.4%	4.8%	5.6%	5.3%
% of intPP reaching 500 m	—	—	—	—	—
% of intPP reaching 1500 m	0.5%	0.6%	0.5%	0.5%	0.5%
% of intPP reaching 3200 m	0.4%	0.4%	0.3%	0.3%	0.3%
Transfer efficiency (1500 m)	8.3%	7.0%	9.4%	8.3%	9.5%
Transfer efficiency (3200 m)	5.8%	4.9%	6.9%	5.3%	6.5%
Difference (1500 – 3200 m)	2.5%	2.0%	2.5%	2.9%	3.0%

deep C_{org} fluxes at 500, 1500 and 3200 m depths, and also the Q-ranges of the intPP and EF. The annual intPP averages $436.2 \text{ mg m}^{-2} \text{ d}^{-1}$ and ranges from a maximum of $534 \text{ mg m}^{-2} \text{ d}^{-1}$ in the winter (JFM) to a minimum of $330 \text{ mg m}^{-2} \text{ d}^{-1}$ in the fall (OND). During the peak flux period, the Q-range of intPP (74 %) is about 50 % larger than during the rest of the year (44–46 %), reflecting a large interannual variability in the magnitude of the spring bloom. The EF averages $29.0 \text{ mg m}^{-2} \text{ d}^{-1}$ annually and ranges from a maximum of $43.2 \text{ mg m}^{-2} \text{ d}^{-1}$ in the winter (JFM) to a minimum of $19.4 \text{ mg m}^{-2} \text{ d}^{-1}$ in the fall (OND). Consistent with intPP, the Q-range of the EF (79 %) is about 40 % higher in the winter (JFM) than the during the rest of the year (54–60 %). This high variability in overlying intPP and EF during the peak bloom period is mirrored by the high variability in the deep C_{org} fluxes (Fig. 5).

4.7. Carbon transfer efficiency within the water column

Three decades of concurrent OFP and BATS in-situ data allow for robust estimates of the climatological average depth penetration of carbon within the water column (Table 4a). Our estimates assume that the BATS PITS-style cylindrical traps average 100 % trapping efficiency. This assumption is supported by the lack of any consistent difference between calculated fluxes based on ^{234}Th profiles and ^{234}Th fluxes in cylindrical traps (Baker et al., 2020 and references therein), and by paired comparison between ^{234}Th -derived and trap fluxes ($n = 19$) at the BATS site, which yielded highly variable ^{234}Th -based estimates of trapping efficiencies but no consistent evidence of over/under trapping or seasonality in trapping efficiency (Buesseler et al., 2007). Similarly, direct comparison of BATS traps with neutrally buoyant traps (NBSTs) ($n = 77$) also find no evidence of consistent under/over sampling due to hydrodynamic biasing (Owens et al., 2013).

We also assume 100 % trapping efficiency for the OFP deep moored traps. This assumption is supported by the excellent agreement between the 3200 m trap flux and flux estimates based upon water column ^{230}Th profiles (Bacon et al. 1985), as well as the minimal impact of swimmers on trap fluxes at 1500 m and 3200 m depths (Appendix 2). We note that current and tilt data (Appendix 1) indicate a possibility of small hydrodynamic biasing and material losses due to swimmers in the 500 m trap (Appendix 2), as found for other traps sampling in mesopelagic

waters. Thus, the measured flux at 500 m depth may slightly underestimate the true flux.

C_{org} export flux efficiency, or e-ratio, is defined as percentage of the intPP that is exported to 150 m depth. The C_{org} export flux efficiency averages 6.6 % annually. About 0.6 % of the annual intPP is exported to 500 m depth, about 0.5 % to 1500 m depth, and about 0.4 % to 3200 m depth. As noted above, the efficiency of export to 500 m depth is a minimum estimate.

There is a clear seasonal trend in the percentage of overlying intPP transferred to depth which correlates with the seasonal cycle in intPP. The export flux efficiency during the season of maximum intPP (JFM, 8.1 %) is 47 % higher than during the spring post-bloom season (AMJ, 5.5 %) and 20–35 % higher than during the summer and fall when intPP is at a minimum.

A similar seasonality is observed in the percentage of intPP exported to deeper depths. At 500 m depth, 0.7 % of overlying intPP is exported during the season of maximum intPP (JFM) but only 0.4 % exported during the production minimum (JAS). Seasonality is much reduced but still apparent at 1500 m depth, with 0.6 % of intPP is exported during the season of maximum intPP (JFM) and 0.5 % during the rest of the year. By 3200 m depth, no seasonality is apparent.

Carbon flux transfer efficiency (TE), defined as the fraction of the EF that is transported to depth, is calculated for the 1500 m and 3200 m depth horizons. The annual TE is 7.6 % at 1500 m depth and 6.2 % at 3200 m depth. Unlike the seasonal pattern in export flux efficiency, the transfer efficiency is highest during the post bloom season (AMJ, 9.1 % and 7.4 %, respectively at 1500 m and 3200 m depths).

The difference between the TE at 1500 m and 3200 m depths increases over the year from 1.6 % in winter (JFM) to 2.4 % in the fall (OND). This divergence indicates that the depth penetration of C_{org} within the bathypelagic zone is about 44 % higher during the JFM period of maximum EF than during the late fall season of minimum EF.

To quantitatively evaluate how our TE estimates might be affected by deep water lateral advection of C_{org} , we estimated an upper limit for the C_{org} advected with the lithogenic material assuming that the primary source of advected C_{org} is resuspended sediments from the US continental slope that have been transported to the Bermuda region via Gulf Stream recirculation. This assumption is supported by the elemental

composition of the lithogenic flux material (Conte et al., 2019) and of deep water suspended particles collected between the US continental margin and Bermuda (Lam et al., 2015). The C_{org} concentration of continental margin sediments at mid-slope (1000–2000 m) depths ranges from < 1% to 3% (Milliman, 1994; Mayer et al., 2002). For this calculation, we assume that the advected sediments have a C_{org} concentration of 3%, which is the maximum concentration of mid-slope sediments off Cape Hatteras near the latitude where the Gulf Stream veers offshore (ibid). We also assume that the resuspended C_{org} is refractory (~ 1300 yrs ^{14}C age) and tightly associated with fine organoclay particles (Mayer et al., 2002) so that there is no remineralization loss of C_{org} relative to lithogenic material during transport. We then use the lithogenic flux data (available for samples collected between 2001–2015, Section 3.2.6) to calculate the maximum C_{org} flux that could be associated with laterally advected resuspended continental slope sediments, and subtract this upper limit from the total C_{org} flux at 1500 m and 3200 m depths.

Table 4(b) gives adjusted estimates calculated after the upper limit estimate of the advected C_{org} flux has been subtracted from the total. (Note that the intPP and C_{org} flux data are calculated for the same 2001–2015 time period as the lithogenic data.) These estimates confirm that deep water lateral advection of sedimentary carbon has a minimal influence on our estimates. The adjusted estimate of the percentage of intPP reaching 1500 m depth is within 0.05% of the unadjusted estimate, and within 0.07% of the unadjusted estimate for 3200 m depth. Differences in estimated TE are small and largely reflect the different integration periods. The annual estimated TE is 0.66% higher at 1500 m depth and 0.46% lower at 3200 m depth. Seasonal trends at both the 1500 m and 3200 m depth horizons do not differ from the unadjusted TE estimates, and the seasonal trend of increasing difference in TE between the 1500 m and 3200 m depth horizons remains clearly apparent.

5. Discussion

5.1. Deep flux magnitude, composition and advected sources

Over the last four decades (1978–2022), the mass flux at the OFP site has averaged $27.2 \text{ mg m}^{-2} \text{ d}^{-1}$, $34.8 \text{ mg m}^{-2} \text{ d}^{-1}$, and $36.8 \text{ mg m}^{-2} \text{ d}^{-1}$ at 500, 1500 and 3200 m depths, respectively (Table 1a). The dominance of biogenic carbonate in the mass flux ($\sim 60\%$ of the total) is characteristic of other oligotrophic and temperate regions (e.g., Fischer et al., 2020; Honda et al., 2017; Honjo et al., 2008; Lampitt et al., 2010; Tan et al., 2020; Waniek et al., 2005). The minor contribution of opal in the mass flux (6.8% of the mass at 500 m depth) is consistent with the low diatom productivity of the Sargasso Sea region (Lomas et al., 2013; Steinberg et al., 2001). The C_{org} flux has averaged $2.4 \text{ mg m}^{-2} \text{ d}^{-1}$, $2.2 \text{ mg m}^{-2} \text{ d}^{-1}$, and $1.8 \text{ mg m}^{-2} \text{ d}^{-1}$ at 500, 1500, and 3200 m depths, respectively (Table 4a), decreasing with depth as C_{org} is remineralized in the water column.

The OFP data underscores the major contribution of advected materials in the deep particle flux, even for this open ocean gyre location. A >5-fold increase in lithogenic flux between 500 m ($1.1 \text{ mg m}^{-2} \text{ d}^{-1}$) and 3200 m ($6.4 \text{ mg m}^{-2} \text{ d}^{-1}$) depths clearly indicates a major deep advected source of lithogenic material, which from its elemental composition appears to have originated from North American continental margin sediments transported to the region via Gulf Stream recirculation and subsequently repackaged into sinking particles (Conte et al., 2019). The incorporation of advected lithogenic materials into the sinking flux leads to a 3-fold increase in the lithogenic fraction of the mass flux between 500 m (6%) and 3200 m (18%) depths. This increases flux ballasting (e.g., Armstrong et al., 2001; De La Rocha and Passow, 2007; Iversen and Ploug, 2010; Klaas and Archer 2002) and potentially increases mean particle sinking rates with depth.

A 3-fold increase in the opal flux between 500 m ($1.8 \text{ mg m}^{-2} \text{ d}^{-1}$) 500 m and 3200 m ($5.2 \text{ mg m}^{-2} \text{ d}^{-1}$) depths similarly points to a significant advected component in the deep opal flux. While deeper traps

have larger surface catchment areas (e.g., Siegel and Deuser, 1997) and thus will collect materials from the area of slightly higher opal production nearer the Bermuda Platform (Conte et al., 2001), differences in opal production in the trap catchment areas are insufficient to account for this nearly 3-fold increase. Comparative microscopic and chemical studies of lithogenic and opaline materials in the deep traps with deep suspended particles upstream of the OFP site could help to identify the source region(s) of advected materials contributing to the deep flux.

5.2. Flux variability

There is significant variability in flux magnitude on multiple time scales (c.f. Figs. 2, 5). A clear seasonal cycle in flux magnitude is observed for all flux components (Fig. 5). This includes the lithogenic component of the flux, which exhibits a seasonal cycle similar to that of the biogenic flux. This covariation provides further evidence of the incorporation of abiogenic materials into the sinking flux as suspended particles are repackaged into sinking particles by particle cycling processes in the water column (e.g., Deuser et al., 1983).

Episodic, high flux events are a characteristic feature of the OFP flux climatology (Fig. 2). Episodic high flux events have been observed in many deep ocean basins (e.g., Billett et al., 1983; Briggs et al., 2011; Dickey et al., 1998; Fischer and Wefer, 1996; Honjo et al., 1999; Lampitt, 1985; Marra et al., 1998; Michaud et al., 2022; Smith et al., 2018). These episodic high flux events have been shown to be particularly important for the transfer of labile carbon and associated elements to the deep ocean (Conte et al., 2019, 2003, 1998; Pedrosa-Pàmies et al., 2019, 2018) and to serve as an important food supply for deep-dwelling and benthic animals (Billett et al., 1983; Durden et al., 2020; Fischer and Wefer, 1996; Gehlen et al., 1997; Gooday, 2002; Graf, 1989; Kiriaoulakis et al., 2001; Lampitt et al., 2001; Pfannkuche, 1993; Rice et al., 1986; Shaw et al., 2020; Smith et al., 2002). The outsized importance of episodic flux events in the transfer of labile materials to depth underscores the need for more research to better understand the upper ocean conditions which trigger these nonseasonal events.

Early winter (Dec – Jan) flux peaks, which occur in some but not all years at the OFP site, can rival the magnitude of the flux peak associated with the spring bloom (Figs. 2, 5). These winter flux events occur during the period when surface stratification is being eroded by seasonal cooling and water column stability is minimal (Lomas et al., 2013; Steinberg et al., 2001), and have been associated with transient blooms induced by physical perturbation of the mixing dynamics by meteorological or mesoscale forcing (e.g. Conte et al., 1998). Early winter flux peaks are particularly apparent in the opal flux (Fig. 5), indicating that these early winter bloom/flux events are enriched in diatoms relative to the main spring bloom. This inference is consistent with observations of enhanced silica production in transient winter blooms in the Bermuda region (Krause et al., 2009).

Flux temporal variability is quantitatively captured by its probability distribution. The value of a distributional approach to the analysis of ocean biogeochemical fluxes is nicely illustrated by the seminal paper of Cael et al. (2018), who showed that while the individual co-located paired measurements of net primary production (NPP) and surface organic carbon export flux (EF) are uncorrelated (Maiti et al., 2013), the relationship between them can be mechanistically linked through their probability distributions. They found that NPP and EF both follow a lognormal distribution over a range of oceanographic settings and are predictably related by their distributions' moments. The smaller log-standard deviation σ of the EF as compared to NPP ($\sigma_{F_{OC}} \sim 0.62 \sigma_{NPP}$) quantified the strong dampening of NPP variability by the processes that collectively transform NPP into the EF. Following on this study, Cael et al. (2021) compared the frequency distributions of NPP and deep fluxes of C_{org} , carbonate and opal at six time series sites- the ALOHA site in the oligotrophic North Pacific (4000 m depth), the CARIOCO site in an anoxic Caribbean coastal basin (1210 m depth), Station K2 in the subarctic northwest Pacific (4810 m depth), Station M

within the California Current upwelling system in the northeast Pacific (3400 m depth), the PAP site in the temperate northeast Atlantic (3000 m depth), and the OFP site (3200 m depth). At all sites, deep fluxes follow lognormal probability distributions whose moments are quantitatively related to that of the overlying NPP. Estimation of deep carbon flux episodicity derived from flux probability distributions indicated that globally about 70 % of total flux occurs during periods of high flux which cover only 30 % of the time.

In this study we show that flux frequency distributions also provide a comprehensive and quantitative evaluation of how flux variability evolves with depth as particles are recycled in the water column. Flux frequency distributions (Fig. 3), together with flux Q-ranges (Table 1), quantify substantial dampening of flux variability and episodicity with depth, especially within the lower mesopelagic zone. At 500 m depth, carbonate and C_{org} flux frequency distributions can be approximated by a Weibull (carbonate) or Gamma (Corg) probability distribution model. These probability distributions are commonly observed in similarly skewed hydroclimatological variables such as rainfall, streamflow and wind speed, with the Weibull probability distribution a better fit in regions that experience more extreme events (De Michele and Avanzi, 2018; Husak et al., 2007; Singh, 1998; Wilks, 1989, 1995; Wilson and Toumi, 2005; Zorzetto et al., 2016). These flux distributions, together with the pronounced compositional variability at 500 m depth (Table 1), suggests that the mesopelagic particle flux is driven by multiple discrete surface export events that are highly variable in both space and time. This inference is supported by glider observations showing high spatial and temporal variability of particle populations within upper mesopelagic waters (e.g., Estapa et al., 2019, 2013; Henson et al., 2023). The high episodicity shown in the 500 m flux distribution is consistent with Flux Stability Indices (FSI, Table 3) indicating that on average, 50 % of the annual carbonate and C_{org} fluxes occur during periods of high flux which cover only 16 % of the time.

The “heavy-tail” distributions (Bryson, 1974) of the carbonate and C_{org} fluxes at 500 m depth indicate that high fluxes are more frequent than predicted by their probability distribution models. “Heavy-tailed” distributions have been shown to occur in hydroclimatology variables when the extreme events arise from a confluence of processes that differ from the typical forcing processes (Wilson and Toumi, 2005; Merz et al., 2022). For example, “heavy-tail” distributions are characteristic of precipitation data in regions that experience extreme rainfall events which arise from an atypical confluence of meteorological forcing (e.g., Cavanaugh et al., 2015; Panorska et al., 2007). “Heavy-tail” distributions are also characteristic of streamflow data in regions that experience floods which arise from a convergence of distinct patterns of atmospheric forcing and nonlinear catchment behavior which differ from the processes responsible for typical streamflow (Merz et al., 2022; Macdonald et al., 2022; Wang et al., 2023). Similarly, the “heavy-tailed” distributions of the mesopelagic particle flux are consistent with evidence that the processes that trigger high flux events, for example short-lived bloom/flux events forced by transient nutrient upwelling/downwelling during passage of energetic mesoscale eddies and extreme storms, differ from the processes which generate typical fluxes.

The marked evolution in flux frequency distributions between 500 m and 1500 m depths (Fig. 3), together with the strong dampening and homogenization of flux magnitude and composition (Tables 1, 2) evidences the intensity of particle recycling within the lower mesopelagic zone. The large compositional differences observed in the biogenic components of the 500 m and 1500 m OFP traps (Pedrosa-Pàmies et al., 2018; Shatova et al., 2012; unpublished OFP microscopic data), also observed in mesopelagic sediment traps in other regions (e.g., Belcher et al., 2023, 2016; Urrère and Knauer, 1981; Wilson et al., 2008), indicates that this depth evolution in the particle flux within the lower mesopelagic zone is strongly mediated by zooplankton grazing activities.

Between 1500 m and 3200 m depths, the changes in flux frequency distributions (Fig. 3) and the dampening of flux variability are much

reduced (Tables 1, 2). This suggests that flux recycling at bathypelagic depths is much less intense, presumably due to lower zooplankton grazing pressure. This inference is supported by lipid biomarker evidence showing minimal change in suspended particle organic composition with depth in the bathypelagic zone at the OFP site (Pedrosa-Pàmies et al., 2018).

An interesting observation is the divergence in carbonate and Corg flux frequency distributions at 1500 m and 3200 m depths. While the carbonate flux closely follow the lognormal probability model at both 1500 and 3200 m depths, the C_{org} fluxes continue to exhibit “heavy tail” distributions. This divergence suggests that particle cycling processes within the deep water column impact the carbonate and C_{org} fluxes differently. This might arise due to the longer water column residence time of carbonate, a quasi-stable mineral, versus C_{org} , which is subject to remineralization. Thus, the deep carbonate flux will be impacted to a greater extent by water column processes which affect deep flux regeneration (e.g. suspended particle aggregation), while the deep C_{org} flux will be impacted to a greater extent by high flux events that enhance C_{org} transfer to deep waters.

5.3. Flux recycling and modulation within the water column

The particle flux in the deep water column is comprised of a mixture of particles having different ages, sources and sinking rates, and includes materials supplied by two distinct flux generating processes. The first is coupled to the biogeochemical and physical processes operating within the surface ocean that drive primary production and the physical and biological pumps (e.g. Boyd et al., 2019) which collectively generate the overlying surface export flux. The second is coupled to the myriad particle cycling processes operating within the water column- e.g. microbial degradation and remineralization, animal feeding activities (e.g., Belcher et al., 2016; Hernández-León et al., 2020; Iversen and Ploug, 2010; Robinson et al., 2010; Steinberg and Landry, 2017; Turner, 2015; Vinogradov, 1962; Zhang and Dam, 1997), particle coagulation and disaggregation (Burd and Jackson 2009 and references therein)- that attenuate and modify the sinking flux and produce fresh sinking particles at depth (e.g. Bishop et al., 1986; Carroll et al., 1998; Shatova et al., 2012; Urrère and Knauer, 1981; Wilson et al., 2008). Collectively, the fraction of the deep sinking flux produced internally within the water column by particle cycling processes has been termed the “intrinsic” flux component (Conte, 2019) to emphasize that it is intimately tied to the ecosystem processes driving particle cycling and flux regeneration via large particle production at depth rather than being directly supplied by the overlying surface export flux. While one fraction of the intrinsic flux produced by aggregation of suspended particles in the water column could be considered a “delayed” flux- albeit aged and compositionally modified- of disaggregated flux materials originally supplied by the overlying export flux, the “intrinsic” flux is also an amalgamation of many additional types of particulate materials which are present in the deep particle pool and not directly supplied by the overlying surface export flux. These materials include deep water animal debris and microbial biomass, materials supplied by lateral advection, authigenic minerals precipitated within the water column, and dissolved materials scavenged by particle surfaces.

As the particle flux is consumed, recycled and regenerated during its downward transport, the fraction of the flux that is directly supplied by the overlying surface export flux will decrease with depth while the fraction of the flux that is supplied by “intrinsic” flux generation will increase. This depth evolution will result in a dampening of flux seasonality with depth, as the ecosystem processes that drive particle cycling and flux regeneration within the water column are less tightly coupled to the seasonal cycle in the surface export flux. This dampening is clearly observed in the OFP record (section 5.2) and has also been observed at the Ocean Station Papa site in the highly productive NE Pacific (Timothy et al., 2013), suggesting that dampening of flux seasonality with depth occurs across all productivity regimes.

The OFP record also evidences a seasonal progression in the relative fraction of the deep particle flux that is supplied by “intrinsic” flux generation processes (e.g. suspended particle aggregation) within the water column. When the overlying export flux is at its seasonal minimum in late summer and fall, the carbonate (and opal) fluxes at 1500 m and 3200 m depths are significantly higher than that at 500 m depth (Fig. 5). Given that the annual carbonate flux is quasi-conservative with depth (Table 1), this indicates that the deep carbonate flux during the season when overlying export flux is at a minimum is largely generated via net aggregation of suspended particles that have accumulated in the deep particle pool from flux disaggregation during the preceding winter/spring high flux period, rather than being directly supplied by the overlying flux.

A notable observation is that the seasonality of the deep lithogenic flux mirrors that of the biogenic flux (Fig. 5) rather than the seasonality of atmospheric dust deposition, which peaks in the late summer and fall (Arimoto et al., 1995). Because the deep lithogenic flux is predominately generated by aggregation of suspended lithogenic particles sourced from deep water lateral advection, this seasonal covariation suggests that the rate of suspended particle aggregation within the water column (i.e. the rate of “intrinsic” flux generation) is sensitive to the magnitude and/or composition of the biogenic flux. There are two nonexclusive reasons why this might be. First, rates of passive particle coagulation, which scale with particle encounter rates (Burd and Jackson, 2009), may be higher during the winter/spring bloom period due a general increase in particle concentrations. Additionally, concentrations of “sticky” mucus-like materials such as TEP are higher during phytoplankton blooms (Engel et al., 2004; Kjørboe and Hansen, 1993; Passow, 2002) which might also increase the efficiency of particle coagulation.

Second, zooplankton grazing activity could be stimulated by the increased availability of labile food particles in the water column during the bloom season, which would lead to higher rates of suspended particle repackaging and large particle production. Epipelagic zooplankton are known to strongly modulate grazing rates in response to changes in food availability (e.g., Deibel, 1998; Flood and Deibel, 1998; Paffenhöfer and Van Sant, 1985). Similar adaptive feeding behavior in deeper dwelling animals has been indicated by recent studies finding that mesopelagic zooplankton alter their feeding strategies in response to flux seasonality (Hannides et al., 2020). Energetically, adaptive feeding behavior also would be expected in bathypelagic animals, which inhabit an extremely food-limited environment. In particular, adaptive feeding by gelatinous zooplankton would have large consequences for intrinsic flux generation because of their exceptionally high water clearance rates, suspended particle retention efficiencies and large particle production rates (e.g. Gorsky and Fenaux, 1998; Katija et al., 2017; Robinson et al., 2010; Yoon et al., 2001; Youngbluth et al., 2008). Given that extremely low particle concentrations limit abiotic particle encounter rates in the deep ocean, an adaptive feeding response by zooplankton to flux lability could be expected to dominate modulation of intrinsic flux generation rates in the deep water column.

Conte et al. (2003) provide a case study indicating modulation of intrinsic flux generation rates by an adaptive feeding response of deep-dwelling zooplankton grazers to an episodic bloom/flux event at the OFP site. In this event, as a highly productive mesoscale feature passed over the OFP site both mass and carbonate fluxes at 3200 m depth abruptly increased 2.5-fold. Phytoplankton biomarkers showed that this abrupt flux increase occurred just at the onset of a bloom/flux event when the fastest sinking phytodetrital material began to arrive at 3200 m depth, before the arrival of the main pulse of phytodetritus from the overlying bloom. Coincident with the abrupt increase in mass flux, fluxes of zooplankton-derived fecal lipids and refractory organic carbon compounds also increased, which indicated that the abrupt increase in mass flux was due to increased zooplankton grazing activity and rates of suspended particle aggregation. This temporal sequence strongly suggested that deep zooplankton grazers abruptly increased their feeding activity upon initial detection of labile phytodetrital food particles,

which increased rates of suspended particle aggregation and large particle production in the water column. Given that particle sinking rates range more than two order of magnitude (e.g. Yoon et al., 2001), a similar adaptive feeding response by zooplankton grazers to variations in flux lability could contribute to the remarkably tight temporal coupling of the particle flux with depth (Figs. 4-6).

5.4. Carbon flux linkage to overlying primary production and flux attenuation with depth

One of the first discoveries of the OFP time-series was the observation that the deep ocean environment was not constant but tightly linked to the seasonal cycle of overlying primary production via the “particle rain” (Deuser, 1986). Here we have quantified this surface- deep ocean linkage by directly comparing 30-year integrated averages in the seasonal cycle of the deep particle fluxes with the seasonal cycle in overlying primary production and surface export flux measured concurrently on monthly BATS cruises (Fig. 7, Table 4a). The comparisons show the C_{org} flux throughout the water column strongly covaries, with no time lag between surface carbon export and the deep flux within the one month temporal resolution of the BATS data. The seasonal maximum in the C_{org} flux lags the seasonal maximum in primary production by approximately one month, providing a robust estimate of the timescale of flux generation in the oligotrophic surface ocean. A comparable time lag (~1.5 months) between primary production and the deep flux was previously suggested from a comparison of the bimonthly-resolved 3200 m flux data and satellite-derived primary production between 1979–1985 (Deuser et al., 1990). Similar time lags between primary production and the carbon flux have been estimated for the PAP site in the NE Atlantic (Frigstad et al., 2015) and sites in the eastern subtropical Atlantic (Neuer et al., 1997; Waniek et al., 2005).

The OFP/BATS data comparison allows the efficiency of carbon export through the water column to be directly calculated (Table 4a). Over this 30-year (1989–2021) period, the percentage of annual intPP transferred to 150 m depth (i.e. the surface export efficiency or e-ratio) has averaged 6.6 %. This annual estimate is about 20 % higher than the 30 year average e-ratio measured at 150 m depth at Station ALOHA off Hawaii ($5.4 \% \pm 0.021$, Karl et al., 2021). The percentage of intPP transferred to 500 m, 1500 m and 3200 m depths has averaged 0.6 %, 0.5 % and 0.4 %, respectively, over this 30 year period.

The data clearly show that efficiency of carbon export to mesopelagic depths varies seasonally and is strongly correlated with the seasonal cycle in primary production. The percentage of the overlying intPP exported to 500 m depth ranges from a high of 0.7 % in winter (JFM) to a low of 0.4 % in summer and fall. Seasonal variation in export efficiency is much reduced at 1500 m depth but is still 20 % higher during the period of peak primary production. By 3200 m depth, no seasonal trends are apparent. These depth trends provide further indication of the dampening of seasonality with increasing depth due to particle cycling.

Carbon transfer efficiency (TE) from 150 m depth to the 1500 m and 3200 m depth horizons average 7.6 % and 6.2 %, respectively, and also varies seasonally but with the maximum efficiency occurring in the post-bloom period (Table 4a). At 1500 m depth, carbon transfer efficiencies range from 7.2 % in during the winter productivity peak (JFM) to 9.1 % during the post-bloom period (AMJ). At 3200 m depth, carbon transfer efficiencies range from 5.6 % in winter (JFM) to 7.4 % in spring (AMJ). These estimates of transfer efficiency at the OFP site are similar to estimates of transfer efficiency at the subtropical oligotrophic western North Pacific at the S1 site (Honda, 2020), where the carbon transfer efficiency at the 4800 m depth horizon was estimated as 4.8 %. Using a first-order assumption of a constant rate of C_{org} flux attenuation with depth in bathypelagic waters ($0.235 \text{ mg } C_{org} \text{ per } 1000 \text{ m}$ for the 1500 m – 3200 m depth interval at the OFP site), the transfer efficiency at the 3200 m depth horizon at the S1 site would be approximately 5.5 %, which is nearly identical to the wintertime transfer efficiency at the OFP site (5.6 %) when C_{org} export flux at both sites are similar. Frigstad et al.

(2015) calculated a slightly lower transfer efficiency (4.4 %) for the 3000 m depth horizon at the PAP site in the NE Atlantic. These direct estimates of carbon transfer efficiency also are comparable to modelled carbon transfer efficiencies to deep waters in subtropical regions that are based upon deep ocean nutrient concentrations (Weber et al., 2016). In contrast, modelled transfer efficiencies in subtropical regions that are based upon ^{234}Th -derived export fluxes (Henson et al., 2012) are significantly higher.

A novel finding is that while the efficiency of carbon export out of the surface ocean is highest during the bloom period, the efficiency of carbon transfer to deep waters is highest during the post bloom period. This finding is consistent with recent studies in the upper mesopelagic waters of the Southern Ocean that found that surface export flux efficiency was highest during the early bloom stage and then decreased, whereas the carbon transfer efficiency increased as the bloom waned (Henson et al., 2023). The difference suggests a seasonal evolution in the strength of coupling between primary production and secondary consumption, which could result from a lag in mesopelagic zooplankton grazing pressure relative to the rise in primary production due to the longer zooplankton generation times. Thus, particle consumption and recycling would be less efficient during the bloom period leading to more surface carbon export, whereas carbon transfer within the water column would be more efficient during the post bloom period as zooplankton grazing activity and intrinsic flux generation within the water column increases.

Another novel finding is that as the season progresses the difference between carbon transfer efficiencies at the 1500 m and 3200 m depth horizons increases by a remarkable 40 % (Table 4a), indicating that carbon remineralization within bathypelagic waters increases as overlying productivity and export flux declines. This seasonal trend in the efficiency of C_{org} remineralization is somewhat enigmatic as the deep C_{org} flux would be expected to become more refractory over the year as overlying surface export flux decreases. One hypothesis is that zooplankton grazing pressure and particle cycling rates at bathypelagic depths increase over the year due to increasing zooplankton abundance and/or changes in trophic structure (e.g., Hannides et al., 2020; Koppelman and Weikert, 1999; Romero-Romero et al., 2019), resulting in more efficient C_{org} remineralization. The larger inference is that carbon transfer efficiency, and the transfer of other elements associated with biogenic debris, is intimately tied to the ecosystem processes that drive particle cycling and remineralization in the deep water column.

6. Concluding remarks

The OFP time-series has produced an unparalleled four decade climatology of the deep ocean particle flux in the oligotrophic western North Atlantic gyre. The OFP data, together with co-located BATS time-series data of primary production and surface ocean carbon export, allow for robust in-situ estimation of the climatological average depth penetration of carbon within the water column that constrains estimates of deep ocean carbon sequestration in the oligotrophic ocean.

Deep particle flux time-series such as the OFP continuously record surface ocean biogeochemical status integrated over mesoscale spatial scales and thus are highly complementary to ship-based time-series, which are extremely limited in both temporal and spatial coverage. Additionally, the recovered flux material retains a rich diversity of biogeochemical signals and can be archived for future study using ever more sophisticated analytical methodologies.

The OFP record clearly shows the deep particle flux is an admixture of materials supplied by the overlying surface export flux and materials that are supplied by the myriad particle cycling processes which attenuate, modify and regenerate the sinking flux within the water column. Although long appreciated that the deep flux is closely tied to processes driving the seasonal cycle in overlying primary production and surface export flux, our knowledge of the ecosystem processes driving particle cycling and flux regeneration within the deep ocean interior remains quite limited and is deserving of greater research focus.

As the OFP time-series moves forward into the future, the deep flux climatology presented here will serve as an invaluable baseline record to assess the consequences of a changing climate on biogeochemical cycles in this northwestern quadrant of the oligotrophic North Atlantic gyre. In particular, change in the frequency of episodic, high flux events would have an outsized impact on carbon and material transfer efficiencies through the water column. The statistics of infrequent events have been shown to be more dependent on system variability than mean state (Katz and Brown, 1992). The implication is that high-frequency variability of the climate-ocean system and how this variability affects ocean biogeochemical processes must be better understood and characterized to improve predictions of how different climate scenarios will impact the oceanic particle flux and deep ocean carbon sequestration.

CRediT authorship contribution statement

M.H. Conte: Writing – review & editing, Writing – original draft, Visualization, Validation, Supervision, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis, Data curation, Conceptualization. **R. Pedrosa-Pàmies:** Writing – review & editing, Supervision, Investigation, Data curation. **J.C. Weber:** Writing – review & editing, Supervision, Investigation, Formal analysis, Data curation. **R.J. Johnson:** Writing – review & editing, Investigation, Data curation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix 1. Deep currents at the OFP mooring site

Current speeds measured by the ADCP at the top of the OFP mooring (300 m depth, 8.0 yr record) and by FSI 2D current meter deployed on the OFP mooring (750 m depth, 3.2 yr record) are summarized in Table A1. Current velocity stick plots for each deployment are shown in Fig. A1.

At 300 m depth, the median current speed was 16 cm s^{-1} . Occasionally, currents increased to $> 35 \text{ cm s}^{-1}$ for short periods during passage of energetic eddies or storm events. These episodes occurred in $\sim 5 \%$ of the record.

Currents are much reduced at 750 m depth. The median current speed was 11 cm s^{-1} . Currents exceeded 35 cm s^{-1} in $< 0.1 \%$ of the record, and 25 cm s^{-1} in only 2% of the record. The tilt of the in-line FSI current meter at 750 m depth averaged 1.1° , exceeding 4.0° in $< 1 \%$ of the record.

We also show data from a current meter mooring deployed at $31^\circ 47' \text{ N}$, $64^\circ 26' \text{ W}$, just west of the OFP site, between Apr 1975 to Jan 1976 (McKee et al., 1981). This mooring deployed vector averaging current meters (VACM, AMF/EG&G Sealink Systems) at 300, 500, 1000, and 1500 m depths. At 300 m depth, current speeds were similar to those measured by the OFP's ADCP. At 500 m depth, current speeds were about 20 % lower. At 1000 m depth, current speeds were 20–30 % of those at 500 m depth, and by 1500 m depth currents averaged $< 1 \text{ cm s}^{-1}$ and never exceeded 5 cm s^{-1} .

Table A1

Current speeds (cm s^{-1}) measured by the the RDI Workhorse ADCP and FSI 2D ACM current meter and deployed on the OFP mooring at nominal depths of 300 m and 750 m, respectively. Record lengths at 300 m and 750 m depths are 8.0 and 2.6 years, respectively. Tilt ($^\circ$) and temperature ($^\circ\text{C}$) data collected by the FSI 2D ACM is also shown. The ADCP data is the average of the lowest 8 m sample bin. Time weighted averages of all deployments are shown.

(a) ADCP deployments (nominal depth 300 m)

Start date	End date	Days	Sampling rate (min)	Depth avg (m)	Current speed (cm s^{-1})		Percentage of Record			
					Avg (range)	Median	$>20 \text{ cm s}^{-1}$	$>25 \text{ cm s}^{-1}$	$>30 \text{ cm s}^{-1}$	$>35 \text{ cm s}^{-1}$
11/26/12	4/17/13	142	60	349	15 ($<1-48$)	14	21 %	12 %	7 %	3 %
11/11/13	10/14/14	337	60	274	12 ($<1-60$)	10	8 %	5 %	5 %	4 %
10/23/14	4/18/15	177	20	280	12 ($<1-40$)	11	15 %	5 %	1 %	0 %
4/19/15	11/14/15	209	60	279	16 ($<1-50$)	14	31 %	16 %	7 %	3 %
11/16/15	4/26/16	162	20	317	25 ($<1-59$)	23	60 %	44 %	29 %	19 %
4/27/16	10/26/16	182	60	284	14 ($<1-41$)	12	24 %	14 %	5 %	1 %
10/29/16	4/24/17	177	20	283	14 ($<1-49$)	13	21 %	9 %	3 %	1 %
4/25/17	8/18/17	115	60	275	14 ($<1-47$)	13	21 %	7 %	3 %	1 %
8/25/17	1/28/18	156	20	337	21 ($<1-65$)	21	53 %	32 %	16 %	8 %
1/29/18	6/3/18	125	20	347	18 ($<1-40$)	19	40 %	15 %	4 %	1 %
6/4/18	10/28/18	146	15	289	24 ($<1-56$)	23	65 %	43 %	28 %	17 %
10/29/18	4/27/19	180	20	296	19 ($<1-52$)	18	43 %	22 %	10 %	3 %
4/29/19	8/10/19	103	60	292	23 ($1-61$)	22	60 %	39 %	24 %	11 %
11/2/19	8/7/20	279	60	268	15 ($<1-48$)	15	27 %	12 %	4 %	1 %
8/9/20	4/17/21	251	60	306	23 ($<1-66$)	22	56 %	40 %	27 %	16 %
4/19/21	10/30/21	194	15	266	13 ($<1-42$)	11	16 %	9 %	4 %	1 %
Weighted average				293	17	16	33 %	19 %	10 %	5 %

(b) FSI 2D ACM deployments (nominal depth 750 m)

Start date	End date	Days	Sampling rate (min)	Depth avg (m)	Temp ($^\circ\text{C}$) avg (range)	Tilt ($^\circ$) avg	Current speed (cm s^{-1})		Percentage of Record			
							avg (range)	Median	$>20 \text{ cm s}^{-1}$	$>25 \text{ cm s}^{-1}$	$>30 \text{ cm s}^{-1}$	$>35 \text{ cm s}^{-1}$
11/20/05	1/15/06	56	120	nd	nd	nd	7 ($<1-16$)	6	0 %	0 %	0.0 %	0.0 %
8/28/06	8/29/07	366	120	nd	10.5 (8.6–12.8)	1.3	13 ($<1-41$)	12	14 %	4 %	0.2 %	$<0.1 \%$
11/4/08	7/23/09	261	120	nd	11.0 (8.7–12.6)	0.9	10 ($<1-26$)	9	2 %	0 %	0.0 %	0.0 %
7/26/09	11/21/09	118	120	nd	11.5 (8.5–13.4)	1.0	12 ($<1-36$)	11	9 %	2 %	0.4 %	$<0.1 \%$
4/15/10	7/14/10	90	20	760	10.4 (9.0–9.0)	nd	11 ($<1-28$)	10	6 %	1 %	0.0 %	0.0 %
7/16/10	9/17/10	63	20	752	11.0 (9.7–11.9)	nd	10 ($<1-31$)	9	7 %	1 %	0.1 %	0.0 %
Weighted average					10.9	1.1	11	11	8 %	2 %	0.1 %	$<0.1 \%$

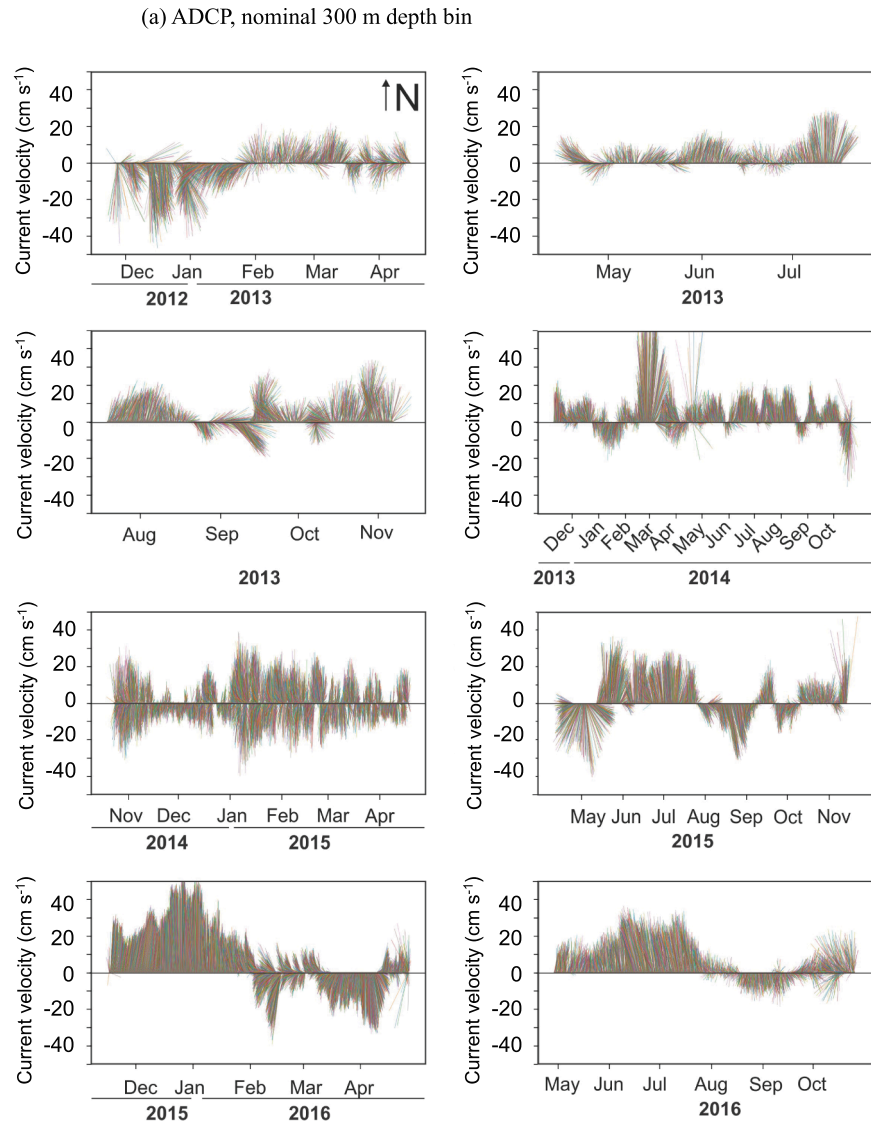


Fig. A1. Current velocity stick plots. (a) Currents in the bottom (8 m) bin of ADCP deployments on the OFP mooring, nominal depth 300 m (b) Currents measured by the FSI 2D ACM on the OFP mooring, nominal depth 750 m. (c) Currents measured by velocity averaging current meters (VACMs) on a mooring deployed just west of the OFP site ($31^{\circ} 46.9'N$, $64^{\circ} 26.2'W$, 4353 m water depth) from Apr 1975 to Jan 1976 (McKee et al., 1981).

(a) cont.

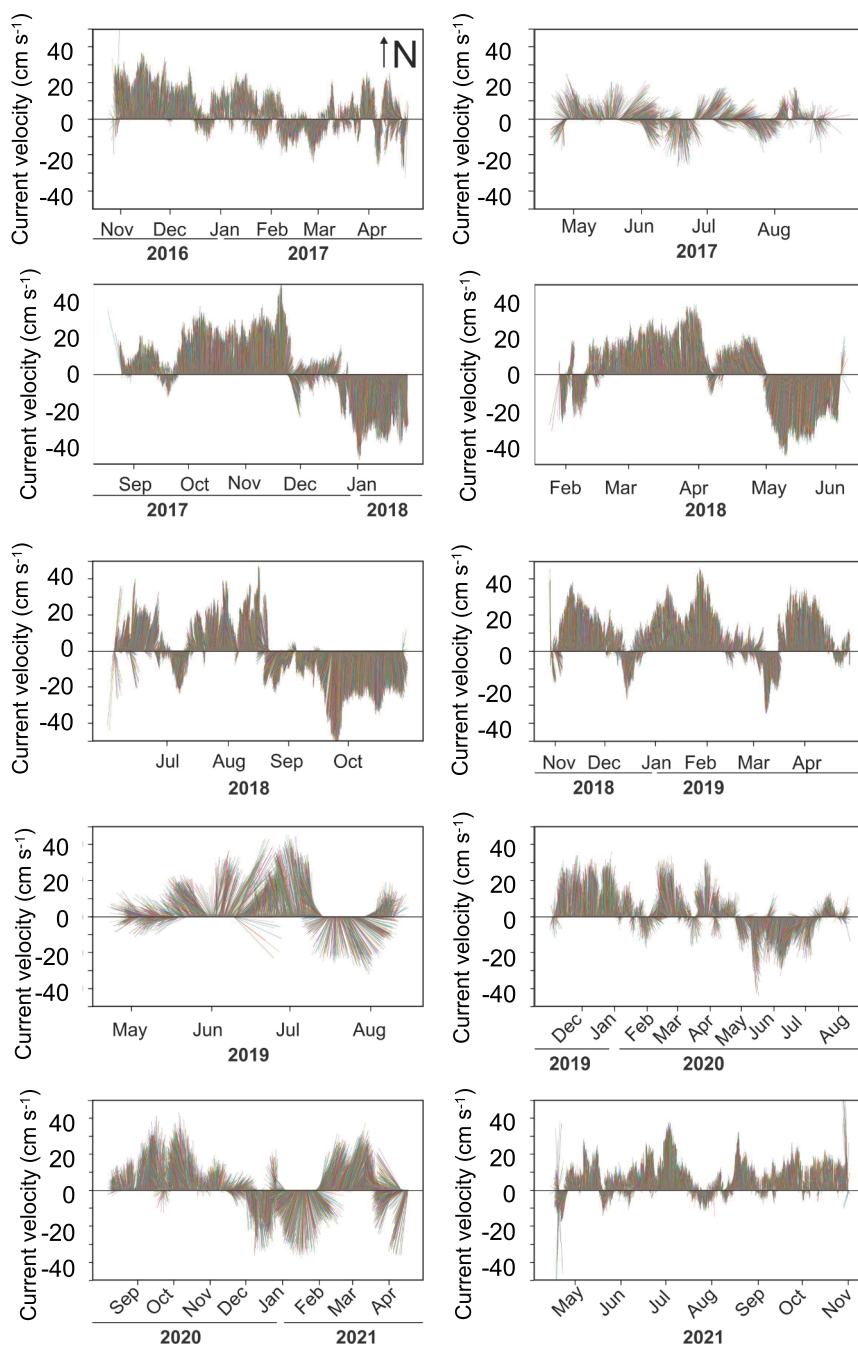


Fig. A1. (continued).

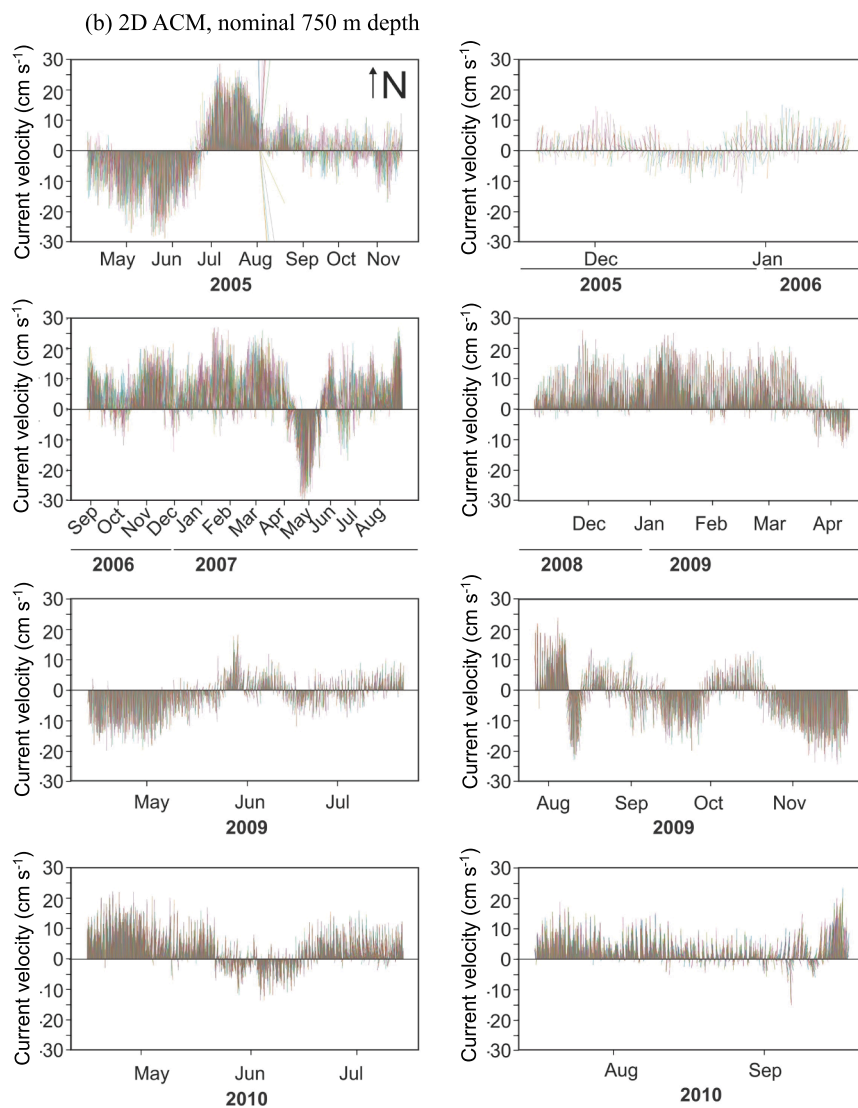


Fig. A1. (continued).

(c) VACM profile, mooring 553 (from McKee et al., 1981)

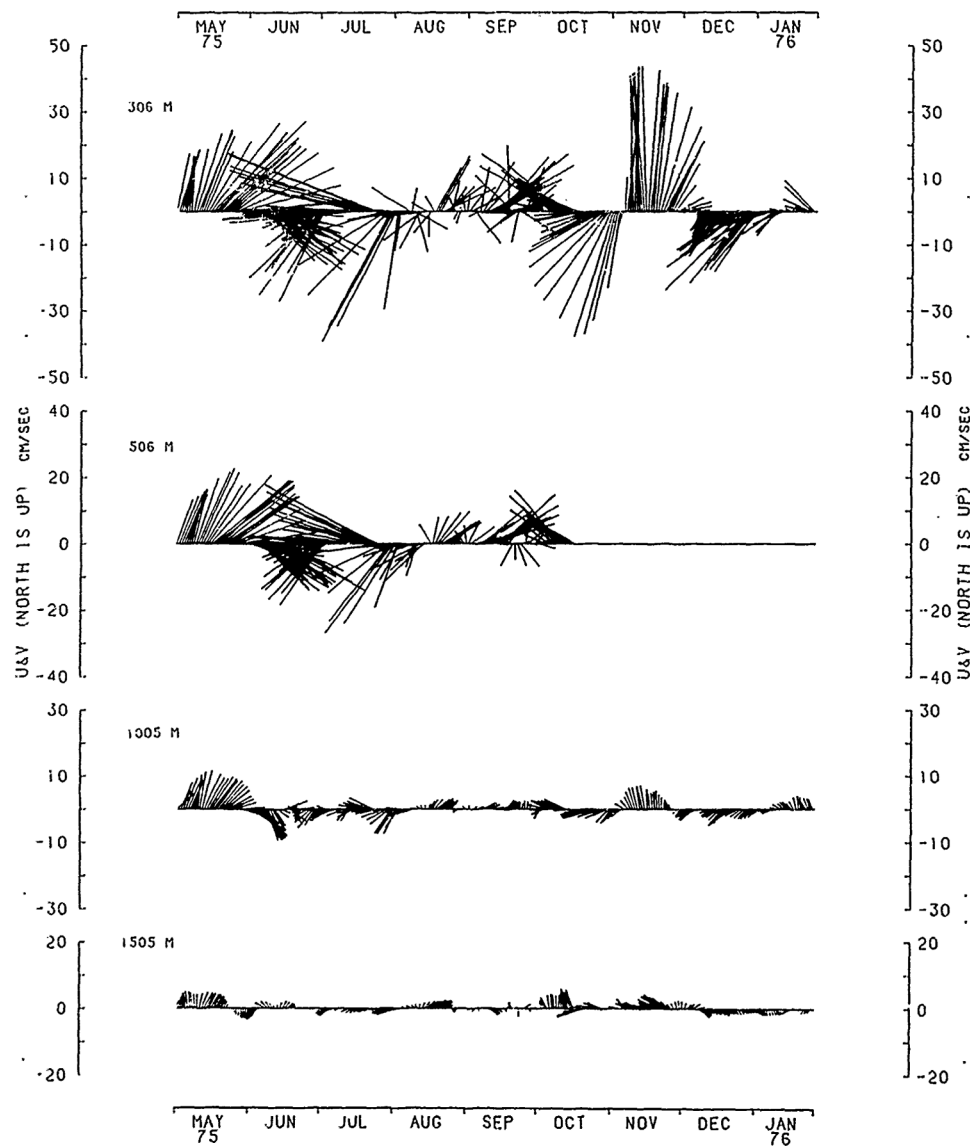


Fig. A1. (continued).

Appendix 2. Swimmers in the OFP traps

Some pelagic animals are attracted to moored equipment, including sediment traps. These animals may affect the flux by feeding upon particles collecting on trap surfaces before they enter the collection cup, or by introducing materials (biomass, fecal material, regurgitated stomach contents) which are not part of the sinking flux. Fish are a well-documented problem for traps sampling at mesopelagic depths. Fish entering the collection cup can seriously contaminate the cup contents, with occasional carryover of fish debris on the sampling carousel contaminating subsequent cups. Debris from fish activity above the trap funnel opening can also fall into the trap.

“Swimmers”, defined as animals that have entered the trap collection cup alive, are removed during initial sample processing of OFP trap material. We examine the > 125 mm size fractions under a stereomicroscope to identify as swimmers any zooplankton that have completely intact body membranes and no visible signs of decay. Swimmers and any fish debris (bone, scales) are carefully removed using tweezers before further sample processing. While swimmers and fish debris usually can be quantitatively removed from the samples, the overall impact of animals potentially grazing on material collecting on the interior trap surfaces is largely unknown.

Since May 2001 we have quantified swimmers in the traps and also preserved them as part of the OFP sample archive. Table A2 summarizes data on swimmer taxa, abundance and composition and fish debris in the 500, 1500, and 3200 m traps (401, 429 and 494 samples, respectively). The most common invertebrate taxa in the OFP traps are crustaceans (amphipods, copepods, mysids, euphausiids, shrimps, ostracods, and cyprid larvae). On rare occasions isopods, polychaetes, nematodes and gelatinous zooplankton are found.

The 500 m trap contains significantly higher swimmer abundances than traps at 1500 m and 3200 m depths. At 500 m depth, about 10 %, 17 % and 24 % of 500 m trap samples contained > 10 individual ostracods, amphipods and copepods, respectively; by 1500 m depth these percentages drop to 0 %, 1 % and 8 %, respectively. Swimmer abundances in the 3200 m trap are very low, with < 5 % of samples having more than 1–2 individuals of any

taxa.

There is a shift in some swimmer taxa with depth. Ostracods were generally found only in the 500 m trap. Mysids were abundant only in the 1500 m trap, with 7 % of samples having > 10 individuals.

Seasonality was only observed in ostracod and mysid abundances. Ostracods in the 500 m trap were most abundant between Jun – Oct. Mysids in the 1500 m trap were most abundant between May – Nov and nearly absent from Jan – Apr.

Fish are far more common in the 500 m trap than in the deeper traps. Approximately 10.3 % of the 500 m trap samples (46 cups) were contaminated by large fish debris and excluded from further analysis. An additional 13.2 % of the samples had fish debris (flesh, scales, bones) that could be removed. Intact small fish, principally *Cyclothone* spp., and fish larvae were present in about 5 % of the 500 m trap samples but < 0.5 % of trap samples at 1500 m or 3200 m depths. The 1500 m and 3200 m traps are largely unaffected by fish. Only 5 samples (1.1 % of the total) in the 1500 m trap were contaminated by large fish debris, and 7.7 % of the remaining samples had small amounts of fish debris. No 3200 m trap samples were contaminated by large fish debris, and only 1.6 % of samples had any fish scales or bones.

These data indicate that the OFP trap at 500 m depth may be moderately impacted by swimmer and fish activity, but that traps at 1500 m or 3200 m depths are minimally impacted.

Table A2

Zooplankton swimmers and fish debris in the OFP trap samples at 500, 1500 and 3200 m depths. Trap cups contaminated by large fish debris (45 and 5 samples, respectively in the 500 and 1500 m traps) were excluded from analysis.

No. individuals in sample	Number of samples									
	Ostracods	Amphipods	Copepods	Mysids	Euphausiids	Caridia	Cyprid-like larvae	Small Fish + Larvae	Fish Scales	Fish Bones
<i>(a) 500 m trap (401 samples)</i>										
0	233	277	209	381	394	352	295	250	276	233
1–2	84	31	36	16	5	38	31	40	34	35
3–5	30	15	31	4	1	5	17	30	38	30
6–10	14	10	28	0	0	3	26	17	16	14
11–20	16	21	37	0	1	1	15	19	8	16
21–30	6	18	28	0	0	1	8	6	8	6
31–50	7	14	13	0	0	1	8	10	6	7
51–100	7	12	19	0	0	0	1	12	5	7
>100	4	3	0	0	0	0	0	23	9	4
<i>(b) 1500 m trap (429 samples)</i>										
0	418	283	70	253	425	422	328	426	347	393
1–2	10	105	91	58	4	6	70	2	33	27
3–5	1	33	142	56	0	1	27	0	5	7
6–10	0	3	93	33	0	0	4	0	9	1
11–20	0	0	30	26	0	0	0	1	7	1
21–30	0	2	2	3	0	0	0	0	2	0
31–50	0	0	1	0	0	0	0	0	4	0
51–100	0	0	0	0	0	0	0	0	5	0
>100	0	3	0	0	0	0	0	0	17	0
<i>(c) 3200 m trap (494 samples)</i>										
0	479	441	294	493	493	490	377	493	468	469
1–2	2	50	177	1	1	4	108	1	22	20
3–5	13	2	23	0	0	0	7	0	1	2
6–10	0	0	0	0	0	0	2	0	1	1
11–20	0	0	0	0	0	0	0	0	0	2
21–30	0	0	0	0	0	0	0	0	0	0
31–50	0	1	0	0	0	0	0	0	0	0
51–100	0	0	0	0	0	0	0	0	0	0
>100	0	0	0	0	0	0	0	0	2	0

Data availability

Data will be made available on request.

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