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Note on volume and distribution of fresh water in the Amazon River plume under low discharge conditions

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Supplementary material for this article is available [online](#)

Abstract

In this research communication, we report the results of a field survey conducted in a part of the plume of the Amazon River between 0° and 5°N and offshore of the 28 m isobaths in November of 2022, during the low river discharge season. By comparing the observed vertical salinity profiles ‘disturbed’ by continental discharges within the plume with the virtually ‘undisturbed’ ones outside the plume, we estimated the total content of fresh water in the area covered by the measurements as $203 \pm 22 \text{ km}^3$, which equals to less than 3% of the average annual Amazon River discharge. Furthermore, we argue that the river-borne continental water was not confined to the upper mixed layer and show that about 37%, or 76 km^3 , of its volume was entrained into the plume-underlying layer between the mixed layer and the salinity maximum. This point is additionally supported by analysis of chromophoric dissolved organic matter (CDOM) fluorescence in water samples, demonstrating significant concentrations of terrigenous CDOM to depths up to 140 m. We also observed that there was a significant direct correlation between the volume of freshwater accumulated in the affected layer and background stratification (expressed as buoyancy frequency) in the unaffected layer below it.

1. Introduction

With its average annual discharge of almost $7,000 \text{ km}^3$, the Amazon River (AR) not only heads the list of the World’s biggest rivers—the volume of water it discharges into the western tropical Atlantic exceeds the combined contributions from the next five entries on the list (i.e., Congo, Ganges, Orinoco, Yangtze, and Rio de la Plata) and amounts to nearly a fifth of the global freshwater discharge. The AR has its maximum runoff of about $2.5 \times 10^5 \text{ m}^3 \text{ s}^{-1}$ in May and a minimum of approximately $1 \times 10^5 \text{ m}^3 \text{ s}^{-1}$ in November, thus demonstrating strong seasonal cycling (e.g., Dai and Trenberth 2002, Gouveia et al 2019a, 2019b). However, even during the low runoff period, the AR discharge still exceeds the outflow from any other of the World’s rivers.

This AR generates a very dynamic and extensive plume, which affects the physical and biogeochemical properties of the ocean over large areas (e.g., Nittrouer and DeMaster 1996). The waters of the AR plume are rich in nutrients and organic matter whose composition is subject to alterations associated with bacterial, planktonic, and photochemical drivers (Medeiros et al 2015). Currents off the Amazon delta have a strong tidal component, which is a major source of kinetic energy for mixing (Geyer 1995), while the non-tidal velocity fields are mainly wind-driven (Lentz and Limeburner 1995). The plume exports fresh water from the western tropical North

Atlantic interacting with the North Brazil Current (Muller-Karger *et al* 1988). Coles *et al* (2013) showed that freshwater associated with the Amazon River influences surface salinity to as far as 20°W and identified principal pathways of the Amazon plume. The Amazon plume exhibits strong interannual variability largely controlled by wind-driven advection (Fournier *et al* 2017) and demonstrates significant correlation with the North Atlantic Oscillation (NAO) and El Niño Southern Oscillation (ENSO) indices, weakening during transitions of the NAO and ENSO from negative to positive phases (Da and Foltz 2022). Numerical experiments demonstrated that the Amazon discharge intensifies near-surface currents, leads to shallowing of the upper mixed layer across the western tropical Atlantic, and forms a barrier layer impeding vertical mixing and exchanges between the upper ocean and deeper parts of the water column (e.g., Varona *et al* 2019).

The AR plume has been a topic of numerous publications, and a considerable number of field campaigns have been conducted in the plume area over the last decades (e.g., Geyer and Kineke 1995, Smith and DeMaster 1996, Silva *et al* 2005b, Coles *et al* 2013, Mascarenhas *et al* 2016, Araujo *et al* 2017). In particular, Silva *et al* (2005a,b) produced detailed maps of the thermohaline structure of the plume in all seasons. Nevertheless, the available *in situ* data are still limited, particularly for the low discharge period. In this research communication, we report results from a single field campaign conducted in boreal autumn. However, we believe that joint measurements of thermohaline fields and terrigenous dissolved organic matter data based on optical properties yielded interesting new findings as to the spatial distribution and the volume of fresh water in the AR plume.

Optical techniques used in this study to analyze chromophoric DOM (CDOM) have become widespread. Spectroscopic indices calculated on the basis of absorbance and fluorescence spectra, such as humification (HIX) and biological (BIX) indices, allow fast determination of predominant DOM sources (Derrien *et al* 2017). Furthermore, PARAFAC decomposition of excitation–emission matrices (EEM) (Bro 1997) resulting in the concentrations of individual components comprising the data, may provide an advanced understanding of organic matter dynamics in aquatic ecosystems. Thus, in the Amazon River plume and the adjacent areas, a strong dilution of seawater with river water was found to be a dominant process in shaping the conservative behavior of humic-like components along the salinity gradient (Hu *et al* 2004, Cao *et al* 2016). The conservative behaviour of some of independent fluorophores along the salinity gradient allow considering humic-like fluorescence as a tracer for terrestrial derived material refractory to bacterial utilization (i.e., Yamashita *et al* 2008, Drozdova *et al* 2022).

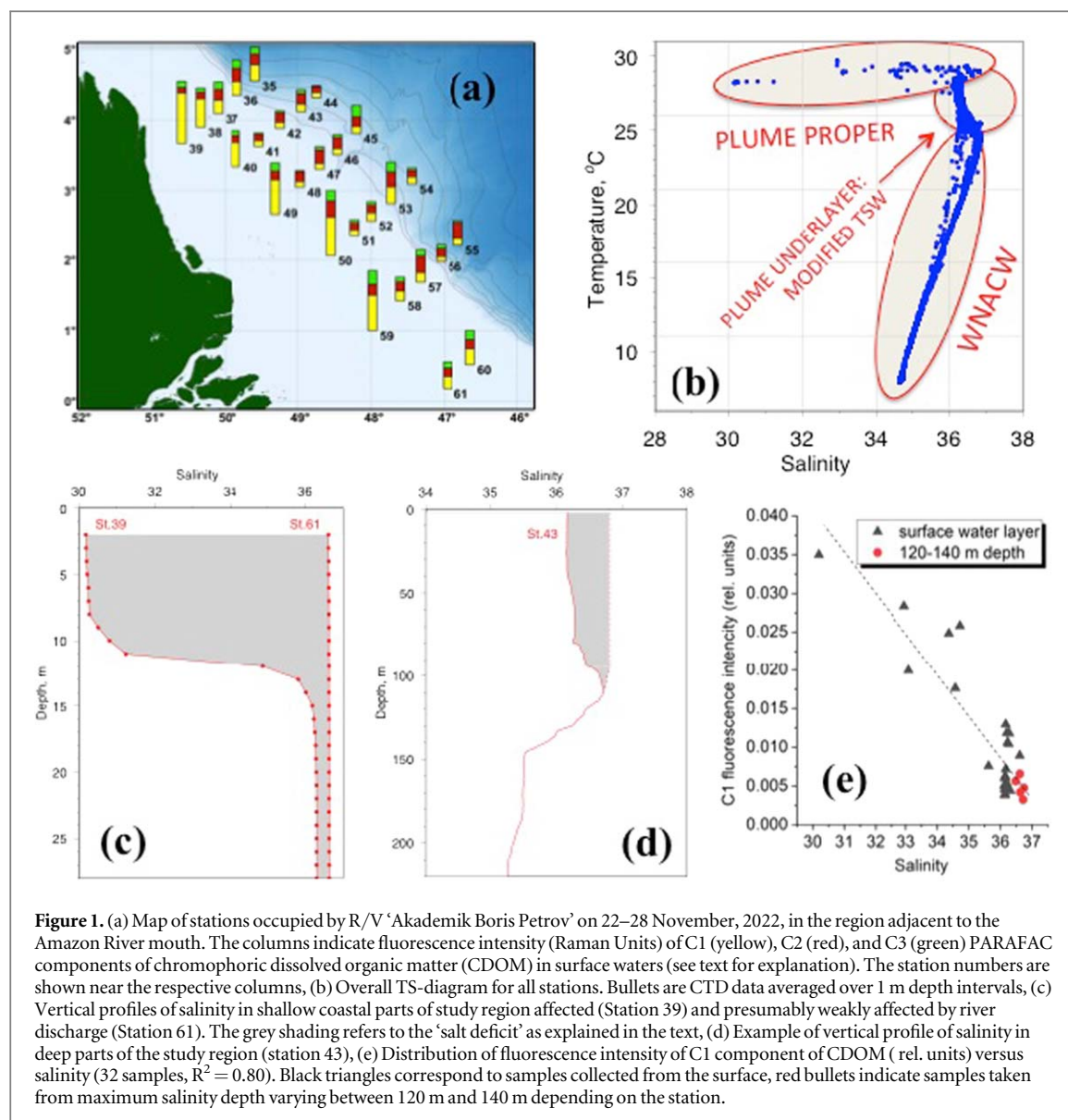
2. Data and methods

The data used in this study were collected in field survey of R/V ‘Akademik Boris Petrov’ from November 20 through November 29, 2022. The map of stations is given in figure 1(a). The location of stations and arbitrary limit bathymetric depths > 28 m were pre-approved by the Brazilian authorities. The measurements at 28 stations were carried out using a CTD profiling of the upper 200m and water sampling with Niskin bottles. An *AML Oceanographic BaseX* CTD probe was used for these measurements. The CTD data were averaged into 1 m depth bins. The LADCP data in 8 m depth bins for the upper 200 m measured by *TRDI WorkHorse Monitor* 300 kHz profiler were obtained only at some stations, namely, stations 35, 46, and 54–56.

At each station, we also collected water samples from different depth levels using Rosette with 5 l Niskin bottles. All samples (27 samples collected from the surface and 5 samples from depth varying between 120 m and 140 m) were filtered through pre-combusted (450 °C) Whatman GF/F filters with a nominal pore size of 0.7 µm. Filtrate was collected into the acid-cleaned 40 ml glass vials and stored under dark conditions at 3 °C until further analysis. Fluorescence and absorbance of chromophoric DOM (CDOM) were examined with an *Aqualog* spectrofluorometer (*Horiba Jobin Yvon, Inc.*) in a 1-cm path-length quartz cuvette at room temperature relatively to a standard pure water sample. The EEM spectra were recorded by scanning the excitation range between 240 and 650 nm (5 nm intervals) and capturing emission spectra over 250 to 800 nm (1.2 nm intervals). Integration time for each sample was 10 s. Potential inner-filter effect has been corrected by multiplying the spectra by the effective absorption coefficient (Lakowicz 2013), although the resulting change did not exceed 5%.

Prior to the analysis, the scattering signal was removed, with widths of ±20 nm for the first order Rayleigh scattering, ±7 nm for the first order Raman scattering, and ±20 nm for the second order Raman scattering. The missing areas were then interpolated using Whittaker smoothing (Krylov and Labutin 2023) with non-smoothness penalties $\lambda_1 = 10^{-2}$, $\lambda_2 = 1$ and negativity penalty of 0.1 (figure S1 of supplemental information).

The PARAFAC model (Bro 1997) was used to decompose the fluorescence intensities into emission spectra, excitation spectra, and relative concentrations of independent fluorophores comprising the data (figure 1(a)). Split-half analysis (DeSarbo 1984) was performed on 10 randomly assigned halves, suggesting a three-component model. Additionally, fluorescence indices have been calculated for all samples through the procedure described by Krylov *et al* (2020). We focus on the humification index (HIX) defined as the ratio of

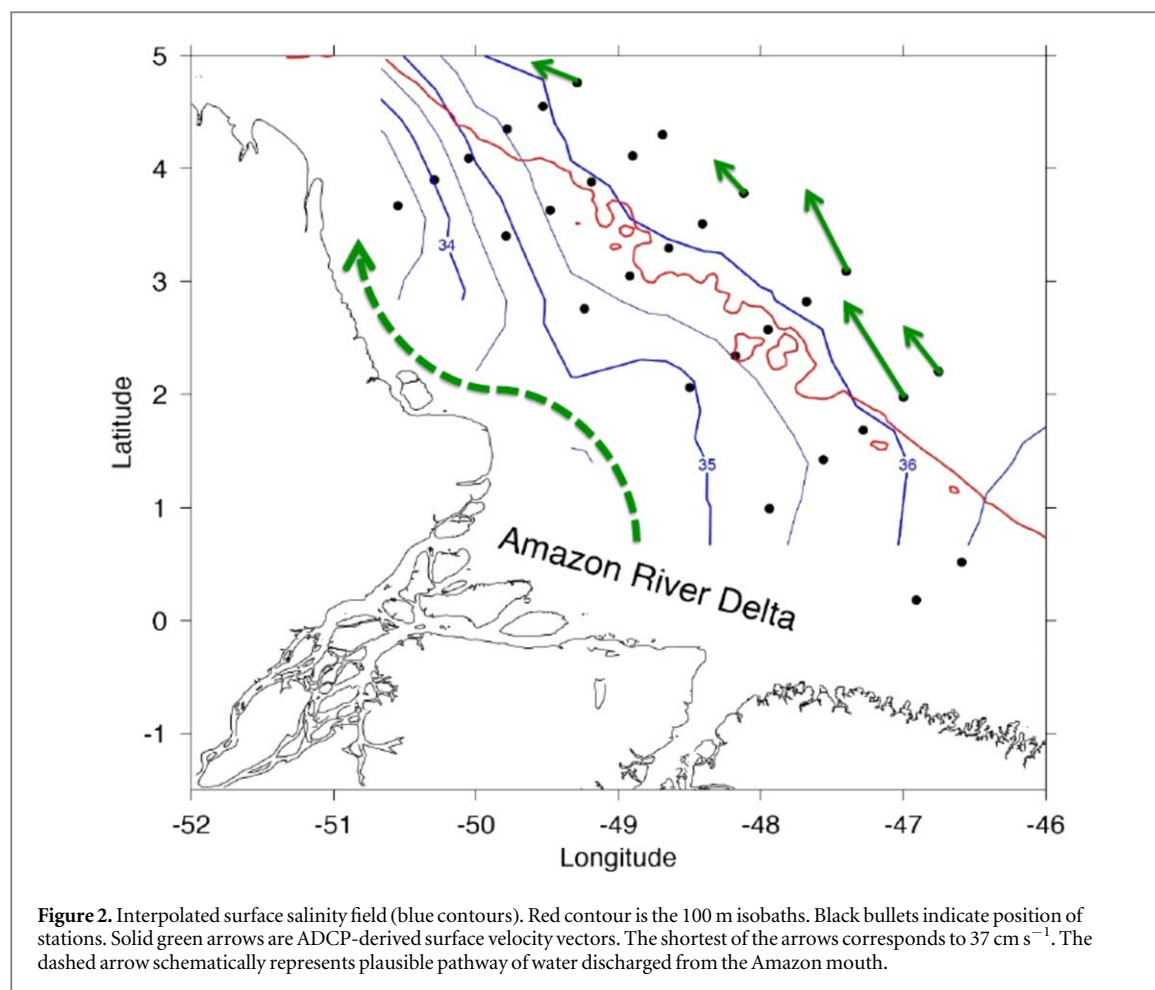


humic-like and protein-like fluorescence intensity excited at 245 nm, and the biological index (BIX) representing the ratio of fluorescence intensity at 380 nm to that at 430 nm when excited at 310 nm. Diagnostic values for important recent autochthonous DOM component are $HIX < 4$ and $BIX > 1$, while strong humic character often related to significant terrigenous contribution is characterized by $HIX > 16$ and $BIX < 0.7$ (Huguet *et al* 2009).

3. Results

Typical vertical profiles of salinity along with the overall TS-diagram are shown in figures 1(b)–(d). The surface salinity distribution interpolated from all stations is depicted in figure 2, which also exhibits the 100 m isobaths as a proxy for the shelf break position, and gives some insights about the circulation. The green arrows in figure 2 are surface velocity vectors actually measured at stations 35, 46, and 54–56. They indicate very energetic, up to 106 cm s^{-1} at station 56, flow generally directed to the northwest along the shelf break, presumably associated with the North Brazil Current. The currents on the shelf are unknown to us, but the salinity pattern suggests that the water discharged from the river was mainly transported in northwesterly direction along the coast, as shown schematically by the dashed arrow.

It is noteworthy that although some stations were located as close to the coast as the 28-m isobaths, salinity at all stations was above 30, and at most of them above 34. Still, the AR plume was evident in the data, manifested as



a pronounced decrease of salinity from well over 36 at the outer stations of the area covered by the measurements to below 35 at the shelf stations and down to about 30 at some near-shore locations.

The vertical thermohaline structure consisted of the upper mixed layer where both temperature and salinity reduced by river discharge were nearly uniform, or ‘plume proper’, whose thickness ranged from about 12 m at station 39 on the inner shelf to over 90 m at some deep stations. Further downwards, it was followed by what we will hereinafter call the ‘plume underlying layer’ (PUL), where salinity gradually increased with the depth and attained maximum at 110–140 m. Several previous studies conducted in the area pointed on the fact that the upper isothermal and isohaline layers generally did not coincide, and introduced the so-called barrier layer as the interface between the uniform salinity layer and the bottom of the isothermal layer (e.g., Pailler *et al* 1999). For the purposes of this study, the mixed layer depth was determined from thermal conditions as the depth where the difference between the local temperature and the temperature at 2 m depth exceeded 0.15°C , while the PUL was defined as the layer between the mixed layer depth and the maximum salinity depth. Hence, the so defined mixed layer would encompass also the barrier layer, if any. However, the barrier layer was not identifiable in this study, as it normally does not develop during low discharge season in boreal autumn (Silva *et al* 2005a).

Some previous publications identified the PUL as the Tropical Surface Water (TSW) layer and suggested that as the plume propagates in the surface layer, it somehow roofs the surrounding TSW—literally, as (Silva *et al* 2005b) put it, ‘in the areas of the plume influence, deepening of TSW takes place’ (translated from Portuguese). We note, however, that TSW in the absence of the plume is generally characterized by salinity either nearly constant or decreasing with depth (e.g., Liu and Tanhua 2021), while in the layer underneath the plume, salinity increased downwards. This can only be explained by the presence of a fraction of continental freshwater not only in the plume itself but also in the PUL where, being admixed to TSW, it modulates the vertical profile. We, therefore, interpret PUL composition as TSW modified by continental discharges, which provides the link between the plume proper and the undisturbed waters below the maximum salinity depth. Accordingly, we assume that the depth of salinity maximum is the lower limit of the freshwater influence. Further downwards, salinity decreased again in a pattern characteristic for the Western North Atlantic Central Water, hereinafter WNACW (e.g., Stramma and Schott 1999, Liu and Tanhua 2021).

As to the CDOM, the PARAFAC decomposition revealed three fluorescent components with one pronounced emission band λ_{em} and one or two excitation maxima λ_{ex} , namely, C1 with $\lambda_{ex} < 400$ nm and $\lambda_{em} = 447$ nm, C2 with $\lambda_{ex} < 240, 275$ nm and $\lambda_{em} = 330$ nm, and C3 with $\lambda_{ex} < 260, 293$ nm and $\lambda_{em} = 354$ nm (see figure S2 and table S1 of supplementary data). Component C1 is characterized by humic-like fluorescence (Coble 1996). It is abundant in rivers and lakes as well as in estuarine, coastal and shelf regions of the oceans. C1 is usually associated with a high impact of terrestrial DOM. Component C2 is also widespread in natural waters, indicating low molecular weight autochthonous DOM. The elevated protein-like fluorescence of C2 was also observed near urbanized areas, which can be explained by both the presence of pollutants with similar spectral characteristics and by an increase in the content of autochthonous material caused by the DOM decay (Parr *et al* 2015). Component C3 exhibited protein-like fluorescence similar to tryptophan amino acid (Coble 1996, Stedmon and Nelson 2015). It has been reported to be derived from microbial metabolism (Romera-Castillo *et al* 2011, Amaral *et al* 2016) and is often linked to aquatic productivity (Castillo *et al* 2010).

In the study area, HIX and BIX varied in the ranges 0.2–4.6 and 0.78–2.02 (table S2 of supplementary information), respectively, which indicates the predominance of the autochthonous CDOM (Zsolnay *et al* 1999, Huguet *et al* 2009, Derrien *et al* 2017). A decrease in the autochthonous DOM contribution in the surface water layer, along with an increase in the allochthonous CDOM component, was recorded at the southern stations of the transects closest to the Amazon delta where the HIX took its maximum, and the BIX—its minimum values. This area was characterized by salinities 30.2–35.6. The larger portion of allochthonous DOM is associated primarily with an increase in the content of humic substances, supplied by the AR runoff, in which humic substances constitute about 60% of the DOM (Ertel *et al* 1986). The distribution of the content of humic substances can be examined from the fluorescence intensity of the humic-like component C1 (figure 1(a)). Conservative behaviour of the C1 component was observed in the studied salinity range ($R^2 = 0.80$, 32 samples) (figure 1(e)), which supports the conclusion about terrestrial origin of C1. Although HIX and BIX indicate the general importance of autochthonous DOM, the terrestrial-derived C1 component often dominates in terms of reduction of the sum of squared residuals due to prevalence of C1 and/or since its peaks being wider than those of the remaining two components, it covers broader area than the narrower peaks of C2 or C3.

One of our objectives was to estimate the total content of continental fresh water within the part of the plume covered by the measurements. To this end, we first calculated the ‘salt deficit’ D for each station as

$$D = \int_{z_{\max}}^0 (S(z) - S_0(z)) dz, \quad (1)$$

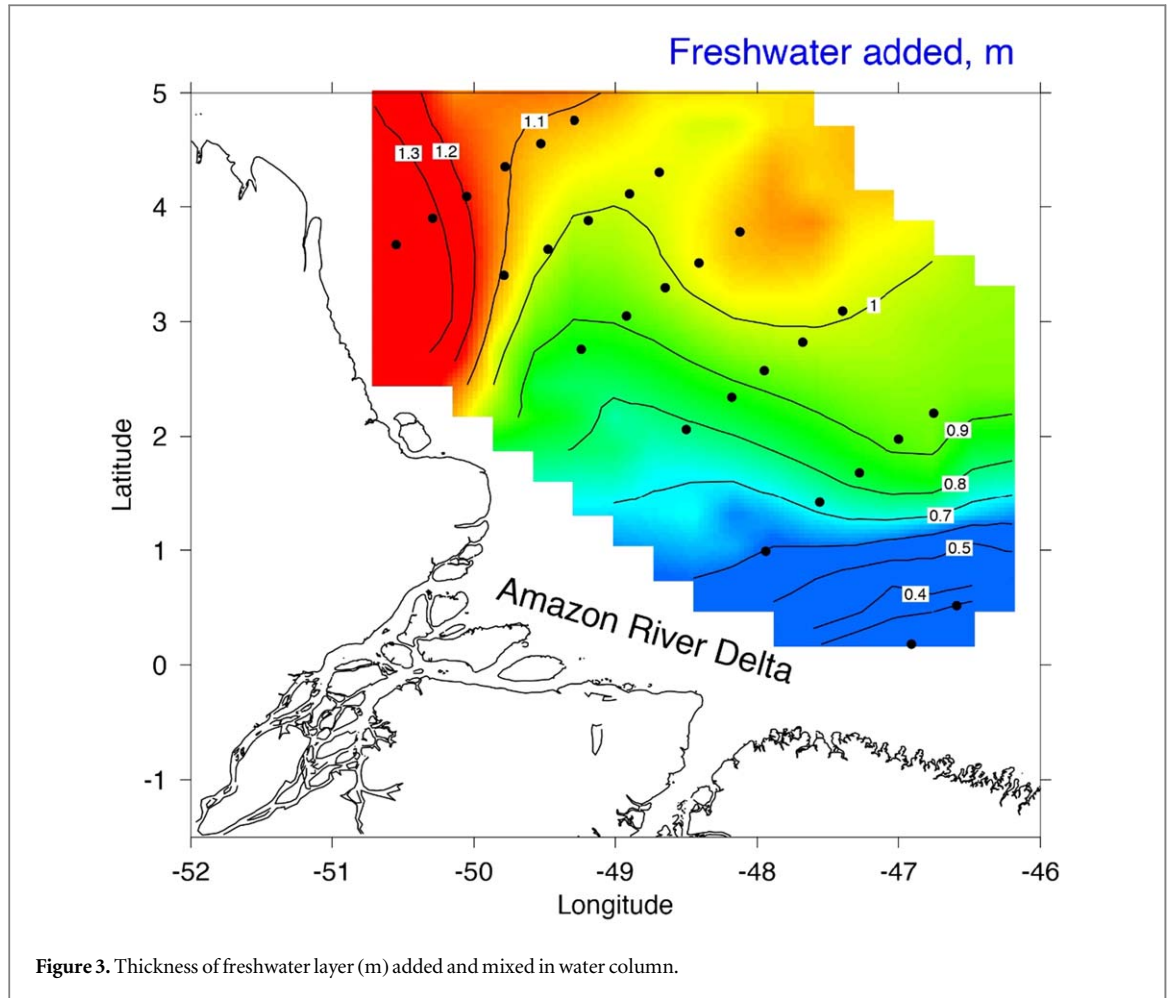
where $z_{\max}$ is the depth of salinity maximum (or the total depth for shallow stations with the depth smaller than the depth of salinity maximum), S is salinity, z is vertical coordinate, and $S_0(z)$ is a hypothetical undisturbed profile in the absence of the plume and the influence of river discharges. Based on the CTD data obtained at the southern extremity of the study area where the freshwater influence was at its minimum, and taking into account literature data, we adopted the following idealized profile for $S_0(z)$:

$$S_0(z) = \begin{cases} 36.7, & z < 95, \\ S(z_{\max}) + \frac{(36.7 - S(z_{\max}))}{(z_{\max} - 95)} (z_{\max} - z), & 95 < z < z_{\max}, \\ S(z), & z > z_{\max}, \end{cases} \quad (2)$$

that is, the idealized ‘would-be-undisturbed’ salinity profile corresponding to an observed ‘disturbed’ one is approximated by a constant value of 36.7 in the upper 95 m, then salinity linearly increases downwards to where the salinity maximum was actually observed, and below it coincides with the observed profile, supposedly undisturbed in this part of the water column. Of course, the shape of the undisturbed profile selected for this analysis is only hypothetical and idealized, but it fits reasonably well the observations at stations outside the plume’s influence or on its periphery (Silva *et al* 2005a, 2005b). In addition, the upper layer salinity value of 36.70 not only lies very close to the upper layer salinity that we observed at the virtually ‘undisturbed’ station 61, namely, 36.68, but also fits well into the typical salinity range of WNACW, i.e., between 36.64 and 36.82, or 36.73 on average (Liu and Tanhua 2021). Hu *et al* (2004) used the interval from 36.47 to 36.80 (36.72 on average) as the ‘marine end-member salinity’, as they put it, for this region. Examples of undisturbed and disturbed profiles for shallow and deep stations are shown in figures 1(c) and (d), where the grey shading indicates the salt deficit.

Equation (1) yields the salt deficit in $\text{kg} \cdot \text{m}^{-2}$. This quantity has a simple sense of the total mass of salt that needs to be added to each m^2 of the sea surface to restore the salt content that would have been there without freshening by river discharges. A map of the salt deficit calculated as described above (not shown here for the sake of brevity) revealed that this quantity spanned from nearly zero at the southern extremity of the study region, where the influence of river discharge was the smallest, to over $60 \text{ kg} \cdot \text{m}^{-2}$ on the inner shelf in the northern part of the region in a plume-like pattern.

The salt deficit D can be converted into the thickness H of the layer that must have been removed from the salty water column and replaced by the river-borne freshwater layer added and mixed into the affected water column whose height is $z_{\max}$ simply as



$$H = z_{\max} \frac{D}{Q}, \quad (3)$$

where $Q = \int_{z_{\max}}^0 S_0(z) dz$ is the total salt content in the ‘undisturbed’ water column. A map of H expressed in meters is depicted in figure 3, with H varying from nearly zero in the southern part of the study area outside the plume to 1.5 m in the northern nearshore corner, where the salt deficit was the maximum.

With this done, we then obtained an estimate for the total content Q of riverine water in the area covered by our measurements. We interpolated the H values computed for the stations onto a regular rectangular grid between 46 and 51°W, and 0 and 5°N using the objective interpolation method described in Koch *et al* (1983) and Levy and Brown (1986). We then integrated H by area by summing up the values of H at all grid nodes filled by interpolation multiplied by grid element area. This procedure arrived at the following value:

$$Q = 203 \pm 22 \text{ km}^3. \quad (4)$$

Here, the error interval was estimated by varying the grid size and the number of grid nodes used for interpolation across the range recommended by (Koch *et al* 1983), in this case, from a minimum of 25 to a maximum of 100.

We also applied the abovementioned procedure to the part of the water column below the upper mixed layer, i.e., the PUL only. To this end, we replace the upper integration limit in equation (1) with the mixed layer depth z_{ml} to obtain

$$D_{pul} = \int_{z_{\max}}^{z_{ml}} (S(z) - S_0(z)) dz, \quad (5)$$

and

$$P = \frac{D_{pul}}{D} \cdot 100\%, \quad (6)$$

where D_{pul} is the salt deficit within the plume-underlying layer, and P is its percentage ratio with respect to the total salt deficit. The values of P ranged from 12% to 54% for different stations, averaging over the entire study area to 37%. Accordingly, the same ratios apply to the relative contents of fresh water in the respective layers.

4. Discussion

Although our measurements by far did not cover the entire area of the plume, the collected data suffice to conclude that both the cross-shore size of the plume (about 200 km or less from the river mouth, if we consider the isohaline 36 isoline as the plume's outer limit, as suggested in some previous publications, e.g., Salisbury *et al* 2011) and the salinity drop magnitude (a drop of only 6 at the cross-shore distance of 200 km from the mouth) are moderate, given the huge scale of the freshwater source. The plumes of other major estuaries of the World, such as the Ob, the Yenisei, or the Rio de la Plata, for examples, typically exhibit much stronger salinity drops of up to 20 at same or greater distances from their estuaries (e.g., Zavialov *et al* 2003, Zavialov *et al* 2015, Osadchiev *et al* 2021). As mentioned above, the discharge rates of the AR, even at their yearly minimum, still greatly exceed those of these rivers, and yet, the plumes that the latter develop are more pronounced than the AR plume, as observed here. We recall that it has been demonstrated in numerical modeling experiments by (Osadchiev and Zavialov 2013) that, under otherwise equal conditions, river plumes in equatorial regions attain smaller horizontal extents than those in moderate latitudes. This effect might be of relevance to the development of the AR plume originating from the river mouth situated exactly at the Equator.

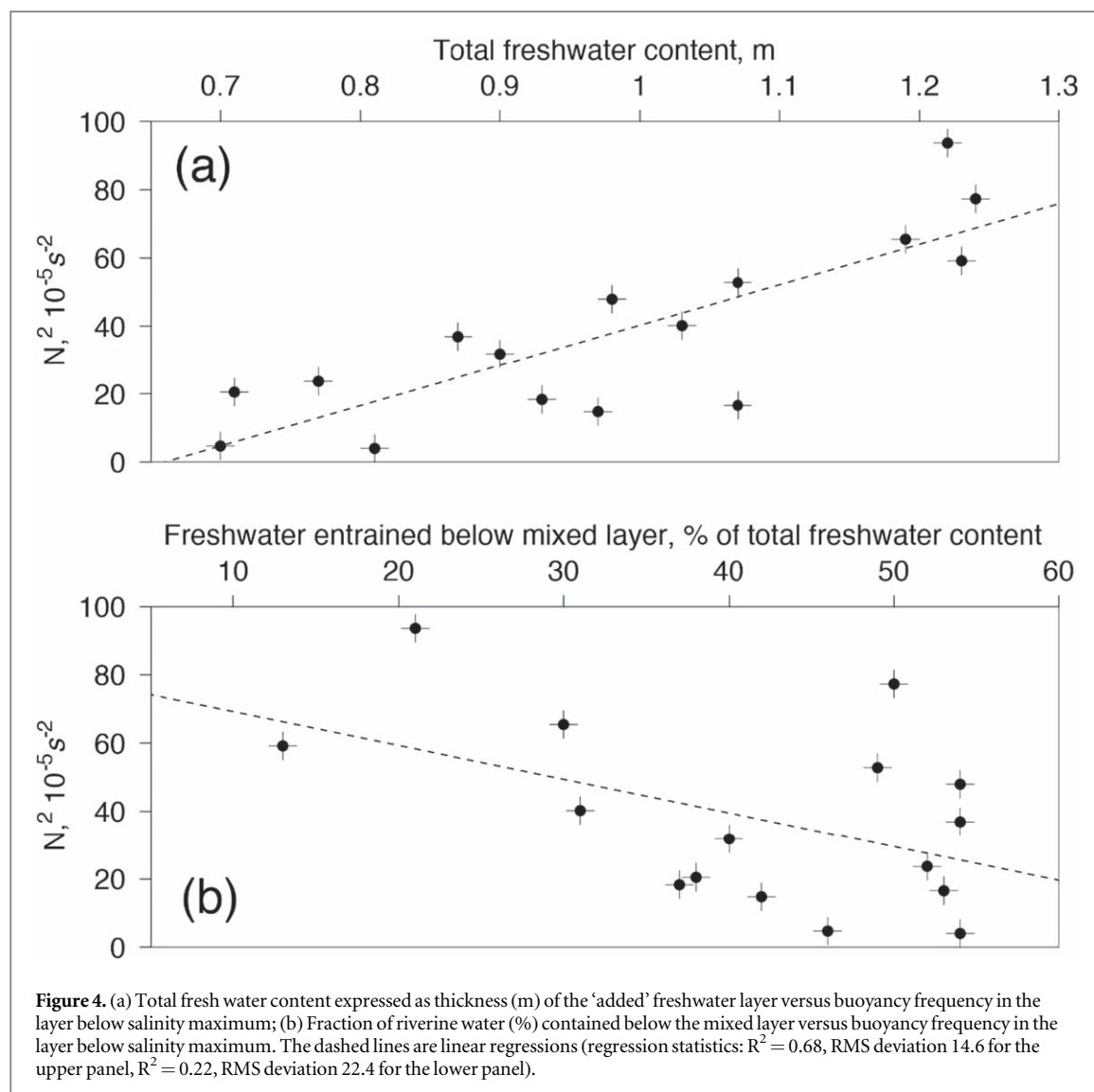
To our knowledge, estimates of the total volume of continental freshwater stored in the plume and its specific parts or layers have not been made yet, except the work by Hu *et al* (2004), where some integrated estimates of the freshwater content in the area of Amazon-Orinoco influence were obtained based on the data from remote sensing. These authors reported the total volume of 800–1200 km³ for the month of November. However, this refers to the entire area between 0–15°N and 45–62°W, i.e., almost 7-fold the area of the present study.

The volume of freshwater accumulated in the area covered by the measurements made in the present study, as estimated above (about 200 km³), is relatively small, considering that it corresponds to less than 3% of the Amazon's annual discharge. For comparison, a similar estimate for the Yenisei plume in the Kara Sea would arrive at 30% at least (Zatsepin *et al* 2010). Of course, the area covered by this Amazon plume survey by no means encompassed the entire plume, as the latter is likely to extend farther northwestward. Besides, the most freshened coastal part of the plume located on the inner shelf at depths below 28 m remained unsampled in this campaign. However, this near-shore area is shallow and, hence, may contain only limited volume of freshwater.

We also emphasize here the presence of freshening not only in the upper mixed layer (dubbed 'plume proper' in figure 1(b)) but also underneath it in a rather thick intermediate layer ('plume's underlying layer', PUL) between the mixed layer and the salinity maximum. This means that a part of fresh water must have been entrained from the mixed layer into the PUL. The percentage of fresh water in the PUL relative to the total freshwater content varied from 12% to 54% at different stations (figure 4(b)) and totalled to 37%, or about 76 km³, over the entire area sampled. Below the salinity maximum, salinity decreased downwards in a pattern characteristic of WNACW. Such a three-layered vertical structure is not typical for the abovementioned large rivers whose plumes usually occupy only the mixed upper portion of the water column with a more or less abrupt jump of salinity at its lower boundary and a layer of undisturbed high salinity seawater immediately underneath it. Accordingly, the Amazon's plume, with its maximum depth of freshwater influence up to 140 m, is significantly 'thicker' than those of the abovementioned other major rivers where such influence is generally confined to the upper 10–20 m (e.g., Zavialov *et al* 2003, Zatsepin *et al* 2010, Osadchiev *et al* 2021). This indicates enhanced vertical mixing of fresh water and might explain the moderate cross-shore spatial extent of the plume and relatively small salinity drop in it.

In the context of vertical mixing, it is interesting to look at the freshwater content in the plume and its distribution between the upper mixed layer and the PUL versus the background stratification - not in the plume layer itself, but in the 'undisturbed' background part of the column underneath the plume. The Brunt-Väisälä (buoyancy) frequency scatter plots, calculated for the 20 m thick layer immediately below the maximum salinity depth at all stations having salinity maximum against the total freshwater content and a fraction of riverine water (%) contained below the mixed layer are shown in figure 4. There is a relatively high ($R^2 = 0.68$ for 16 data points) positive correlation between the freshwater content and stratification in the 'undisturbed' layer, indicating that the stronger the background stratification is preventing the plume from mixing downwards, the more massive the accumulation of riverine water is within the plume (figure 4(a)). There is also a less pronounced but significant negative correlation between the stratification and the percentage of freshwater entrained into the underlying layer, meaning that the smaller the background stratification, the more intense the plume's freshwater proper entrainment into the PUL (figure 4(b)). Altogether, these data suggest that vertical mixing plays important role in shaping the plume.

The idea of enhanced vertical mixing entraining continental waters rather deep down to the maximum salinity depth is also supported by CDOM analysis, which demonstrates the presence of clearly terrestrial humic substances (C1 component) even at the depths of up to 140 m, where their concentrations are comparable with those in the samples taken from the surface layer. For example, at station 35, the terrigenous C1 concentration at



140 m depth reached 89% of its mean concentration in surface waters of the nearby transect (stations 40–44), and 60% of its mean concentration of all surface water samples. The samples collected from depths over 110 m characterized by higher salinity values still exhibited C1 concentrations comparable to those from surface waters. This fact can be related to both formation of deep waters during the period of higher Amazon River discharge characterized by higher DOC concentration (Seidel *et al* 2016) and faster photodegradation of humic-like CDOM in the surface layer (Cao *et al* 2016).

5. Conclusions

Based on *in situ* CTD and CDOM data collected in a field survey in the plume of the AR in November of 2022, we note a modest cross-shore extent of the Amazon plume and a relatively small salinity drop in it. Apart from the low discharge season, it can be explained by the fact that freshwater influence was evident in the very thick, up to 140 m, upper portion of the water column, including the plume proper in the upper mixed layer and the plume-underlying layer extending from the mixed layer to the maximum salinity depth. This plume-underlying layer accumulated about 37% of the total freshwater volume in the study area, along with significant content of terrigenous dissolved organic matter. The total freshwater content accumulated at individual stations was positively correlated with the background stratification below the layer of freshwater influence, while the percentage of freshwater stored in the plume-underlying layer was negatively correlated with the stratification (parameterized as buoyancy frequency), thus pointing on importance of vertical mixing in shaping the plume. Furthermore, by comparing the observed vertical salinity profiles disturbed by continental discharges with the idealized undisturbed ones, we estimated the total river-borne fresh water volume in the study area (46–51°W,

0–5°N) as $203 \pm 22 \text{ km}^3$, which is less than 3% of the average annual Amazon River discharge. It suggests rapid removal of freshwater from the plume due to vertical and lateral mixing and advection.

Data availability statement

The data that support the findings of this study are openly available from the authors at request.

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