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The Seasonal cycle of the eastern boundary currents of the North Atlantic Subtropical Gyre

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Key Points:

- The Canary Current varies its path and strength seasonally, moving westward from winter to fall and being the strongest in fall.
- At the Lanzarote Passage, the flow is southward throughout the year except in fall when it reverses at the surface and intermediate layers.
- The Lanzarote Passage seasonal cycle and its amplitude correlate with the seasonal cycle of the AMOC as measured by the RAPID-MOCHA array.

Keywords: Canary Current, seasonal cycle, Atlantic Meridional Overturning Circulation, Eastern boundary, Lanzarote Passage, North Atlantic Subtropical Gyre.

Abstract

For the first time, four dedicated hydrographic cruises – one in each season – took place in 2015 around the Canary Islands to determine the seasonality of the flows at the eastern boundary of the North Atlantic Subtropical Gyre. The Canary Current (CC) is the eastern boundary current of the North Atlantic Subtropical Gyre and links the Azores Current with the North Equatorial Current. The 2015 estimations show that the CC has a seasonal behavior in its path and strength, flowing on its easternmost position in winter (3.4±0.3 Sv), through the Canary Islands in spring (2.1±0.7 Sv) and summer (2.0±0.6 Sv) and on its westernmost position in fall (3.2±0.4 Sv). At the Lanzarote Passage (LP), the dominant flow is southward except in fall, where a northward transport is observed at the surface (1.1±0.3 Sv) and intermediate (1.3±0.2 Sv) layers. Combining all the available transport estimations a historical composite observational seasonal cycle is constructed and fits the 2015 seasonal cycle. Hence, a solid seasonal cycle is constructed supported by all the available observations in the area. The LP seasonal cycle and seasonal amplitude match

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the seasonal cycle of the Atlantic Meridional Overturning Circulation (AMOC) measured by the RAPID-MOCHA data array. These indicates that the seasonal cycle of the AMOC of the RAPID-MOCHA data array is driven by the dynamics of the eastern boundary currents.

Plain Language Summary

In this study, we explore for the first time the seasonal cycle of the main ocean currents existing at the eastern side of the North Atlantic Subtropical Gyre using data from four cruises that took place in each season in 2015. The Canary Current (CC) is the southward flow that connects the Azores Current with the North Equatorial Current. The 2015 estimations show that CC shifts westward from winter to fall. In addition, it also changes transport seasonally being the weakest in spring and summer, and the strongest in fall. The Lanzarote Passage (LP) is the geographical area between Lanzarote and Africa, it is considered to be the eastern end of the North Atlantic Subtropical Gyre. Through the LP the flow is mainly southward except in fall when the flow reverses. A composite seasonal cycle arises from combining all the available observations with the outputs of this study. The LP seasonal amplitude and cycle match the ones of the Atlantic Meridional Overturning Circulation (AMOC) measured by the RAPID-MOCHA data array. These indicates that this seasonal cycle of the AMOC is driven by the dynamics of the eastern boundary currents.

1. Introduction

At the eastern region of the North Atlantic Subtropical Gyre, the Canary Current (CC) connects the Azores Current with the North Equatorial Current (Hernández-Guerra et al., 2005; Pérez-Hernández et al., 2013) (Figure 1a). The first studies carried out in the region used historical hydrographic data to determine the existence of a seasonal change in the structure of the eastern subtropical gyre. These studies found a CC flowing closer to the African coast in summer and through the western islands in winter (Stramma & Isemer, 1988; Stramma & Müller, 1989). Using one hydrographic cruise per season carried out between Madeira and north of the Canary Islands (at 28-32ºN) in 1997 and 1998, Machín et al. (2006) reported that the CC presents a seasonal cycle characterized by a southward mass transport of 1.7± 1.0 Sv in winter, 2.8± 1.0 Sv in spring, 2.9±

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1.1 Sv in summer, and 4.5±1.2 Sv in fall with a shift westward from spring to fall. Pérez-Hernández et al. (2013) confirmed that CC migrates west of the Canary Islands in fall and described a mass transport of 5.8±0.2 Sv southward. In addition, both Fraile-Nuez & Hernández-Guerra (2006) using Argo data, and Mason et al. (2011) using a Regional Ocean Modeling System (ROMS) found that the variability of the CC is mainly driven by the curl of the wind stress, following the Sverdrup balance.

The area between the islands of Lanzarote and Fuerteventura and the African Coast, known as the Lanzarote Passage (LP) has a specific dynamic (Figure 1b). Several studies, (Hernández-Guerra et al., 2003; Machín et al., 2006; Pérez-Hernández et al., 2015 and Casanova-Masjoan et al., 2020), have reported that the seasonal variability across the LP is different from the variability west of the LP. Throughout the year, the flow across the LP is southward except during fall when it reverses carrying a mean northward transport of about 2-3 Sv of North Atlantic Central Water (NACW). This northward flow in fall has been attributed, using satellite data as a proxy -high correlations between the NACW transports and those from altimetry where found in Vélez-Belchí et al, (2017)- to a recirculation of the CC (Pérez-Hernández et al., 2015 and Vélez-Belchí et al., 2017). In addition, Casanova-Masjoan et al. (2020) and Hernández-Guerra et al. (2017) have shown, with in-situ observations, that a branch of the CC feeds the northward flow at the LP in fall.

The northward flow in surface and thermocline layers within the LP in fall is usually accompanied by a flow at intermediate levels that provides a higher Antarctic Intermediate Water (AAIW) contribution in the area (Fraile-Nuez et al., 2010). This intermediate flow has been called the intermediate Poleward UnderCurrent (iPUC in Pérez-Hernández et al., 2015), or Canary intermediate Poleward Undercurrent (CiPU in Vélez-Belchí et al., 2021, and used hereafter), and contributes nearly 1 Sv to the overall northward transport (Hernández-Guerra et al., 2005, 2017; Laiz et al., 2012; Pérez-Hernández et al., 2015; Vélez-Belchí et al., 2021). For these intermediate layers, Machín et al. (2010) attributed the northward flow during Fall to an isopycnal stretching generated by wind forcing further south. On the other hand, Vélez-Belchí et al. (2021) proposed that the existence of this northward flow at intermediate levels is due to a meridional alongshore pressure gradient generated by the density difference between the Mediterranean Water (MW) and the AAIW.

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The Atlantic Meridional Overturning Circulation (AMOC) is an important component of the climate system since it makes the largest oceanic contribution to the meridional transport of heat (Ganachaud, 2003; Caínzos et al., 2022). The strength of the AMOC is continually monitored along 26.5°N by the U.K.–U.S. Rapid Climate Change–Meridional Overturning Circulation and Heatflux Array (hereafter the RAPID-MOCHA array). Chidichimo et al. (2010) found, with the first 4 years of data, a peak-to-peak seasonal cycle of the AMOC of 6.7 Sv, and attributed 5.2 Sv of this seasonal cycle to the eastern boundary. From the three main components of the AMOC at 26.5°N - the Gulf Stream (T_{GS}), Ekman (T_{EK}), and upper-mid-ocean (T_{UMO}) transports - Kanzow et al. (2010) found that T_{UMO} is the largest contributor to the seasonal cycle of the AMOC, with a peak-to-peak seasonal amplitude of 5.9 Sv. Pérez-Hernández et al. (2015) compared the seasonality of the RAPID-MOCHA array to that of the EBC4 mooring located at the LP (Figure 1) and found significant correlations between the T_{UMO} and both the upper (0.7 correlation coefficient) and intermediate (0.8 correlation coefficient) layers of the Lanzarote Passage. This relation is further explored using hydrographic cruises that took place in different years during the spring and summer seasons. Vélez-Belchí et al. (2017) and Casanova-Masjoan et al. (2020) showed a seasonal amplitude of approximately 4 Sv that matches that of the AMOC for the fall-2013/spring-2014 (Vélez-Belchí et al., 2017) and fall-2016/spring-2017 cruises (Casanova-Masjoan et al., 2020b), while a larger amplitude of nearly 7 Sv was observed for the fall-2017/spring-2018 estimation(Casanova-Masjoan et al., 2020).

The objective of this paper is to describe the seasonal variability of the circulation in the eastern region of the North Atlantic Subtropical Gyre at the latitude of the Canary Islands by using four identical cruises carried out in each season and, in the same year, 2015. The remainder of the paper is organized as follows. In section 2, the different data sets are described. In section 3, we next describe the water masses existing in 2015. Section 4 presents the inverse box model. Section 5 shows the seasonal circulation around the Canary Islands estimated using the geostrophic approach and the inverse model applied to the hydrographic data. Section 6 discusses the seasonal cycle of 2015 and its amplitude. Section 7 introduces the seasonal cycle of the different AMOC components. Section 8 discusses the results with previous publications to create a composite seasonal cycle from historical data. Section 9 compares the AMOC seasonality with the one described here and finally, Section 10 exposes the main conclusions of the manuscript.

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2. Data

2.1 Hydrographic data

Since 2003, the *RaProCan* project, which is the Canary Islands component of the Spanish Institute of Oceanography (IEO) ocean observing system (Tel et al., 2016; Vélez-Belchí et al., 2015), monitors the Canary basin. In 2015, the *RaProCan* project joined efforts with the *Seasonal Variability of the AMOC: Canary Current (SeVaCa*n) project of the Instituto de Oceanografía y Cambio Global (IOCAG-ULPGC) to increase the temporal resolution of the observations. Thanks to this, in 2015, a hydrographic cruise took place in each season (February, April, July, and November) to study the seasonal cycle of the basin (Table 1 and Figure 1b). These 2015 cruises had a box-shaped track around the archipelago consisting of 51 stations divided into 24, 6, and 21 stations for the northern, eastern, and southern transects, respectively (Figure 1b). The northern transect of this box samples the 18 standard *RaProCan* stations (stations 6 to 24 in Figure 1b, red color). In each station, conductivity, temperature, and pressure were measured with redundant temperature and salinity sensors from a Seabird 911+ CTD, which were calibrated at the SeaBird laboratory before the cruise. Onboard salinity calibration was carried out with a Guildline Autosal model 8400B salinometer with a precision better than 0.002 for single samples. Density will be given as neutral density following Jackett and McDougall (1997). At each station, velocity data were acquired from a Lowered Acoustic Doppler Current Profiler (LADCP) system composed of a 150 kHz LADCP downward-looking (master) and a 300 kHz LADCP upward-looking (slave), with a shared battery pack. The LADCP data were processed according to Fischer and Visbeck (1993). Data were acquired at each station from the surface down to 10 m above the bottom. Distance intervals between stations were approximately 50 km except for the African slope stations, which were 4-5 km apart. The topographic data used in Figure 1 is from the Smith and Sandwell (1997) database.

2.2 Surface data.

Maps of Absolute dynamic topography (MADT) were obtained from the Space Oceanographic Division of the Collective Localization Satellites (CLS) through the Archiving, Validation, and Interpretation of Satellite Oceanographic Data project (AVISO; https://www.aviso.altimetry.fr/en/home.html). The MADT is a merged product from all available Absolute Dynamic Topography (ADT) data from TOPEX/Poseidon, Jason-1, Jason-2, Envisat,

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and GFO satellites. The MSLA data have a temporal resolution of one day and are gridded in 0.25º×0.25º spatial bins on a Mercator grid.

Wind data are estimated using the Weather Research and Forecasting (WRF) model (version 3.9.1), developed at the National Center for Atmospheric Research. This model has the advantage of obtaining wind data in high temporal and spatial resolution to resolve the orographic perturbation of the wind as it flows through the islands and the wind variability during the cruise. A complete description of this model can be found in Skamarock et al., (2008). Data from the operational analysis performed every 6 hr, at 1° horizontal resolution at the National Center for Environmental Prediction (NCEP final analysis) were used as initial and boundary conditions for the simulations. For this study, we have set a horizontal grid spacing of 0.125° and 50 terrain-influenced vertical levels. A full description of the configuration of the simulations is given by Cana et al. (2020). The model output covers our period of sampling and includes the zonal and meridional wind velocities measured at 10 m (U10 and V10, respectively). Both are used to estimate the Ekman transport to be included in the shallowest layer of the inverse box model.

3. Water Masses

Surface waters (SW, γ^n <26.85 kg m^{-3}), as they are in contact with the atmosphere, show the largest seasonal signal being the warmest during fall (reaching up to 25.9ºC) when the trade winds stop, and the coolest during winter (reaching up to 19.0ºC). This is especially remarkable in the open ocean regions (Figure 2 a, b, and c). Immediately beneath this layer, the NACW expands on the density range 26.85 < γ^n < 27.38 kg m^{-3}, following the nearly straight line described in Harvey & Arhan (1988) in all seasons and geographical areas. SW and NACW compose the thermocline layers of our study.

At intermediate levels (27.38< γ^n <27.82 kg m^{-3}), the water mass is a mixture of AAIW and MW (Hernández-Guerra et al., 2005; Machín et al., 2006). Interestingly, the content of AAIW/MW varies seasonally and geographically as seen in Figure 2. On the northern transect, winter high salinities indicate the presence of a Meddy (Mediterranean Eddy) which is a dynamical structure often seen in the Canary Basin as anticyclonic eddies containing high amounts of MW in their core (Machín & Pelegrí, 2016) (Figure 2a). In all the areas the intermediate level during winter is the most homogeneous, except at the northern transect where the Meddy is found. In both the northern

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and western transects, the intermediate levels do not present any significant seasonality (Figures 2a and b). On the southern transect, the intermediate layers evolve from presenting a high MW content in summer to the low salinities of AAIW in spring. At the stations sampled along the LP and African shelf (Figure 2 d and e), a clearer transition from MW in winter to AAIW in fall is observed. The presence of AAIW along the African coast in fall has been widely documented (Fraile-Nuez et al., 2010; Hernández-Guerra et al., 2017; Pérez-Hernández et al., 2013; Vélez-Belchí et al., 2017, 2021).

The deepest water mass corresponds to North Atlantic Deep Water (NADW, $\boldsymbol{\gamma^n}$ >27.820 kg m^{-3}), which is only seen on the open ocean transects (Figure 2a to c). This water mass can only be observed in the oceanographic stations deeper than ca. 1500m (the Northern, Western, and Southern transects) and it does not present any seasonality in our dataset.

4. Geostrophic Velocities.

To describe the seasonal change in the ocean circulation in the eastern boundary of the North Atlantic Subtropical Gyre, we have initially estimated the geostrophic velocities using the thermal wind equation with a level of no motion. Here, the density level γ^n=27.975 kg m^{-3} (roughly 1950 m; Table 2) located at the interface between the MW and the NADW is used for the oceanic waters, while the density level of γ^n=27.380 kg m^{-3} (roughly 750 m; Table 2) found between the NACW and the AAIW is used for the stations located in the Lanzarote Passage as previous studies carried out in the area (Hernández-Guerra et al., 2005). The geostrophic velocities were integrated over 13 neutral density layers following the water mass characterization given in Section 3 (Table 2). The thermocline waters occupy the first four layers, the intermediate water masses the next two, and the deep-water masses are found in the densest layers corresponding to NADW.

Once the initial geostrophic velocities described above were estimated, an adjustment is carried out in two steps. First, following Comas-Rodríguez et al. (2011) and using the LADCP data, the velocities were adjusted on a station-by-station analysis. This analysis consisted of selecting the vertical range of the LADCP profile where the vertical shear resembles that of the initial geostrophic velocities, and, then, computing a mean of those vertical LADCP velocities. Once these reference velocities are added, the new geostrophic transports are computed, but still do not accomplish mass balance for each hydrographic cruise, as shown in Table 3. Second, to

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reduce the mass transport imbalance obtained using the thermal wind equation and LADCP reference velocities, and therefore increase the reliability of the mass transport estimates, we use an inverse box model. An inverse box model is based on the conservation of mass and allows the estimation of new velocities at the reference level once adjusted to LADCP data. Following Hernández-Guerra et al. (2005) and Pérez-Hernández et al. (2013), we have applied an inverse model to the volume enclosed by the hydrographic stations and the African coast. The box model includes the conservation of mass per layer, the total, and an adjustment for the initial Ekman transport:

$$\iint \rho b \, dx \, dz = -\iint \rho V_{rel} \, dx \, dz + E_k \quad (1)$$

where x and z are the along the transect and vertical coordinates, respectively; ρ is the density of each layer. The integral terms are derived from the reference velocity (b) and the relative geostrophic velocity adjusted to LADCP data (V_{rel}). The term E_k designates the Ekman transport, wich is calculated for each transect and cruise, ranging between –0.32 Sv to 0.36 Sv. The inverse model slightly adjusts these Ekman transports.

Once discretized, the equations of mass transport per layer and the total mass transport form the following matrix equation:

$$A\,x + n = -\Gamma \quad (2)$$

where A is a matrix with the number of layers(Q)×stations(N), n is a column vector whose elements are the noise for each equation, Γ is a vector representing the degree of initial imbalance in each layer, and x is the column vector containing the unknowns of the system (one row for the velocity at the reference level for each station pair and the last one for the adjustment of the Ekman transport):

$$x = \begin{pmatrix} (b_i), i = 1, \dots N_{pair} \\ \Delta T_{Ek} \end{pmatrix} \quad (3)$$

To solve the inverse problem, we applied the Gauss-Markov method which produces a minimum error variance solution from the initial estimates of the unknowns (Wunsch 1996). The solution provided by the method depends on the a priori variances of each component. For mass transport, we chose $(0.1\ \text{Sv})^2$ for each layer, $(1\ \text{Sv})^2$ for the total, and for the velocities $(0.02\ \text{m/s})^2$ for the open ocean stations (>1500 m) and $(0.04\ \text{m/s})^2$ for stations located in shallower waters. For

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the winter cruise, the a priori variance of the mass transport is slightly different, being $(1\ Sv)^2$ on the first layer, $(0.5\ Sv)^2$ in layers 2 to 7, $(0.25\ Sv)^2$ for the remaining layers, and a $(2\ Sv)^2$ for the total. These a priori variances have been extensively used in the Canary Basin (Hernandez-Guerra et al., 2017; Casanova-Masjoan et al., 2020).

The main outcomes of the inverse modeling are a new set of velocities at the reference level that help to achieve mass balance inside the box. As seen in Figure 3 these velocities are mainly non-significantly different from zero. The difference between these velocities and the initial geostrophic velocities referenced to the LADCP measurements is generally small, except in the shallowest areas near the African continent (stations 1-5 and 45-50). The uncertainties are quite similar to the imposed a priori variances, in agreement with other inverse model results.

The velocities at the reference level estimated using the inverse model allow us to compute the adjusted geostrophic mass transport. After the inverse modeling, the initial geostrophic mass transport imbalance estimated with LADCP significantly decreases, being lower than 0.2 Sv for the upper layers (layers 1 to 4), except for summer when it reaches 0.5 Sv, and smaller than 0.1 Sv for the remaining layers as indicated in Table 3. In the following section, the seasonal circulation will be described to estimate the shape of the seasonal cycle for the upper and intermediate layers. The seasonal cycle of the NADW layer will not be studied as the transport is not significantly different from zero.

5. Seasonal circulation in 2015.

5.1. The Canary Current and the flow along the African slope.

Figure 4 illustrates the accumulated mass transport of the thermohaline layers and highlights the behavior of the surface layer. Surface transport is also computed for each season from the AVISO MLSA data product. In total, the AVISO estimated transport presents a high correlation with the thermocline transport estimations, ranging from 0.7 to 0.8 for the surface layers (1 and 2), and from 0.6 to 0.8 for the NACW layers (3 and 4) (Table 4). These correlations are higher during the summer cruise and lower during the fall cruise. Though the sign criteria in Figure 4 is negative/positive standing for in/out of the box, hereafter geographical signs will be used (positive being northward/eastward and negative southward/westward). Table 5 shows the transport of each water mass. To allow a better understanding of the flow along the African Slope,

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the flow across the northern and southern transects was split. On the northern transect by the island of Lanzarote, and on the southern transect at each side of station 40 for the upper flow (SW and NACW) and each side of station 42 for the intermediate flow (IW) attending to the pattern observed on the accumulated transport shown in Figures 4 and 5. The area between Lanzarote and the African shelf is the Lanzarote Passage (Figure 1).

Figure 4 a and Table 5 show that in winter, all the southward flow at the northern transect (-3.4±0.3 Sv) flows east of Lanzarote, through the LP. Likewise, at the southwestern transect, east of station 40, a mass transport of -2.9±0.4 Sv flows near the African coast (Table 5). Therefore, in winter the CC flows southward across the easternmost stations with a total mass transport of nearly 3 Sv at both the northern and southern transects (Figure 4b). This flow is stronger at the NACW level than at the SW level (Table 5). An inflow of +0.9±0.6 Sv enters the region through the western section and leaves through the southern transect.

In spring, the southward flow concentrates west of Lanzarote (Figure 4 c) and hence, the flow along the African Slope is very weak (Table 5). Throughout this season, the CC not only shifts westward but also weakens, carrying -2.1±0.7 Sv at the northern transect and -1.2±0.6 Sv at the southern transect (Table 5). The accumulated mass transport and the AVISO geostrophic velocity fields suggest that the CC enters the basin between Lanzarote and station 15 (15.5ºW) and leaves in the vicinity of station 35 (16.9ºW) (Figure 4c and d).

The summer circulation presents a more complex situation where the southward flow of the CC at the northern transect is observed on both sides of Lanzarote (Figure 4e). A steady southward flow of about 1 Sv flows parallel to the African coast crossing the LP and the eastern side of the southern section (Table 5). Figure 4e shows that the second half of the CC flows between Lanzarote and station 15 (stations 11 to15), carrying -2.0±0.6 Sv consistently with Figure 4f. Interestingly, this flow exits the study area through the western transect (Table 5 and Figure 4f). This western diversion of the flow seems to be driven by the large mesoscale features observed in the AVISO field of Figure 4f.

An abrupt change in the circulation is observed in fall when a northward flow appears along the African slope at both the LP (+1.1±0.3 Sv) and the southern transect (+1.7±0.4 Sv) (Figure 4g and Table 5). In addition, a relatively strong flow with a magnitude similar to the CC (+3.2±0.4

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Sv) enters the box on the western side (Figure 4g and Table 5). A big portion of it (2.0±0.6 Sv) diverts northward and leaves the box between stations 18 and 20 (Figure 4g and h) while the remaining part of the flow (-1.4±0.6 Sv) leaves the area through the southern transect around station 32 (Figure 4g and h). In both the AVISO field (Figure 4h) and the accumulated transports (Figure 4g), we can observe the recirculation of the CC taking place as the flow leaves the study area between stations 31-36 and comes back as a northward circulation east of station 36.

In fall, the AVISO field (Figure 4h) shows that the CC flows west of our westernmost stations, and partially enters the area through the western transect. Hence the transport measured on the western transect might underestimate the CC fall transport that has been reported to be as large as 5.8±0.2 Sv (Pérez-Hernández et al., 2013).

5.2. The CiPU and the intermediate circulation.

The intermediate northward flow along the African slope of the CiPU can be observed in three seasons (Table 5). During winter, the CiPU carries a weak transport of +0.3±0.2 Sv across the southern section, and of +0.5±0.1 Sv over the westernmost side of the LP (Table 5 and Figure 5). In summer, while the CiPU at the southern transect is similar to the one in winter (+0.6±0.3 Sv), it decreases considerably at the LP (Table 5). In fall, the CiPU reaches its maximum strength carrying northward around 1 Sv in both sections (Table 5 and Figure 5).

During winter, and also in fall a cyclonic structure appears at the northwestern corner of our study area (Figure 5a, b, and e). During fall, a large structure develops along the southern transect with an identical behavior of the upper layers (Figure 4g and 5a). This structure is the intermediate flow that recirculates together with the upper layers of the CC. In the remaining seasons, a large mesoscale activity exists.

6. The seasonal amplitudes of 2015.

In Figure 6 the seasonal cycle of the CC is defined as in Section 5. For the sake of comparing with previous studies, and due to the presence of large mesoscale features on the southern transect - as a result of the mesoscale structures generated by the islands in the path of the Canary Current and Trade Winds (Borges and Hernández-Guerra, 2004 ; Hernández-Guerra et al., 1993) - here we have only considered the transport across the northern section in all seasons

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for the CC and the LP. An exception was made for the fall CC since its transport is defined by the flow across the western transect.

The CC has a seasonal cycle that shows a maximum net southward transport in fall and minimum transport in summer (Figure 6a). During fall, both the thermocline and the intermediate layers contribute to the maximum southward transport. However, the transport in fall at the thermocline layer is slightly weaker than the winter estimation. The seasonal amplitude of the CC net transport is 4.5±1.2 Sv, (Table 6). In addition, the intermediate layers have a larger seasonal amplitude (3.3±0.9 Sv) than at thermocline waters (1.4±0.7Sv) (Figure 6a).

Figure 6b reveals that the seasonal amplitude in the LP is dominated by the thermocline layers (see also Table 6 where the). The maximum southward transport at the LP is achieved in winter when the CC flows through it (Figures 4a and b). In contrast, the maximum northward transport occurs in fall, when the CC recirculates northward and the CiPU develops (sections 5.1 and 5.2). The net LP seasonal amplitude of 5.3±0.6 Sv is slightly larger than the seasonal amplitude of the CC.

The seasonal amplitudes for the eastern boundary are next compared with results from previous cruises. Machín et al. (2006) showed a seasonal amplitude for the CC and LP (extending only to the thermocline layers) of 2.2 ±1.0 Sv and 2.8±0.1 Sv, respectively. Their amplitude estimations are slightly weaker than our thermocline seasonal amplitude estimations (Table 6). Likewise, the comparison with Vélez-Belchí et al. (2017) shows that their estimations of the total seasonal amplitudes (thermocline plus intermediate layers) of the CC (4.1±0.5 Sv) match our results (4.5±1.2 Sv). In contrast, the amplitude at the LP is weaker (3.7±0.4 Sv) than the 5.3±0.6 Sv estimated in this study. A later study by Casanova-Masjoan et al. (2020) obtains a seasonal amplitude at the LP of 4.2±0.4 Sv to 7.6±0.6 Sv for the periods 2016-2017 and 2017-2018 respectively, agreeing with the 5.3±0.6 Sv described here. In both Vélez-Belchí et al. (2017) and Casanova-Masjoan et al. (2020), the seasonal cycle is estimated as the transport difference between a fall and a spring cruise when all the cruises have been carried out in the Canary Islands, which

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might not be capturing the full seasonal cycle. For example, in Figure 6 the maximum southward transport at the LP is in winter rather than in spring.

7. A composite seasonal cycle for the CC and LP from previous observations

In Figure 7, all the available observations in the area have been used to compute a composite seasonal cycle for the CC and LP (dots and blue line), and to compare them with the estimations from this work. The longitudinal shifts and thermocline transports of the CC shown in Figure 4 agrees with Machín et al. (2006) where the CC is reported to shift westward from spring to fall. However, in our estimations the winter CC is further east and stronger than in Machín et al. (2006) probably because their cruise took place in early-January while ours took place in late winter (mid-February).

In spring and summer, the CC flows through the archipelago mainly along the thermocline layers (Figures 5c and d, 6c and 7a). In spring, the CC matches the estimations of 2.8±0.8 Sv from Machín et al. (2006) and 2.4 ±1.1 Sv from Vélez-Belchí et al. (2017). A stronger flow (3.4 to 3.1 Sv) is reported by Casanova-Masjoan et al. (2020), although their lack of uncertainty prevents the comparison. Fewer cruises have taken place in summer, nevertheless our average southward CC of 2.0±0.6 Sv (Figure 7a) agrees with the 2.9±0.8 Sv reported in Machín et al. (2006) and with the -2.1±0.9 Sv (for 1992) and -2.3±1.1 Sv (for 2011) shown in Hernández-Guerra et al. (2014).

Most of the available measurements of the CC have taken place in the fall. Our estimations in this season are 5.6±0.6 Sv southward (3.2±0.4 Sv at surface layers and 2.4±0.5 Sv at intermediate layers, Figures 5e, 6c, and 7a). This transport agrees with the 6.5±0.4 Sv described in Vélez-Belchí et al. (2017) for fall 2013. The thermocline transport of the CC in fall is stronger than the 1.5±0.7 Sv estimated in Hernández-Guerra et al., (2017), but lies within the 1.9 to 3.4. Sv range given in Casanova-Masjoan et al. (2020). In our results, the AVISO fields (Figure 4h) suggest that the CC is flowing west of our section and only partially entering our study area. This agrees with previous studies in which stronger transports are reported west of La Palma island (west of 19ºW) as in Hernández-Guerra et al. (2005) (4.7±0.8 Sv) or Pérez-Hernández et al. (2013) (6.2±0.6 Sv).

At the LP, the flow is mainly southward throughout the year except in the fall, when a northward recirculation of the CC through the LP is observed in Figures 4g, h, 5a, e. and 7. This northward recirculation of the CC at surface and intermediate levels is also accompanied by the

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onset of the CiPU (Figure 5e). The fall southward to northward reversal transport of the upper layers along the African slope has been widely reported in previous studies with a mass transport ranging between 1 and 3 Sv (Casanova-Masjoan et al., 2020; Fraile-Nuez et al., 2010; Hernández-Guerra et al., 2002, 2017; Knoll et al., 2002; Machín et al., 2006; Pérez-Hernández et al., 2015).

Hence, we can confirm that the seasonal cycle of 2015 matches the historical composite seasonal cycle, and we can add these new estimations to the composite historical seasonal cycle.

8. The seasonal cycle of the eastern boundary and its contribution to the AMOC.

Figure 8 presents the seasonal cycle of the AMOC and each of its components from the RAPID-MOCHA array from 2004 to 2018. The seasonal cycle of the AMOC has an amplitude of about 5.0 Sv (Figure 8a). The Upper-Mid Ocean (T_{UMO}) presents the largest contribution to this seasonal cycle, presenting an amplitude of 5.5 Sv (Figure 8d). The AMOC presents three peaks: a minimum in March and two maximums in July-August and November (Figure 8a). Several studies have described the seasonality of the Ekman transport and the Gulf Stream and have attributed their variability to be the main driver of the July AMOC peak (Atkinson et al., 2010; Meinen et al., 2010). The main contributor to the T_{UMO} seasonal pattern is a positive dynamic height that develops in the eastern basin from August to December (Chidichimo et al., 2010; Kanzow et al., 2010; Pérez-Hernández et al., 2015). Further exploring the structure of the UMO using the dynamic height at the east and west end of the RAPID-MOCHA array, Pérez-Hernández et al. (2015) showed that the UMO seasonal variability is driven by the variability on the eastern boundary.

From the previous section, it is noticeable that the seasonal amplitude of the AMOC (5.0 Sv) and T_{UMO} (5.5 Sv) matches the total seasonal amplitude at the LP (5.3±0.6 Sv). In Pérez-Hernández et al. (2015), a significantly high seasonal correlation (higher than 0.7) was estimated for the T_{UMO} and the transport at the LP at the surface and intermediate layers. This was done using data from a mooring located at the LP (the EBC4) and the RAPID-MOCHA array. Here, a similar comparison is done in Figure 9 where the normalized transport (removing the mean from the values and diving them by the standard deviation) of the T_{UMO} is shown together with a composite historical seasonal cycle estimated for the LP (this cycle is estimated as in Figure 8, including also

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the 2015 results). The T_{UMO} and the composite LP transport present the same variability and a correlation of 0.8 (p-value 0.003).

9. Conclusions.

In 2015, the RaProCan project joined efforts with the SeVaCan project to estimate the seasonal cycle in the eastern boundary of the North Atlantic Subtropical Gyre with cruises carried out during the same year (February, April, July, and November) for the first time. Previous studies assessed the seasonal cycle of the CC by combining cruises from different years (Vélez-Belchí et al., 2017; Casanova-Masjoan et al. 2020). This study can be considered as a southward extension of the study carried out by Machín et al. (2006) between Madeira and the Canary Islands over a time spanning 18 months. Likewise, Fraile-Nuez et al. (2010) gave a description of the seasonal cycle of the flow existing just in the LP using a single mooring deployed over 9 years. Here we have combined the results from the 2015 cruises with all the available observations in the area to achieve a better understanding of the seasonal cycle of both the CC and the LP.

Although both the CC and the LP present similar seasonal amplitudes (4.5±1.2 Sv and 5.3±0.6 Sv, respectively), only the shape of the seasonal cycle at the LP resembles the seasonal cycle of the T_{UMO} (Figure 9). The seasonal cycle at the LP can be summarized as a southward flow that is maximum in winter and that reverses during fall due to the recirculation of the CC and the presence of the northward CiPU. A large agreement has been corroborated between the seasonal cycle amplitude of the LP transport and of the T_{UMO} as previously reported with moorings data in Pérez-Hernández et al (2015), and with hydrographic cruises in Vélez-Belchí et al. (2017), and Casanova-Masjoan et al. (2020). This confirmation has been done with the data from the four hydrographic cruises used in this study, and also with the total historical hydrographic data compiled in the area. Hence this study concludes that the LP and T_{UMO} have similar seasonal transport amplitudes (ca. 5 Sv) and that their seasonal cycle has a 0.8 correlation coefficient. In early studies, part of this seasonality was explained with a Rossby wave model (Kanzow et al., 2010; Pérez-Hernández et al., 2015), although Vélez-Belchí et al. (2017) showed that this model was very dependent on the longitudinal extent of the wind-stress chosen.

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Open Research

Hydrographic data from the *RaProCan* project (https://www.oceanografia.es/raprocan/), which is the Canary Islands component of the Spanish Institute of Oceanography (IEO) ocean observing system (Tel et al., 2016) can be found at https://ocean.ices.dk/core/iroc and https://cdi.seadatanet.org/search [Dataset] . The RAPID-AMOC time series was downloaded from the project website (https://rapid.ac.uk) [Dataset] .

Maps of Absolute dynamic topography (MADT) were obtained from the Space Oceanographic Division of the Collective Localization Satellites (CLS) through the Archiving, Validation, and Interpretation of Satellite Oceanographic Data project (AVISO; https://www.aviso.altimetry.fr/en/home.html) [Dataset] .

Wind data are estimated from the National Center for Environmental Prediction (NCEP Reanalysis Derived data; https://psl.noaa.gov/data/gridded/data.ncep.reanalysis.html) using the Weather Research and Forecasting (WRF) model (version 3.9.1), developed at the National Center for Atmospheric Research (Skamarock et al., 2008) [Dataset] .

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Tables

Table 1. Hydrographic cruises

Cruise	Dates	Number of stations
SeVaCan1502	26 February / 6 March 2015	51
RaProCan1504	07-15 April 2015	51
SeVaCan1507	17-25 July 2015	51
RaProCan1510	01-09 October 2015	51

Table 2. Deepest limit (dbar) of the γ^n isoneutral layers used in the mass transport analyses

Layer	**γ^n (kg m^{-3})**	**Deepest limit (dbar)**	**Water masses**
1	26.440	20	SW
2	26.850	298	SW
3	27.162	540	NACW
4	27.380	740	NACW
5	27.620	967	AAIW
6	27.820	1314	AAIW MW
7	27.922	1656	NADW
8	27.975	2015	NADW
9	28.008	2241	NADW
10	28.044	2592	NADW
11	28.072	2965	NADW
12	28.0986	3519	NADW
13	28.1295 (Bottom)	3923	NADW

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Table 3. Accumulated imbalance (in Sv) for the net transport of each water mass after applying the LADCP correction to each cruise and after the inverse box model.

Cruise	Imbalance after LADCP			Imbalance after inverse box modeling		
	SW/NACW	AAIW/MW	NADW	SW/NACW	AAIW/MW	NADW
SeVaCan1502	0.78	0.12	1.41	0.2±0.1	-0.1±0.1	-0.1±0.2
RaProCan1504	0.20	-0.13	0.78	0.1±0.1	0.0±0.1	0.0±0.2
SeVaCan1507	2.08	0.00	-1.33	0.5±0.2	-0.1±0.1	-0.0±0.2
RaProCan1510	-4.49	-3.25	-8.51	0.2±0.1	0.0±0.1	0.1±0.2

Table 4. Correlations between the accumulated mass transport at the surface (layers 1 to 2 from Table 2) and NACW (layers 3 to 4 from Table 2) layers and the estimated accumulated mass transport from AVISO per season. The p-value is shown in brackets, and non-significant correlations have been marked with an *.

	Winter		**Spring**		**Summer**		**Fall**	
	SW	**NACW**	**SW**	**NACW**	**SW**	**NACW**	**SW**	**NACW**
Lanzarote Passage	0.91 (0.00)	0.95 (0.00)	-0.32 (0.37)*	0.59 (0.00)	0.82 (0.00)	0.74 (0.00)	0.80 (0.01)	0.44 (0.00)
Northern transect	0.00 (0.98)*	-0.40 (0.15)*	0.63 (0.01)	0.38 (0.29)*	0.80 (0.00)	-0.66 (0.04)	0.41 (0.13)*	0.45 (0.19)*
West transect	0.94 (0.00)	0.96 (0.00)	0.65 (0.11)*	0.52 (0.04)	0.55 (0.20)*	0.58 (0.02)	0.57 (0.20)*	0.19 (0.49)*
Southern transect	0.59 (0.00)	0.61 (0.00)	0.91 (0.00)	0.65 (0.11)	0.93 (0.00)	0.48 (0.28)	0.84 (0.00)	0.71 (0.08)*
total	0.76 (0.00)	0.69 (0.00)	0.85 (0.00)	0.61 (0.00)	0.84 (0.00)	0.83 (0.00)	0.75 (0.00)	0.65 (0.00)

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Table 5. Estimated seasonal transports (Sv) for the different transects. For the northern transect, the flow is split between East (African Slope - Afr. Slp.) and West of Lanzarote (W. of Lz.). For the southern a transect, the flow is split into African Slope and West of Lanzarote by station 40 for SW and NACW and by station 42 for the intermediate waters (IW). SW stands for the surface waters (layers 1 and 2 in Table 2), NACW for the North Atlantic Central Water layer (layers 3 and 4 in Table 2), and IW is for the intermediate water layer (layers 5 and 6 in Table 2). The sign convection of this table is geographical (positive to the north/east and negative to the south/west).

		Winter		**Spring**		**Summer**		**Fall**	
		North	**South**	**North**	**South**	**North**	**South**	**North**	**South**
Afr. Slp.	**SW**	-1.3±0.2	-1.3±0.3	-0.3±0.1	-1.1±0.2	-1.0±0.1	-1.2±0.1	-0.2±0.1	-0.3±0.2
	NACW	-2.1±0.3	-1.6±0.3	0.3±0.2	0.3±0.3	-0.1±0.3	0.2±0.3	1.3±0.3	1.9±0.3
	Total	-3.4±0.3	-2.9±0.4	0.0±0.3	-0.7±0.3	-1.1±0.3	-1.0±0.4	1.1±0.3	1.7±0.4
	IW	0.5±0.1	0.3±0.2	0.0±0.2	-0.1±0.3	0.2±0.2	0.6±0.3	1.3±0.2	1.1±0.3
W. of Lz.	**SW**	-0.4±0.6	-0.6±0.6	-1.1±0.5	-1.0±0.4	-1.2±0.2	-0.2±0.3	0.4±0.3	-0.3±0.2
	NACW	0.6±0.5	-0.7±0.5	-1.0±0.7	-0.1±0.4	-0.8±0.5	-0.2±0.4	1.6±0.5	-1.1±0.5
	Total	0.2±0.8	-1.3±0.8	-2.1±0.7	-1.2±0.6	-2.0±0.6	-0.4±0.5	2.0±0.6	-1.4±0.6
	IW	2.1±0.8	0.9±0.8	0.1±0.8	3.3±0.7	0.9±0.6	-0.2±0.6	2.5±0.7	0.2±0.7
Western	**SW**	-0.1±0.4		0.5±0.3		-1.2±0.2		1.1±0.2	
	NACW	1.0±0.4		-0.8±0.4		-0.9±0.4		2.1±0.3	
	Total	0.9±0.6		-0.3±0.5		-2.1±0.4		3.2±0.4	
	IW	1.5±0.6		-1.6±0.5		0.8±0.5		2.4±0.5	

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Table 6. Transport seasonal amplitudes (Sv) for the LP, for the CC, and the eastern boundary, at thermocline and intermediate layers and the net.

	LP	**CC**	**East. boundary**
Thermocline	4.5±0.4	1.4±0.7	6.5±0.2
Intermediate	1.3±0.4	3.3±0.9	3.8±0.4
Net	5.3±0.6	4.5±1.2	9.1±0.4

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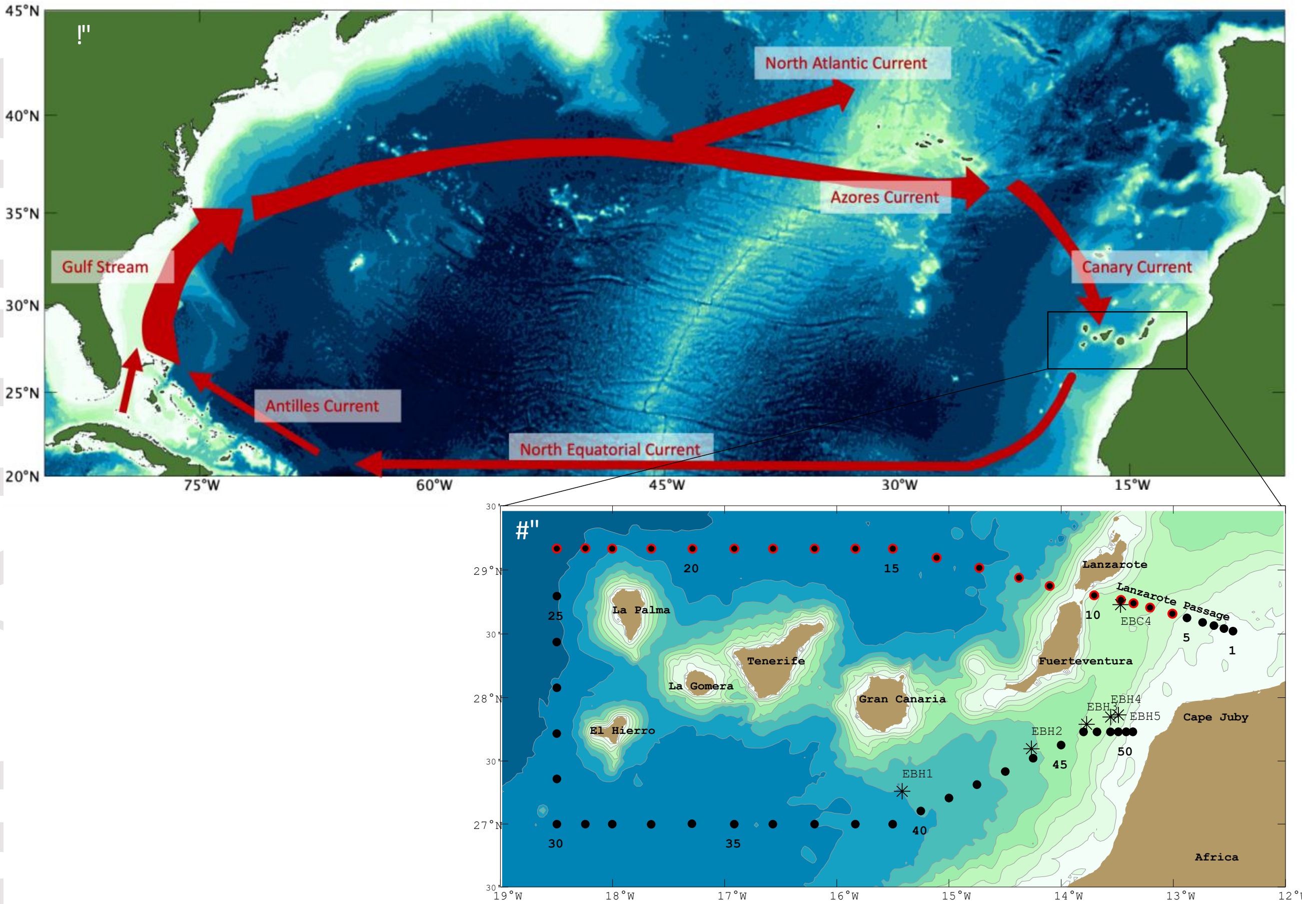

Figure 1. a) Schematic circulation of the North Atlantic Subtropical Gyre showing the main currents in the area (red) and the study area (black box. b) Zoom on the study area showing the main topographic and geographical features referred to in the text. The black circles are the stations sampled in each 2015 cruise, while the standard *RaProCan* stations are highlighted in red. One in every five stations for each cruise has been labeled. The asterisks indicate the position of the main moorings of the RAPID-MOCHA array at the eastern boundary and the mooring EBC4. Bathymetric lines range from 0 to -9000m every 500m except for the 100m isobath (grey color).

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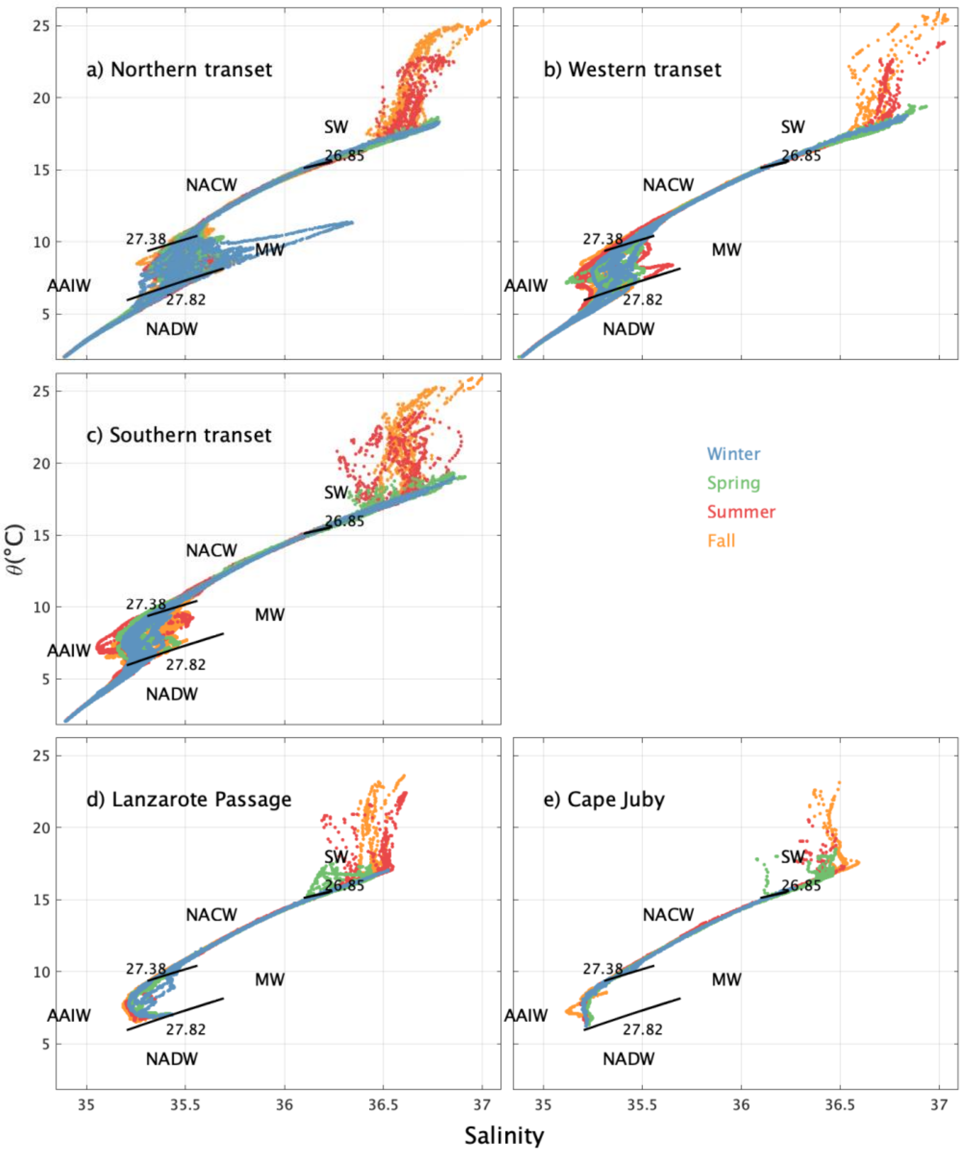

Figure 2. Seasonal θ/S diagram for each relevant geographical location: (a) Northern transect (stations 11 to 24), (b) Western transect (stations 25 to 30), (c) Southern transect (stations 31 to 46), (d) Lanzarote Passage (stations 1 to 10), and (e) Cape Juby (stations 47 to 51). Each hydrographic cruise is represented in a different color, blue for winter, green for spring, red for summer, and orange for fall. The grey thick lines correspond to the isoneutrals used in the inverse model to divide the water column into surface, central, intermediate, and deep (Table 2). NACW stands for North Atlantic Central Water, MW for Mediterranean Waters, AAIW for Antarctic Intermediate Waters, and NADW for North Atlantic Deep Waters.

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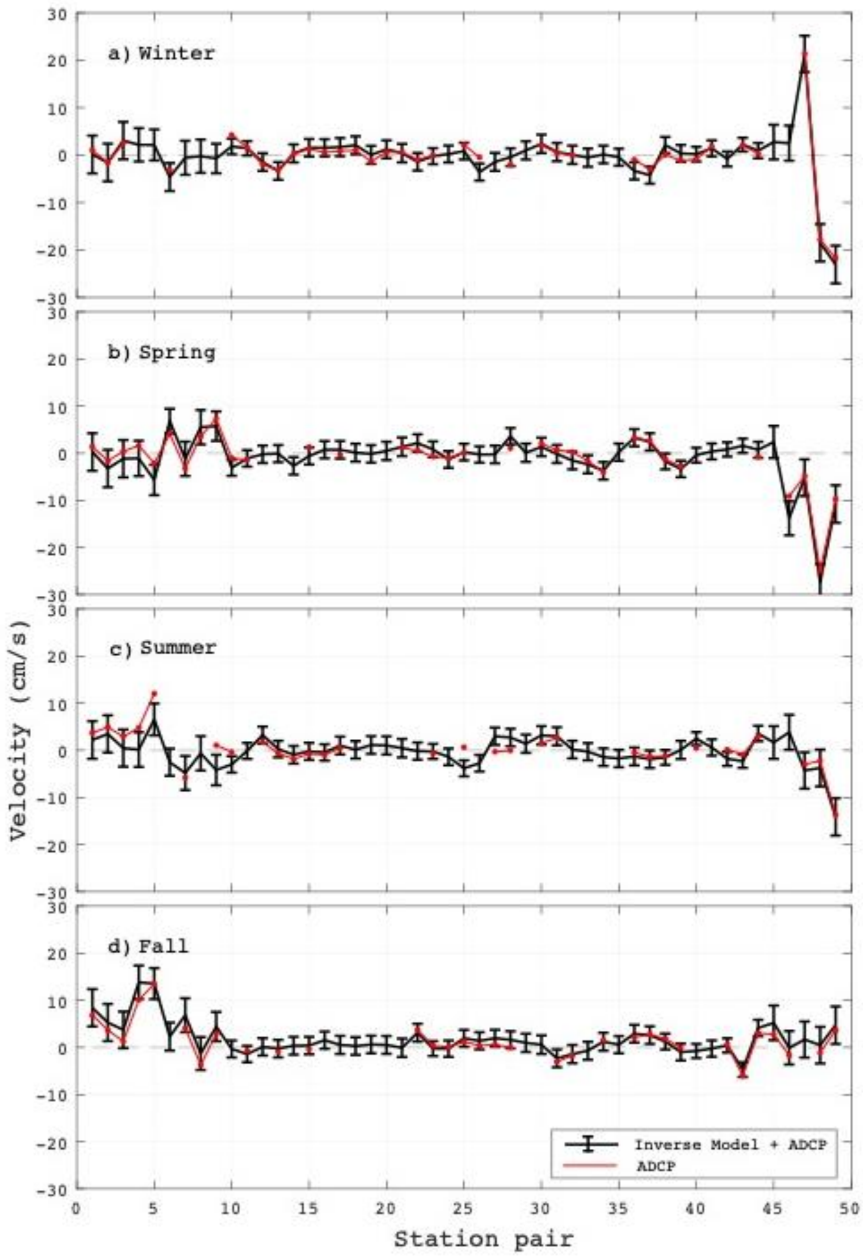

Figure 3. Velocities at the reference level – please recall that the reference level is located in the density level γ^n=27.975 kg m^{-3} for the oceanic waters, while the density level of γ^n=27.380 kg m^{-3} for the shallower stations of the LP- for each station pair as determined by the ADCP (red) and by the ADCP plus the inverse calculations with their error bars (black) for each season/cruise. (a) Winter - SeVaCan1502, (b) Spring - RaProCan1504, (a) Summer - SeVaCan1507 and (d) Fall - RaProCan1510.

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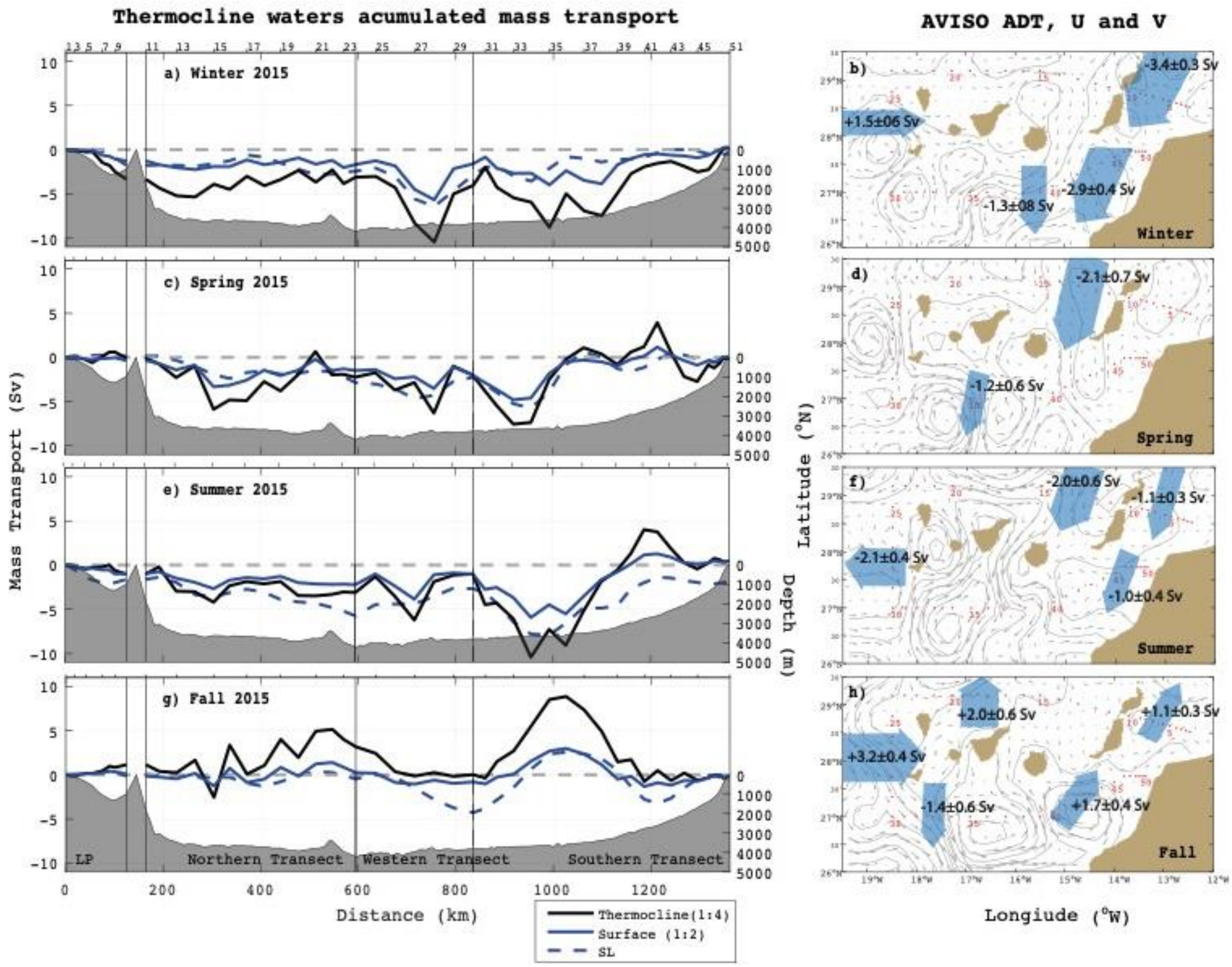

Figure 4. On the left panel: Accumulated mass transport in the thermocline layers during the (a) winter, (c) spring, (e) summer, and (g) fall cruises for the surface (layers 1:2 in blue), and the thermocline layers (layers 1:4 in black). The accumulated mass transport obtained using the surface geostrophic velocities from the AVISO MSLA product integrated to the depth corresponding to the lower limit of the seasonal thermocline waters (27.38 kg m^{-3}) are also shown (dashed blue). For reference, the bathymetry has been superimposed, a vertical black line indicates the corner of each track labeled in the lower panel. The sign convection of this figure is negative/positive standing for in/out of the box. On the right panel: AVISO Absolute Dynamic Topography (ADT) and geostrophic velocities for the (b) winter, (d) spring, (f) summer, and (h) fall cruises together with a schematic representation of the main transports.

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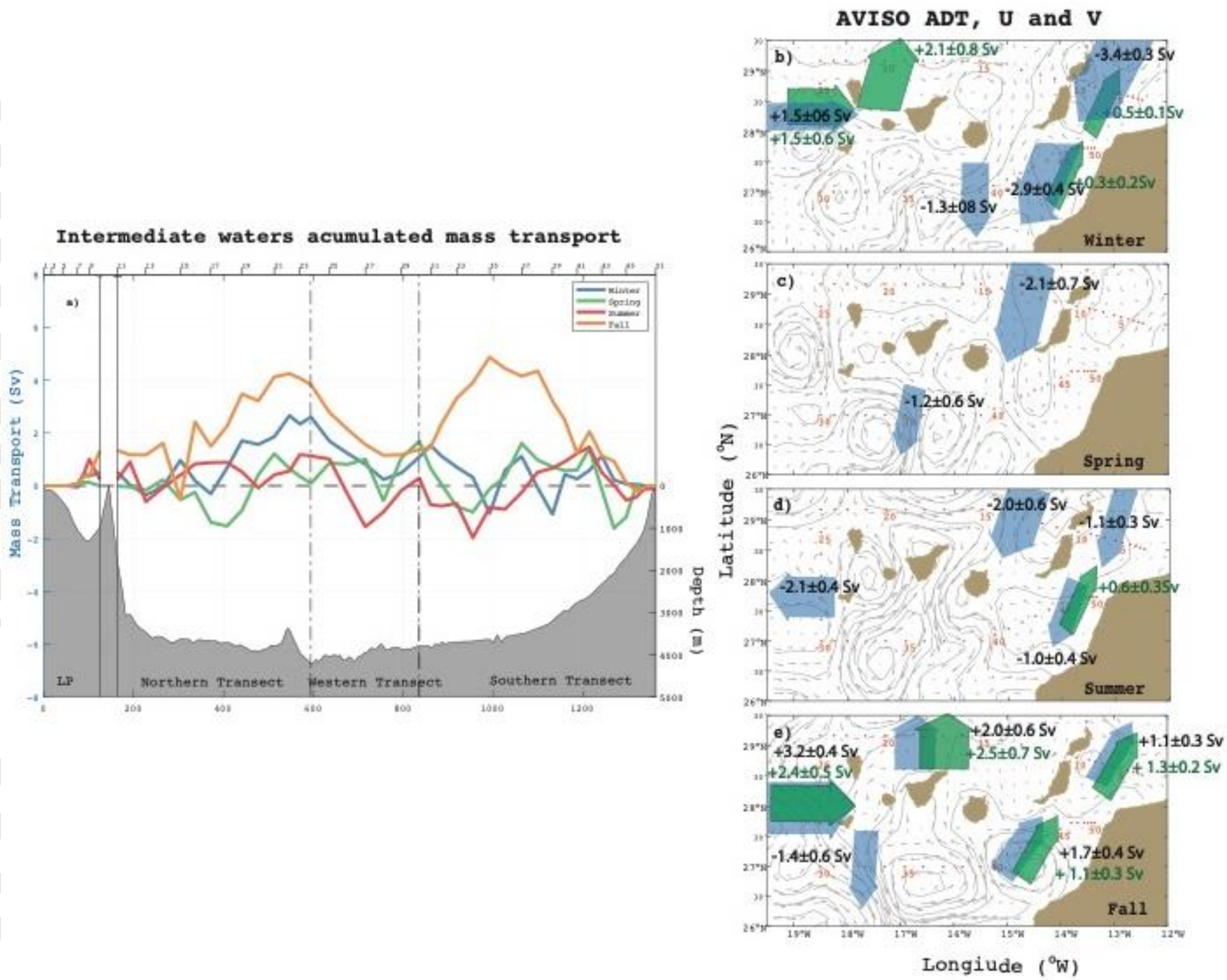

Figure 5. (a) Accumulated mass transport for the intermediate layers during (blue) winter, (green) spring, (red) summer, and (orange) fall cruises. For reference, the bathymetry has been superimposed, a vertical black line indicates the corner of each track labeled in the lower panel. The sign convection of this figure is negative/positive standing for in/out of the box. The right panels are identical to the one in Figure 4 but the schematic representation of the main intermediate circulation features overlayed with green arrows and labels for (b) winter, (c) spring, (d) summer and (e) fall.

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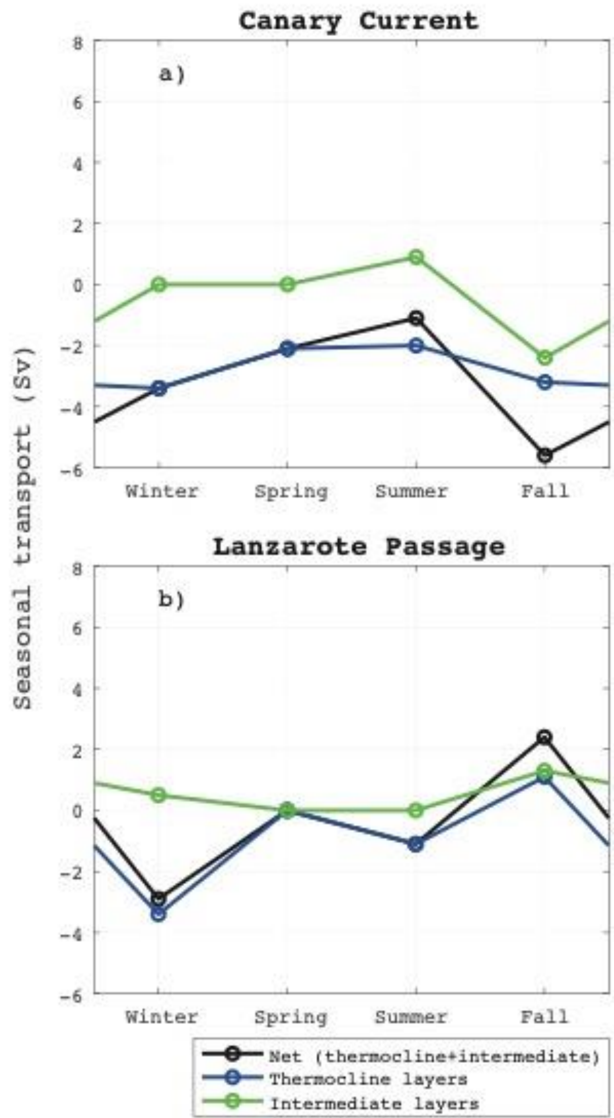

Figure 6. The seasonal cycle of the estimated transports on the thermocline layer (blue line), intermediate layer (green line), and the net (black line) for (a) CC and (b) the Lanzarote Passage. Note that the transports plotted here are those across the northern section in all seasons for the CC and the LP. An exception was made for the fall CC since its transport is defined by the flow across the western transect. In the case of winter, when the thermocline transport of the CC passes through the LP, the transport is the same in both a and b.

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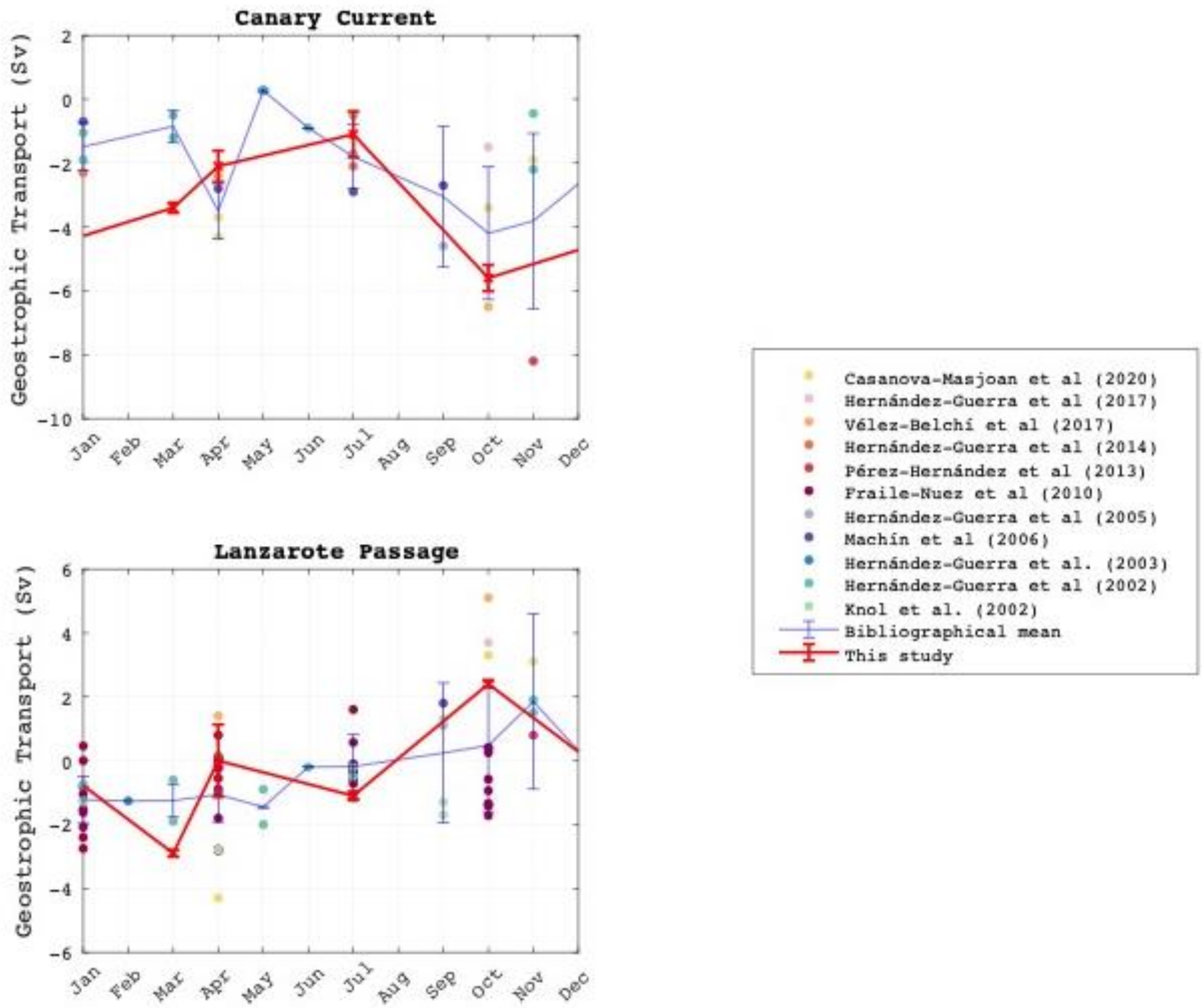

Figure 7. Total seasonal cycle (red line) for 2015 together with all the historical measurements (color dots) for the Canary Current (a) and Lanzarote Passage (b). An average composite seasonal cycle done with all the historical measurements is shown with its standard deviation (blue line and error bars).

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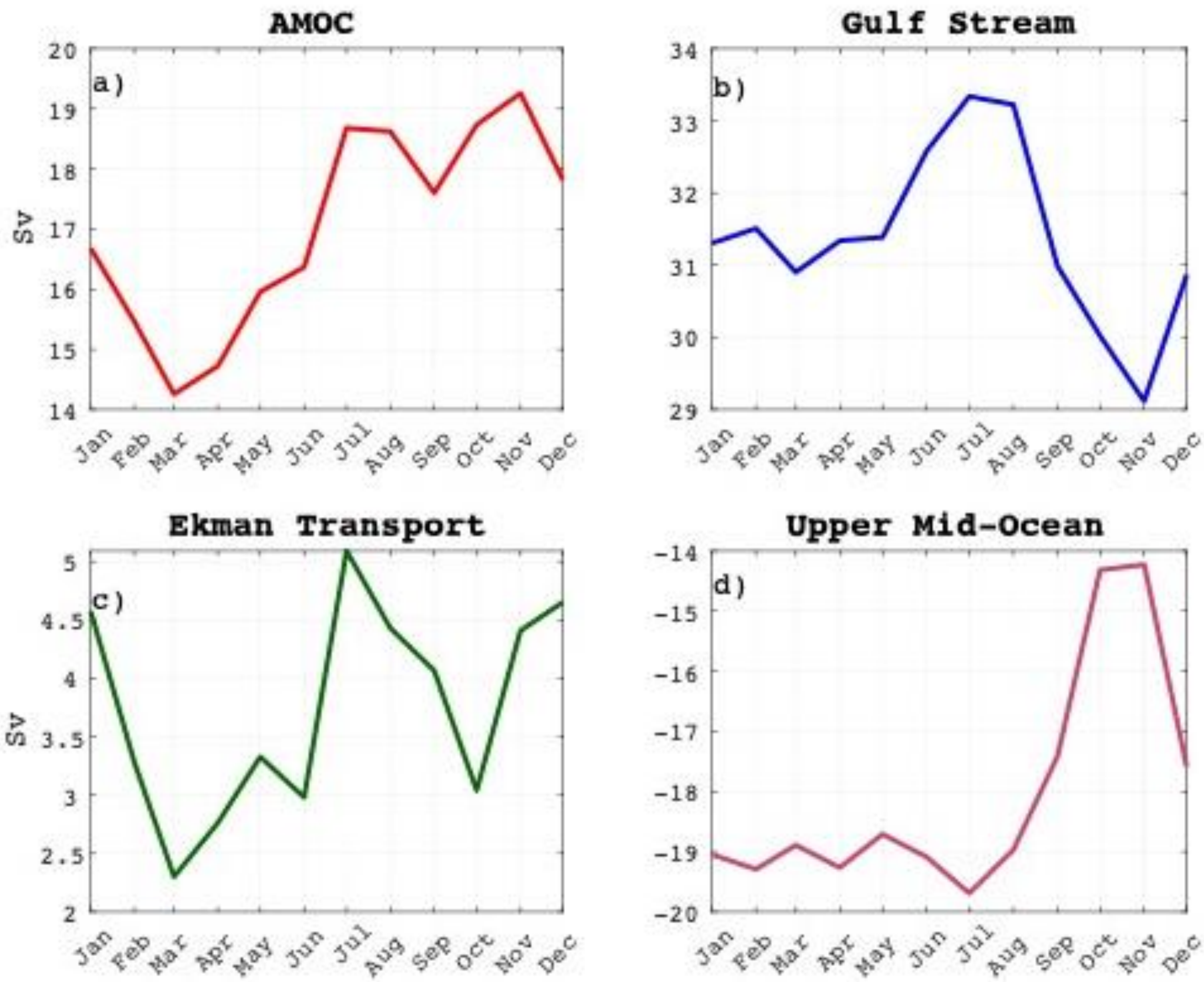

Figure 8. Seasonal cycles of the transport (Sv) of the AMOC (a) and its main components: the Gulf Stream (b), Ekman Transport (c) and Upper-Mid Ocean (d) computed using the RAPID-MOCHA time series. Note different y-axis ranges in each plot.

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Figure 9. Seasonal cycles of standardized transports (Sv) of the Upper-Mid Ocean (red) obtained from the RAPID-MOCHA array and of the historical composite seasonal cycle at the Lanzarote Passage (blue).

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