

1

1

2

3

Seasonal cycle in surface seawater isotopic composition

4

5

6

7

Gilles Reverdin¹, Claire Waelbroeck¹, Lola Goeme¹, Antje H. L. Voelker^{2,3}, Hans-Christian Steen-Larsen⁴

9

10

11

12

¹ Laboratoire LOCEAN/IPSL, Sorbonne Université-CNRS-IRD-MNHN, Paris, 75005, France

14

² Instituto Português do Mar e da Atmosfera (IPMA), Algés, Portugal

15

³ Center of Marine Sciences (CCMAR), Universidade do Algarve, Faro, Portugal

16

⁴ Geophys. Inst., Univ. Bergen, Bergen, Norway

17

18

Corresponding author: Gilles Reverdin gilles.reverdin@locean.ipsl.fr

19

Code de champ modifié

20 Abstract

21 Data of stable isotopologs of water in the quality-controlled ~~CISE-LOCEAN~~ database from
 22 the isotopic platform facility at LOCEAN (hereafter “CISE-LOCEAN”) provide an
 23 unprecedented sampling of the surface Atlantic Ocean and the southwest Indian Ocean. This
 24 investigation of $\delta^{18}\text{O}$, $\delta^2\text{H}$ and deuterium excess (d-excess) in surface sea water for 20
 25 subregions indicates a large variability of the regional regression lines with salinity as a
 26 function of latitude for $\delta^{18}\text{O}$ and $\delta^2\text{H}$, with weakest slopes near the equator, but also to a
 27 lesser extent with weaker slopes in the western Atlantic compared with the eastern
 28 Atlantic at low and mid-latitudes. These regressions often explain a large share of the
 29 variability in the data (40 to 70%), except for d-excess at the higher latitudes and $\delta^{18}\text{O}$,
 30 $\delta^2\text{H}$ near the equator. The data are sufficient to extract a seasonal variability in these
 31 relationships and an average seasonal cycle in some regions of the North Atlantic
 32 subpolar gyre with signals probably related to changes in the input of heavier sea ice
 33 melt and depleted water of Arctic origin. A seasonal cycle in $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess
 34 correlated to the one in salinity is also found in the central equatorial Atlantic with the
 35 strong contrast between the vicinity of the equator and the region to its north,
 36 associated with the seasonal shift of the ~~precipitation belt~~ Inter-Tropical Convergence
 37 Zone and with the seasonality of the equatorial upwelling. ~~In those regions there is also~~
 38 ~~some evidence in seasonal variability of the d-excess that is not currently explained.~~
 39 Elsewhere, the seasonal signal usually does not ~~come out of~~ emerge from the noise. An
 40 independent 7-8 over 7-years long long-time series of nearly monthly surface sampling
 41 at the BATS site near Bermuda (31°50'N, 64°10'W) also indicates a rather well-defined
 42 seasonal cycle of ~~SSS~~ surface salinity, $\delta^{18}\text{O}$, $\delta^2\text{H}$ and ~~deuterium d-~~ excess. ~~There is also~~
 43 ~~some~~ Its indication of interannual variability in $\delta^{18}\text{O}$, which might not be does not appear
 44 closely related to salinity.

Mis en forme : Police :Non Gras

1. Introduction

Seawater isotopic composition ($^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$ ratios expressed as $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in ‰ (per mille) in the VSMOW/SLAP scale) is classified as an Essential Ocean/Climate Variable (EOV/ECV) in international programs such as GEOTRACES and GO-SHIP, respectively aiming to improve the understanding of biogeochemical cycles, and to develop a globally coordinated network of sustained hydrographic sections as part of the global ocean/climate observing system. Except for fractionation during phase changes, thus mostly at the sea surface, and meteoric inputs from the atmosphere and the land masses, the water isotopic composition is nearly conservative in the ocean, with chemical and biological processes having minimal effect. These

Thus, stable seawater isotopes are used, in addition to salinity, to trace sources of freshwater (precipitation, evaporation, runoff, melting glaciers, sea ice formation and melting), as well as to characterize water masses, both at the ocean surface and in the ocean interior (Schmidt et al., 2007; Hilaire-Marcel et al., 2021). More regionally, in addition to salinity, they are used to characterize water masses and their mixing pathway into the ocean (Frew et al., 2000; Voelker et al., 2015). At the sea surface, isotopic properties of the evaporation (Benetti et al., 2014) and of regional precipitation can be used to better characterize air-sea freshwater fluxes from the surface distribution of water isotopes (Benetti et al., 2017a; Benway and Mix, 2004; Conroy et al., 2014).

~~A major emphasis in the data collection has been in~~ In on the high latitude oceans. ~~There,~~ continental (or icebergs) glacial melt, formation or melt of sea ice, and high-latitude river inputs (for the Arctic) leave imprints on the surface ocean isotopic composition, and for the Southern Ocean in the top 800 m water column close to ice shelves (Aoki et al., 2023; Hennig et al., 2024). These processes, as well as changes in ocean circulation vary on seasonal time scales, as well as on longer time scales up to interdecadal, which can influence the relationship between salinity and the water isotopes. At low latitudes, the surface water masses get their isotopic characteristics from ocean circulation, with in particular upwelling of water originating from subtropical evaporation-dominated

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

regions, as well as from local evaporation, rainfall (Benway and Mix, 2004), and from the freshwater inflow of large tropical rivers, such as the Amazon River in the Atlantic Ocean (Benetti et al., 2017a). The more investigated low latitude Pacific Ocean also presents a large variability in these processes which shape the relationship between salinity and water isotopes, in particular on seasonal and interannual (i.e. ENSO) to decadal modes of variability (e.g. Horikawa et al (2023) for the southwest equatorial Pacific, Conroy et al (2017) for the central equatorial Pacific, Dassié et al (2018) for the South Pacific Convergence Zone). Even rainfall properties might change interannually, depending on the origin of the rain water, such as illustrated in the central-western Pacific in Conroy et al (2014) or in the eastern Pacific north of the equator in Benway and Mix (2004).

Mis en forme : Anglais (États-Unis)

To some extent, these isotopic signatures in ocean water are archived in past climate records, for example, for $\delta^{18}\text{O}$ in carbonates from warm water corals, giant clams or planktonic foraminifera (Mulitza et al., 2003; Abram et al., 2020; Dissard et al., 2021; Dong et al., 2025), or for $\delta^2\text{H}$ in lipids (Schefuss et al., 2005; Schouten et al., 2006; Leduc et al., 2013). They are acquired from the seawater isotopic properties at the time when these organisms grow. Measurement of seawater isotopes in the upper ocean at low latitudes are often vital for paleoclimatic studies, as they are needed is strongly recommended, often at a seasonal resolution, to understand these archives. Thus, dedicated efforts should ideally combine collection of living (or recent) organisms nearly simultaneous to the seawater samples, which has been done for example at low latitudes within the frame of paleoclimatic research programs (e.g., PAGES CoralHydro2k working group; Konecky et al., 2020, Atwood et al., 2025, 2026) or to calibrate planktonic foraminifera $\delta^{18}\text{O}$ signals (e.g., Wilke et al., 2009; Rebotim et al., 2019; Salgueiro et al., 2020). to calibrate proxies of past ocean variability in marine carbonate records such as corals and foraminifera (e.g., PAGES CoralHydro2k working group; Konecky et al., 2020, Atwood et al., 2025). Additionally, surface ocean seawater isotopes are also used to characterize evaporation rates and air-sea interactions (Benetti et al., 2017a).

Except for fractionation during phase changes, thus mostly at the sea surface, the water isotopic composition is nearly conservative in the ocean, with chemical and biological processes having minimal effect. Because Except where sea ice forms or melts, because of that the tendency for precipitation to be depleted in the heavier isotopes compared to

seawater and evaporation to cause an increase in the heavier isotopes, there tends to be regionally quasi-linear relationships between ocean surface salinity SSS (practical salinity with no unit) and water isotope properties ($\delta^{18}\text{O}$, $\delta^2\text{H}$, but also the derived d-excess defined as $\delta^2\text{H} - 8 \cdot \delta^{18}\text{O}$, together later referred to as Wisotopes) (Craig and Gordon, 1965)., as exemplified for $\delta^{18}\text{O}$ in Legrande et al. (2006). In their paper, regional relationships are used between $\delta^{18}\text{O}$ and S are thus used by Legrande et al. (2006) to produce an average climatology to reconstruct global distribution of water isotopes in the surface ocean, as well as through the water column. This is used to benchmark in order to provide a climatology to which isotope-enabled models (e.g., Delaygue et al., 2000; Roche and Caley, 2013). can be referred to. Regional relationships are also used to characterize water masses or different freshwater sources during their formation (Frew et al., 2000; Voelker et al., 2015).

Regional relationships between water isotope properties and salinity are further used in Reverdin et al (2025) to identify systematic biases between different data sets collected in the same years, but not necessarily in the same season. That study, as well as Glaubke et al. (2024) also show how the relationships between salinity and $\delta^{18}\text{O}$, $\delta^2\text{H}$, or d-excess vary spatially vary across the southern subtropical Indian Ocean, an indication of the possible limitation of need to refine the concept to smaller regions. Similar spatial variability was also suggested across the North Atlantic subtropics in Benetti et al (2017a).

At high latitudes, Wisotopes sampling primarily occurs during the summer, while at low latitudes, data sampling is more spread out throughout the year, but remains quite scarce. As a result, there are relatively few low latitude regions with enough data to accurately estimate seasonal variability (Atwood et al, 2025)., which reduces the capability of current datasets to benchmark models or calibrate archived climate records. Additionally, Atwood et al (2025) assume that there are Potential remaining biases in data and possible no remaining biases in the data sets and that interannual/decadal variability is minimal or can be modelled can also result in apparent variability in the relationships between Wisotopes and salinity. However, often, different relationships were found in nearby regions Thus, it can be difficult to interpret differences in the relationships between water Wisotopes and salinity that were

~~sampled in different seasons in relatively nearby regions which may not have been sampled in the same season~~, such as the western and eastern North Atlantic subtropics in Benetti et al. (2017a), ~~which raises the question as whether these different differences are artifacts caused by unresolved seasonal or interannual signals.~~

Fortunately, there have been some concerted attempts to resolve surface Wisotopes seasonal variability, away from coastal areas. This has been the case, for example, ~~since late 2010~~, in the surface North Atlantic subpolar gyre (NASPG) between Iceland and North America (Reverdin et al., 2018), ~~in particular between 2010 and 2018, extended and more recently since 2018~~ between southeastern Greenland and the North Sea. An ~~early~~ subset of these NASPG data was used in Benetti et al. (2014) to identify seasonal variability of the relationship between ~~water~~ Wisotopes and salinity. These authors found seasonal changes on the Labrador and Newfoundland shelves and close to them in the western NASPG that were attributed to the cycle of sea ice formation in winter and sea ice melt in spring, ~~often usually~~ upstream of the site of data collection. In the subtropics, a concerted effort was also undertaken at the BATS site ~~near~~ Bermuda (31°50'N, 64°1'W) to regularly sample surface-water isotopes since 2011.

In this study, we aim at investigating the seasonal cycle in surface Wisotopes, and their relation to salinity, across a large part of the Atlantic Ocean and in the south-west Indian Ocean, based on water samples mostly measured in the LOCEAN stable isotope laboratory.

2. Data and Methods

2.1 Data

We primarily use the CISE-LOCEAN database (Reverdin et al., 2022), which originates from samples exclusively analyzed in the isotopic laboratory of LOCEAN (Sorbonne Universités, Paris) since 1999 or in rarer instances by LOCEAN personnel elsewhere. Associated salinity data usually originate from calibrated CTD casts or from discrete samples measured on a salinometer.

Stable seawater isotope data have been produced in the last fifty years by a variety of methods. For example, most data compiled in the “GISS Global Seawater Oxygen-18 Database -V1.21” for stable seawater isotopes (LeGrande and Schmidt, 2006) originate

from Isotope-ratio Mass Spectrometry (IRMS). They were mostly measured by dual-inlet technology (highest precision), whereas, since 2000, the continuous-flow method (lower precision) also became widespread for seawater isotope analysis. Since 2010, laser spectroscopy, in particular cavity ring-down spectroscopy (CRDS), turned into another commonly used method as it allows parallel measurement of $\delta^{18}\text{O}$ and $\delta^2\text{H}$, but with often lower precision, at least early on (e.g., Voelker et al., 2015). As CRDS (and other laser techniques) have been commonly used for seawater isotope analysis since 2010, they contribute a significant part of the data produced globally (see for example, the ~~compilation of the released~~ CoralHydro2k seawater data base with a focus on the tropics (35°N-35°S) (Atwood et al., 202~~6~~5)).

The CISE-LOCEAN dataset merges data measured since roughly 1998 by IRMS switching to CRDS-based instrumentation since 2011. However, there are potential differences between the data produced by the two methods, discussed in Reverdin et al. (2022) in the context of the CISE-LOCEAN database. In this database, among the CRDS measurements, part of the samples collected in 2010-2015 ~~have been~~were first distilled, whereas more recent samples are directly introduced in the vaporizer associated with a Picarro ~~L2130~~L2130-CRDS (usually, L2130 model), through a salt-liner. Each sample was injected at least 6 times, and up to 10 times, and we dismiss most of the early injections when averaging them. For CRDS measurements, there are also issues of memory effects due to the ~~properties propensity~~ of water molecules to stick to the surfaces, and more so for $^1\text{H}^2\text{H}^{16}\text{O}$ than for the other isotopologs of the water molecule (cf fig. 4 of Steen-Larsen and Zannoni, 2024). In this version of the database, these memory effects have been roughly corrected based on regularly inserting ~~some a~~ reference material ~~regularly~~ (every three ~~seawater~~ samples) with isotopic composition close to ~~them~~those of the seawater samples, as well as by running duplicates of all samples at the beginning and end of each daily run, as well as of ~~for~~ samples with large standard deviations between the different injections ~~retained~~averaged,

Furthermore, an attempt was made to adjust the data from these three measurement methods (IRMS, CRDS with or without distillation), based on the analysis of Benetti et al (2017b). Uncertainties are largest in the earlier IRMS measurements, because of uncertainties on the reference material used at the time. As described in Reverdin et al (2022), this was partially corrected based on deep water ~~measurements~~samples

Mis en forme : Couleur de police : Automatique

measured at the time, for which we expect little variability over the last 30 years.

However, there remains subsets of $\delta^{18}\text{O}$ data produced by IRMS which could not be fully assessed. This is for example the case for surface samples collected in the equatorial Atlantic in August-September 1998 and 1999. Compared with more recent data, they

present a negative $\delta^{18}\text{O}$ offset ~~larger than~~ on the order of -0.150‰ . This ~~could be~~ is probably largely related to a calibration issue ~~or, but potentially also to~~ interannual variability. Nonetheless, the slope of the regression of $\delta^{18}\text{O}$ with respect to salinity of these 1998-1999 data is consistent with the slope for the more recent data, thus suggesting similar processes involved in the spatial property distribution in this earlier period. As there are no measurements ~~for in those years for~~ the other seasons for these years and region or other ~~isotopes~~, we have excluded those 1998-1999 data when analyzing the summer season in the equatorial Atlantic ~~region~~. Interestingly, the slope of the regression of $\delta^{18}\text{O}$ with respect to salinity of these 1998-1999 data is consistent with the one of the more recent data, thus suggesting similar processes involved in the spatial property distribution in this earlier period.

~~There was also~~ In the CISE-LOCEAN database, there is also an attempt to identify and flag potential issues of evaporation of the samples, happening mostly because of long storage in bottles which exchanged with the outside air, or defective bottles or caps (Reverdin et al., 2022). Although these issues are less common in the last 10 years of collected data, ~~this anomalies~~ still sometimes happens, in particular with samples extracted from water bottles for salinity measurements collected on Voluntary Observing Ships (VOS). ~~Although these VOS data tend to present less consistency, we retain them in the data base, because in some tropical subtropical areas, they contribute a significant share of the observations~~ (Alory et al, 2015).

The validation steps, as well as the large percentage of surface data points mostly in the open ocean, ~~away from coasts~~ in the CISE-LOCEAN data base (~~over close to 8200-9000~~ validated $\delta^{18}\text{O}$ near-surface data associated with a valid salinity since 1998) represent significant differences with respect to other published data compilations such as the “GISS Global Seawater Oxygen-18 Database -V1.21” for ~~stable~~ seawater stable isotopes which does not include the CISE-LOCEAN data (LeGrande and Schmidt, 2006; referred to as GISS-V1.21), or the CoralHydro2k seawater data base, which includes ~~only part an~~

earlier version of the CISE-LOCEAN data (Atwood et al., 20265). GISS-V1.21 is a compilation of different individual data sets originating mostly from ocean cruises (covering several decades) and measured by ~~Isotope-ratio Mass Spectrometry (IRMS)~~, for which there was an attempt to correct systematic biases between data subsets, ~~and~~ but includes fewer surface data points. The CoralHydro2k data base does not offer systematic correction of potential biases between different datasets, but it contains significant metadata that could be used for further evaluation.

Furthermore, the large proportion of surface data produced by CRDS in the LOCEAN data base (~~more than~~close to 870% of the surface data) offers the opportunity to investigate $\delta^2\text{H}$ and d-excess, in addition to $\delta^{18}\text{O}$, these two other variables been much less frequent in the GISS-V1-21 archive.

Originally, ~~t~~The ~~sampling samples~~ in the CISE-LOCEAN data base ~~were mostly collected during research cruises, often repeated in different years for the same season, as well as on ships of opportunity with collection by a LOCEAN-recruited ship rider (Benetti et al., 2017a), quite commonly in the North Atlantic subpolar gyre (Reverdin et al., 2018). More recently (mostly since 2018), the database also incorporates samples collected on vessels of opportunity (VOS) by a mariner or a sailor, and that often involves a later transfer of the sample from a salinity bottle to a dedicated vial. Although these VOS data tend to present less consistency, we retain them in the database, because they are the only source of data in most tropical-subtropical areas and in some seasons. The resulting data distribution is heavily quite skewed towards better sampling in some subregions of the ocean, such as the northern to equatorial the Atlantic Ocean and nearby seas, the south-western Indian Ocean and between Tasmania and Antarctica. In some of~~Even in most of these regions, except in the NASPG, there ~~is~~are usually some seasons with very limited or no sampling: for example, during boreal summer in the equatorial Atlantic, austral fall in the southwestern Indian Ocean, and late boreal fall to early spring south of Tasmania (App. A). For this last reason, we will not present results for the region~~seasonal sampling, albeit not much south of Tasmania, which will be ignored here, as well as other world regions. (Figure 1)~~which are less-well sampled in the CISE-LOCEAN database. Although IRMS surface data of $\delta^{18}\text{O}$ are less common, they still contribute significantly to the sampling in some regions and seasons (equatorial Atlantic in boreal summer, southwestern Indian Ocean in boreal winter, see App. A).

Mis en forme : Police :Symbol

Mis en forme : Expositant

is heavily skewed to some subregions of the ocean, such as the Atlantic Ocean and nearby seas, the south-western Indian Ocean and between Tasmania and Antarctica. In some of these regions, there is usually some seasonal sampling, albeit not much south of Tasmania, which will be ignored here, as well as other world regions (Figure 1).

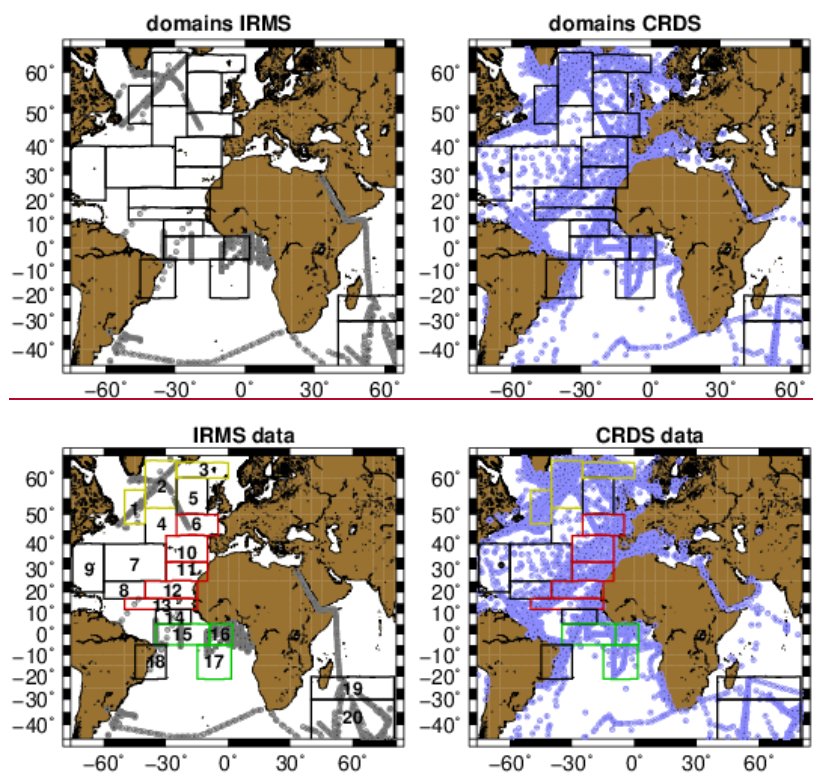


Figure 1: Maps of Wisotopes data distribution with the contours of the different domains boxes outlined with their reference number (the color of the outlines corresponds to region A (boxes 1-3, yellow), region B (red), and region C (green)). Left one panel for IRMS measurements, and right one for CRDS measurements. The black dot in the western subtropical Atlantic on the left-right panel marks is at the BATS station site. A large number of plotted data points are associated with more than one data point, for example when they correspond to repeated cruises, often happening at the same time of year (cf seasonal maps of data density in App. A).

We also ~~independently used-considered~~ surface samples collected mostly monthly ~~at~~ ~~close to~~ the BATS hydrographic site ~~near Bermuda~~ (31°50'N, 64°10'W) in 2015-2022. These samples stored in plastic bottles were frozen and stored at the University of Bergen (~~Norway~~) before been analyzed in 2023 ~~before been refrozen~~. They were ~~again~~ unfrozen in November 2025 to be transported to LOCEAN where they were analyzed between December 5th 2025 and January 11th 2026, ~~but-although~~ on a CRDS unit which experienced a set of technical problems reducing the accuracy. There are also indications that some of the samples had evaporated (five samples, in the Bergen analyses, and 19 samples in the LOCEAN analyses). This is due to the ~~se~~ plastic bottles not been completely proofed ~~against-to~~ air exchange when closed and not frozen.

Mis en forme : Exposant

Mis en forme : Exposant

2.2 Methods

~~$\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess will be considered which will be generically addressed as Wisotopes.~~ We aim at a finer scale analysis ~~of Wisotopes~~ than in Legrande and Schmidt (2006) where most of the surface North Atlantic is included in one single region, ~~when considering their relationship to salinity~~. Indeed, this should be viewed as a continuation of Benetti et al (2017)'s regional analysis, as well as water mass analyses such as in Frew et al. (2000) and Voelker et al. (2015), but with ~~only-astrict~~ focus to the surface- ~~(top 10 m)~~. We consider ~~sub-regionsboxes bound in latitudes and longitudes~~ (Figure 1), in which only samples ~~with afor which~~ salinity ~~is~~ within specified ranges are retained to avoid the freshest waters ~~on-shelves~~ directly influenced by strong local freshwater inputs (rivers, heavy rain ~~events~~, large sea ice melt) which can lead to very skewed salinity distributions (Bingham et al., 2002), ~~and possibly different relations between salinity and Wisotopes.~~

The ~~regions regional boxes are large enough in order to retain data throughout the seasonal cycle, but-should not too large to~~ correspond to identifiable oceanographic domains with respect to ocean circulation as well as for specific regimes of the water air-sea exchange (~~by~~ Evaporation (E) ~~—and~~ Precipitation (P)) and its seasonal cycle or of other freshwater sources. ~~In practice, the boxes have usually to be relatively large in order to retain data throughout the seasonal cycle-These constraints result in rather~~

large domains. Despite that, except in the NASPG boxes 1 to 3, there is often at least one season much less in the sampled than the others. central subtropical North Atlantic, as well as in the South Atlantic and south-western Indian Ocean, due to poor data coverage. In most of these regions, there is one main source of seawater advected through the region and subject to some

In each region, horizontal and vertical mixing between different water masses and their advective paths, as well as freshwater inputs and evaporation resulting in a spatial spread of surface salinity, as well as in $\delta^{18}O$. However, this is not always the case as discussed in. Sometimes the different water masses are not isotopically distinguishable in a [$\delta^{18}O$, salinity] diagram. For example, Benetti et al. (2017b) for the equatorial Atlantic region and the eastern subtropical gyre of the Atlantic Ocean, where could not distinguish the signatures of upwelling also contributes to the upwelled waters from the ones of the surface water masses advected from higher eastern subtropical latitudes in addition to the horizontally advected water. However, Benetti et al (2017a) could not clearly distinguish their respective properties in a ($\delta^{18}O$, salinity) space.

In each region, the data are averaged by month to estimate a seasonal cycle, with unresolved spatial or interannual variability in the sampling contributing to the standard deviation between data. The climatological salinity fields present spatial gradients, which likely also imprint the $\delta^{18}O$ fields (cf Legrande and Schmidt, 2006). With the limited and spatially inhomogeneous sampling, the spatial gradients contribute to a representativeness error on the monthly mean relative to the climatological regional mean. As discussed in App. A, in most regions and for a large part of the seasonal cycle, the spatial standard deviation of climatological monthly salinity contributes to a rather large representativeness error which exceeds 0.1 on salinity with the sampling of the CISE-LOCEAN data set. The notable exception is in the better sampled boxes of the North Atlantic subpolar gyre. In some regions, such as in the central Atlantic equatorial box, we also find that for some months with repeated cruises, the average salinity deviates strongly from the box average salinity, most likely due to the spatial sampling, thus inducing large differences in the salinity seasonal cycle.

To resolve the seasonal cycle of $\delta^{18}\text{O}$, it is useful to establish regional monthly relationships between $\delta^{18}\text{O}$ and S . In principle, the better-defined monthly spatial fields of salinity could then be used to reconstruct domain-averaged $\delta^{18}\text{O}$, adopting monthly linear relationships, as was done in LeGrande and Schmidt (2006) with large scale annual linear relationships between $\delta^{18}\text{O}$ and S , and assuming that these relationships do not vary interannually. However, $\delta^{18}\text{O}$ data are usually too scarce for estimating a linear relationship on a monthly climatological time scale in individual regions. However, to partially circumvent this issue of insufficient isotopic sampling, the compromise adopted here is to group in each box the regional data into four quarters (months 1-3, 4-6, 7-9, 10-12), which, as will be discussed in the results section, often-sometimes produces too uncertain linear regressions, in which case one resorts to estimate an annual relationship by averaging the linear relationships obtained for each of the four seasons. The caveat of this approach is that the resulting uncertainty on the average is strongly influenced by the less well sampled, more uncertain season. To reduce this uncertainty, we also attempted to average the seasonal relationships of nearby boxes in larger regions (regions A, B, C on Figure 1).

The data did not warrant going further than linear relationships, especially within the restricted salinity range retained. These are obtained by linear regression of $\delta^{18}\text{O}$ as a function of S with least square means in the (S , $\delta^{18}\text{O}$) space, thus minimizing non-explained residual variance in $\delta^{18}\text{O}$. To reduce the influence of rare-less common measurements data at the lowest or highest salinity as well as outliers, we also estimate a slope and 0-intercept (value of the regression line at $S=0$) by directly fitting the averages of two classes of data, one for the data below and one for the data above the seasonal salinity average.

The comparison of the two approaches indicates that the unbiased, but slightly noisier second approach often leads to a significantly larger slope. In the four North Atlantic subpolar gyre boxes (1, 2, 3, 4), this excess slope averaged ~4%, 6%, 10% of the mean slopes for $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess, respectively, whereas these increases tend to be smaller and less certain in the eastern mid-latitudes and northern tropics. The differences are usually largest for d-excess, and always are within the uncertainty estimates of the slopes in individual regions. Thus, for simplicity, we will only report the second estimate in the results section.

In each region, the isotopic data were also averaged by month to estimate a seasonal cycle, with unresolved spatial or interannual variability in the sampling contributing to the standard deviation between data points. The climatological monthly salinity fields present spatial gradients in each region, which likely also imprint on the $\delta^{18}\text{O}$ fields, assuming a quasi-linear relationship with respect to salinity. This, based on a limited and spatially inhomogeneous discrete sampling, contributes to a representativeness error on the monthly mean of the discrete samples. We test this effect in App. B using two climatological products for years rather close to the core of the isotopic data (described in Reverdin et al, 2007, and Boutin et al., 2021), with similar conclusions.

As discussed in App. B, in most regions and for a large part of the seasonal cycle, the spatial standard deviation of climatological monthly salinity contributes to a rather large representativeness error which often exceeds 0.1 on salinity with the sampling of the CISE-LOCEAN dataset. For example, in the central equatorial Atlantic (box 15 on Fig. 1), sampling in specific months (in particular March) (see App. A) is from repeated cruises along dedicated tracks resulting in a strong deviation of the average salinity of the discrete data points from the domain-average salinity. The month-to-month changes in sampling thus strongly contributes to the estimated box-averaged salinity seasonal cycle. The expectation is that this would also be the case to some extent for the $\delta^{18}\text{O}$. We will illustrate for two boxes (14 and 15) attempts to correct this by using the established linear relationships of $\delta^{18}\text{O}$ to salinity. The notable exception, where undersampling is less an issue, is in the three more regularly sampled boxes of the North Atlantic subpolar gyre (1, 2, 3 on Fig. 1), but where the seasonal cycle is, however, relatively weak.

3. Results

3.1 Monthly time series

In most regions, monthly standard errors are often large compared to the amplitude of the seasonal cycle. However, there are a few regions where there is sufficient agreement in the salinity seasonal cycle with the climatology, such as the central and western parts

of the NASPG, or the central Atlantic north-equatorial (5–12°N). This last region presents a good agreement with salinity climatologies, whereas the equatorial region (5°N–5°S) just to its south presented much less agreement, in part because of large month-to-month spatial differences in the sampling (cf App. A). Thus, we chose to illustrate the average seasonal cycles for surface salinity, $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess in these two contrasting equatorial regions with the monthly means and the standard deviations of the monthly samples (Fig. 2).

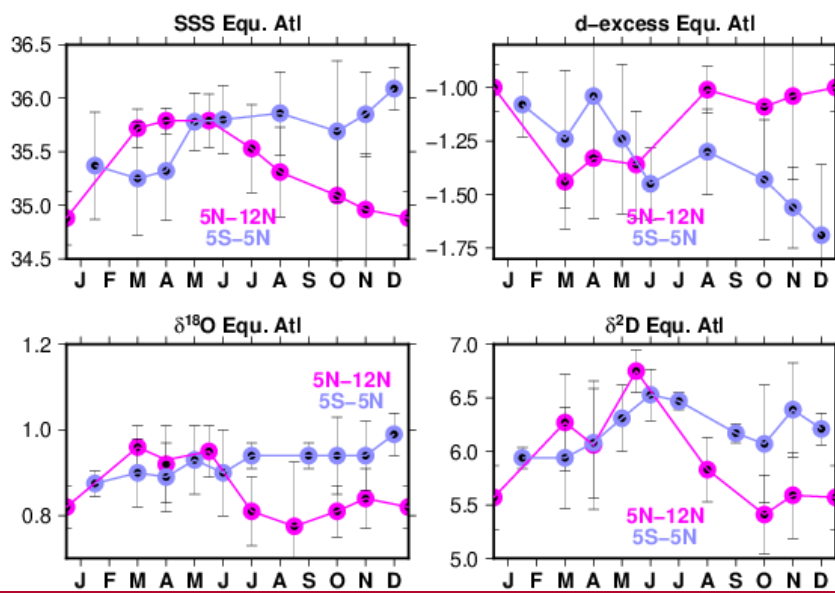


Figure 2: Average seasonal cycle in two central equatorial Atlantic regions for SSS (practical surface salinity, with no unit), $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess (from top to bottom, in ‰). The averages are usually provided monthly, except bi-monthly in a few instances with very insufficient monthly sampling. The standard deviation among the data is plotted (± 1 standard deviation).

$\delta^{18}\text{O}$ and $\delta^2\text{H}$ present a variability roughly in phase with SSS in the 5–12°N region, but less so in the 5°N–5°S region (Fig. 2). Furthermore, in each of the two regions, d-excess presents a variability opposite to the one seen in SSS. The two equatorial Atlantic regions are also roughly in phase opposition to each other, at least for SSS and d-excess.

Mis en forme : Français (France)

In months well sampled, for all variables, the standard error (not presented) is small relative to the amplitude of the seasonal cycle. Typically, for monthly $\delta^{18}\text{O}$ values in the equatorial domain, this estimated standard error varies between 0.01 and 0.10‰ (smallest in March and largest in July to September); whereas in January–February, July or August, there are not enough independent data to estimate a standard error. The standard error is typically 6 times larger for monthly $\delta^2\text{H}$ values and 4 times larger for d-excess values than for $\delta^{18}\text{O}$ values. In those two regions it is small enough to resolve a seasonal cycle for all isotopes, albeit not in all months. As mentioned above, for the equatorial region, there is the caveat illustrated on salinity (cf App. A), that part of this apparent seasonal variability is likely associated with seasonal differences in the spatial sampling.

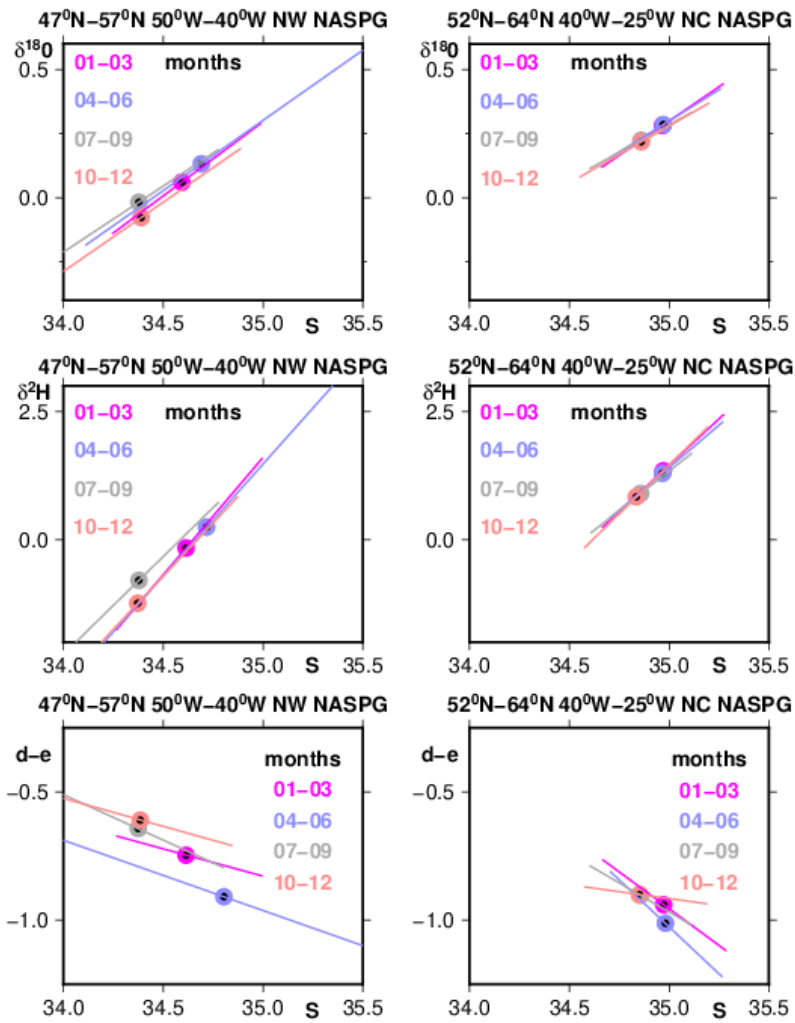
Mis en forme : Français (France)

3.2.1 Linear Seasonal linear regressions in well-sampled regional boxes (NASPG)

Mis en forme : Anglais (États-Unis)

In the particularly well-sampled box 2 (Irminger Sea and nearby areas), the residuals of individual data points with respect to the seasonal linear regression have a standard deviation of 0.046, 0.35 and 0.21 ‰ for $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess, respectively (averaged over the four seasons). In this box 2, which does not present a large spatial or temporal variability in salinity, a significant share of the variance of the individual data in a given season is explained by the mean linear regression to S for $\delta^{18}\text{O}$, $\delta^2\text{H}$ (Pearson correlation coefficient r of 0.69 and 0.67, averaged over the four seasons, corresponding to correlations which are always significantly different from 0 at $p=0.01$), but not so for d-excess (R^2 of 0.11). This encourages us to check whether there are seasonal differences in the regressions. Uncertainties are often large on individual seasonal regressions, in particular when sampling includes less than 10 individual data and for d-excess. This is due to a worse noise/signal ratio for this derived variable, with a smaller proportion of the variability explained by the regression for d-excess. In the particularly well-sampled region south-east of Greenland (Irminger Sea and nearby areas), the residuals of individual data with respect to the linear regression have a standard deviation of 0.046, 0.35 and 0.21 ‰ for $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess, respectively (averaged over the four seasons). In this north-central NASPG region, which does not present a large salinity spatial or

temporal variability, a significant share of the variance of the individual data in a given season is explained by the mean linear regression to S for $\delta^{18}\text{O}$, $\delta^2\text{H}$ (R^2 of 0.69 and 0.67, averaged over the four seasons, corresponding to correlations which are always significantly different from 0 at $p=0.01$), but not for d-excess (R^2 of 0.11). These proportions of course vary from region to region, but for $\delta^{18}\text{O}$, $\delta^2\text{H}$ they decrease towards the lower latitudes (and are as low as 20%–30% near the equator, thus corresponding to correlations which are only significantly different from 0 at $p=0.02$ or 0.05), and for d-excess, they are only high in the equatorial Atlantic domains where the mean regression explains roughly 40% of the variance in individual data (R^2 of 0.40, thus correlations significant at $p=0.01$).



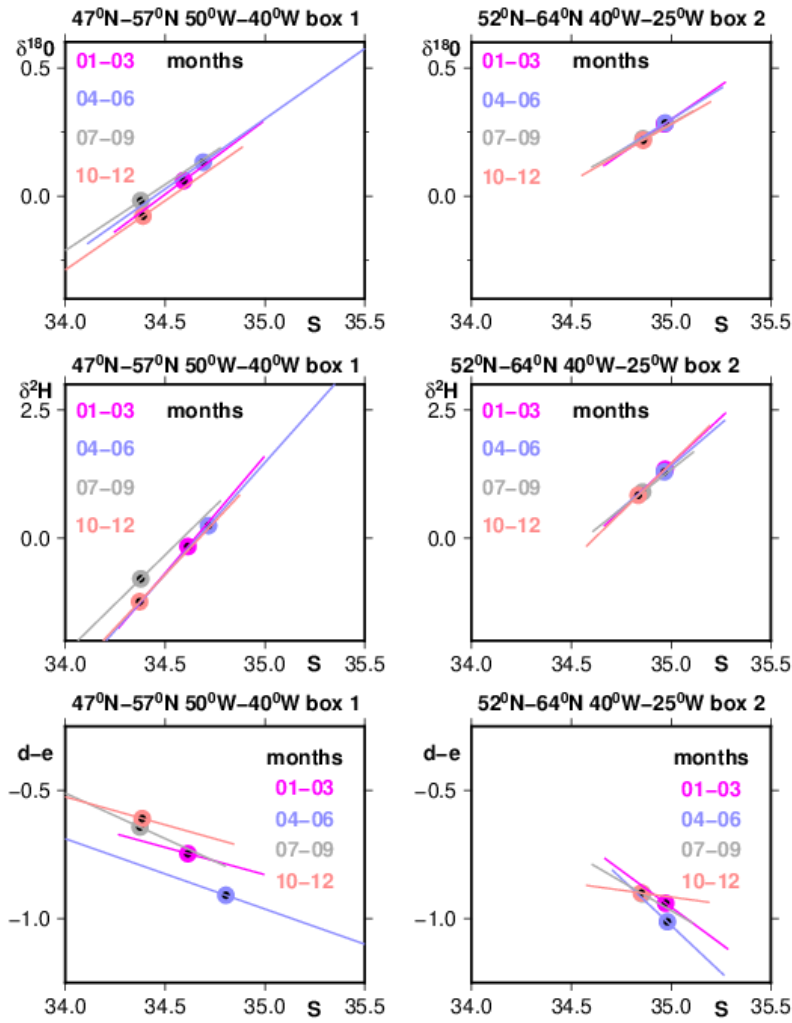


Figure 23: Seasonal linear fits of isotopes (‰) versus SSS (from top to bottom for $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess) in for two regions boxes (1 and 2 on Fig. 1) which are part of region grouping 1 in the north-western and central North Atlantic subpolar gyre. Data have only been retained for SSS within 34 to 35.5 and the linear fit is provided within the seasonal SSS range observed. The dots correspond to the averages, with their standard error always within the plotted dots for these two well sampled regions (for

box 1, the seasonal data number varies between 26 (months 4-6) and 77 (months 1-3);
whereas in box 2, it varies between 102 (months 1-3) and 225 (months 7-9)).

The seasonal distributions of the linear fits of Wisotopes to SSS ~~is-are~~ illustrated for two rather well sampled ~~regions-boxes~~ in the ~~western and~~ north-central ~~and western~~-NASPG (1 and 2 on Fig. 1) (Fig. 23). The average salinity cycle roughly corresponds to what is expected in these two regions (Reverdin et al., 2018) with lower summer-fall salinity and ~~higher-larger~~ winter-spring salinity. In both regions the different seasonal distributions for $\delta^{18}\text{O}$, $\delta^2\text{H}$ versus S nearly overlay each other with maybe a small negative isotopic shift in ~~months~~ 10-12 contrasting with a positive shift in ~~months~~ 7-9, at least for the westernmost region. ~~There are also slightly larger slopes in months 1-3. The latter was attributed to input of sea ice melt to the surface water (Reverdin et al., 2018; Benetti et al., 2016). There are also slightly larger slopes in season JFM, which are attributed to a contribution of depleted brine release for the water samples at the lowest salinities in this season (Benetti et al., 2016).~~

The seasonal d-excess versus S distribution also suggests differences according to the seasons, with d-excess apparently shifted to lower values in the spring season (4-6). This is also found in another region with reasonable sampling south-east of Iceland. The d-excess deviations across seasons exceed the uncertainty on the averages. For instance, assuming that all data ~~points~~ provide independent realizations, the standard errors on the d-excess averages in season 4-6/ 7-9 are respectively 0.04/0.03, and 0.015/0.016‰ for the western and north-central regions of the NASPG (1 and 2 on fig. 1), respectively. There is also a suggestion of a ~~lowweaker~~ slope in season 10-12 with slopes altogether ~~weaker-lower~~ in the western than in the central NASPG ~~region(1 and 2 on figure 1).~~ However, the estimated standard errors for these d-excess versus S seasonal slopes are large and are estimated to be on the order of 0.15 ‰/psu in ~~season 10-12 as well as for the other-all~~ seasons. Thus, even in these well-sampled regions, the differences between seasons in the slope of d-excess versus S linear relationships are within the error bars.

3.2 Seasonal linear regressions in other regions

Except for the well-sampled boxes in the NASPG, uncertainties are often large on seasonal regressions, in particular when sampling includes less than 10 individual

Mis en forme : Anglais (États-Unis)

Mis en forme : Paragraphe de liste, Hiérarchisation + Niveau : 2 + Style de numérotation : 1, 2, 3, ... + Commencer à : 2 + Alignement : Gauche + Alignement : 0,63 cm + Retrait : 1,29 cm

Mis en forme : Police :Couleur de police : Noir

Mis en forme : Retrait : Gauche : 0,63 cm

data points, in particular for d-excess. For this derived variable, this is due to a worse noise/signal ratio, with a smaller proportion of the variability explained by the regression to salinity. These proportions vary between boxes, but for $\delta^{18}\text{O}$, $\delta^2\text{H}$ the part of the variance explained by the linear regressions tends to decrease towards the lower latitudes. The lowest values (20%-30%) are near the equator, thus corresponding to correlations which are only significantly different from 0 at $p=0.02$ or 0.05). On the other hand, for d-excess, the linear fit holds best in the equatorial Atlantic boxes where it explains on the order of 40% of the variance in individual data points (R^2 of 0.40, and correlations significant at $p=0.01$).

3.3 Spatial distribution of linear regressions

For more than half the regions, the uncertainties on the individual seasonal slopes (and 0-intercepts estimated by the extrapolation of the regression line to $S=0$) are larger than between them. To portray the spatial variability in the slopes and 0-intercepts, these seasonal slopes (0-intercepts) are first averaged over the years, retaining only those based on more than 3 individual data.

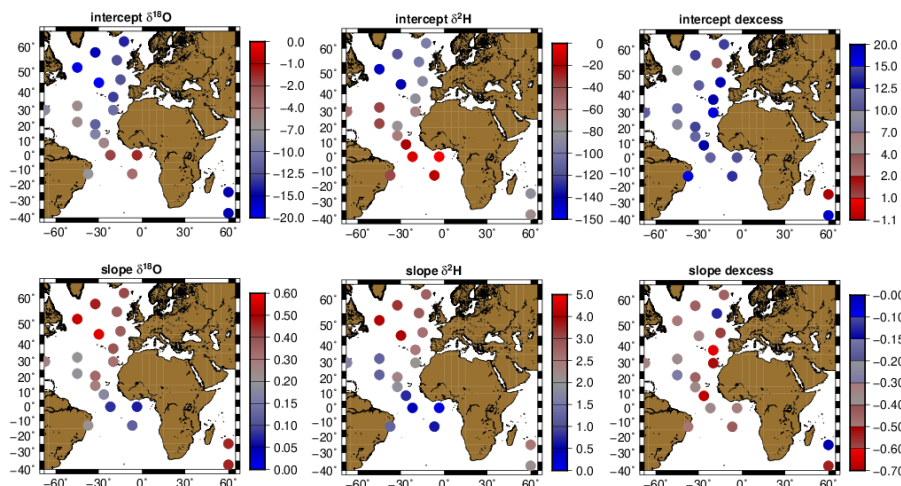


Figure 4: Annual averages of the slopes (‰/pss) and 0-intercepts (‰) of Wisotopes as a function of SSS in the different domains. These annual averages are obtained by averaging the regressions for the 4 seasons, except in the South Indian Ocean, where the data only sample three ($\delta^{18}\text{O}$) or two ($\delta^2\text{H}$ and -excess) seasons. Left for $\delta^{18}\text{O}$, middle for

Mis en forme : Police :Couleur de police : Noir

$\delta^2\text{H}$, and right for d-excess (cf table 1 for standard errors on these averages, estimated from the dispersion between the different seasonal averages).

Table 1: Annual averages of the slopes (‰/pss) and 0-intercepts (‰) of isotopes as a function of SSS in the different domains. Left for $\delta^{18}\text{O}$, middle for $\delta^2\text{H}$, and right for d-excess. These averages (and their standard error in parentheses) are obtained by averaging the regressions for the 4 seasons, except in the South Indian Ocean (last two lines) where they are obtained by averaging only three ($\delta^{18}\text{O}$) or two ($\delta^2\text{H}$ and d-excess) seasons, respectively. Left for $\delta^{18}\text{O}$, middle for $\delta^2\text{H}$, and right for d-excess. Regions with large errors usually have at least one season with poor sampling. Slopes and 0-intercepts are willingly reported with more digits than their uncertainties, in order to be able to reproduce the linear fit to the distribution of data (centered on a mean S very different from $S=0$).

Region center	$\delta^{18}\text{O}$		$\delta^2\text{H}$		d-excess	
	Slope	0-intercept	slope	0-intercept	slope	0-intercept
1 45°W/52°N	-0.544 (0.013)	-18.77 (0.44)	4.03 (0.17)	-146.1 (5.7)	-0.26 (0.04)	8.4 (1.2)
2 32°W/58°N	-0.473 (0.023)	-16.27 (0.83)	3.45 (0.16)	-119.3 (5.7)	-0.47 (0.13)	15.4 (4.6)
3 12°W/62°N	-0.409 (0.058)	-14.0 (2.0)	2.87 (0.45)	-98.9 (16.0)	-0.36 (0.07)	11.6 (2.3)
4 30°W/45°N	-0.579 (0.025)	-20.0 (0.9)	4.24 (0.18)	-147.1 (6.5)	-0.42 (0.14)	13.7 (5.1)
5 12°W/55°N	-0.386 (0.050)	-13.1 (1.7)	3.09 (0.54)	-106.5 (18.8)	-0.14 (0.10)	3.6 (3.5)
6 15°W/47°N	-0.354 (0.036)	-11.9 (1.3)	2.57 (0.23)	-87.5 (8.3)	-0.63 (0.24)	21.3 (8.8)
7 45°W/32°N	-0.199 (0.012)	-6.1 (0.4)	1.21 (0.15)	-36.6 (5.5)	-0.37 (0.09)	11.7 (3.1)
8 45°W/22°N	-0.196 (0.074)	-6.1 (2.8)	1.27 (0.57)	-39.5 (21.4)	-0.26 (0.32)	7.5 (11.8)
9 67°W/30°N	-0.270 (0.018)	-8.6 (0.7)	1.82 (0.13)	-58.1 (4.8)	-0.29 (0.16)	9.1 (6.0)
10 20°W/37°N	-0.385 (0.024)	-13.0 (0.9)	2.45 (0.13)	-83.0 (4.7)	-0.66 (0.20)	22.4 (7.2)
11 20°W/30°N	-0.328 (0.057)	-11.0 (2.1)	2.13 (0.44)	-71.2 (16.1)	-0.48 (0.08)	16.0 (2.7)
12 35°W/21°N	-0.339 (0.049)	-11.4 (1.8)	2.29 (0.35)	-77.4 (13.0)	-0.42 (0.08)	13.6 (3.0)
13 32°W/14°N	-0.275 (0.034)	-9.0 (1.2)	1.87 (0.30)	-61.3 (10.7)	-0.36 (0.09)	11.3 (3.1)
14 26°W/8°N	-0.194 (0.054)	-6.0 (1.9)	0.79 (0.38)	-22.4 (13.5)	-0.74 (0.17)	25.1 (6.1)
15 22°W/0°N	-0.073 (0.010)	-1.7 (0.3)	0.38 (0.07)	-7.5 (2.5)	-0.36 (0.03)	11.7 (1.1)
16 4°W/0°N	-0.061 (0.030)	-1.3 (1.1)	0.18 (0.03)	-0.5 (0.9)	-0.31 (0.06)	9.6 (1.9)
17 7°W/13°S	-0.119 (0.022)	-3.3 (0.8)	0.62 (0.03)	-16.1 (0.9)	-0.49 (0.02)	16.3 (1.6)
18 35°W/13°S	-0.223 (0.022)	-7.0 (0.8)	1.27 (0.23)	-39.2 (8.6)	-0.50 (0.07)	16.6 (4.9)
19 60°E/25°S	-0.481 (0.10)	-16.4 (3.4)	2.45 (0.23)	-82.3 (8.2)	0.00 (0.05)	-1.4 (3.8)
20 60°E/37°S	-0.442 (0.032)	-5.0 (1.2)	2.17 (1.02)	-72.7 (36.2)	-0.54 (0.20)	18.0 (13.7)

The average slopes of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ as a function of SSS (Fig.4, Table 1) are largest in the northern and western NASPG, as well as South Indian Ocean (for $\delta^{18}\text{O}$), whereas they are smallest near the equator. Slopes are also smaller in the western subtropical North

Atlantic than in the eastern part. The 0-intercepts follow the patterns of the slope (most negative in northern NASPG and south Indian Ocean).

For d -excess, the pattern is less clear with larger relative errors. Negative slopes are however largest in the eastern Atlantic (10°N – 45°N), as well as in the southern region of the South Indian Ocean, where it contrasts with weak values further north (but within the error bars). The d -excess 0-intercepts present a similar pattern (opposite signs), with large positive 0-intercepts for large negative slope values. The overall anti-correlation of slope and 0-intercept could be real, but could also arise to some extent from noise, as $\text{SSS}=0$ is very far from the range of salinity data in which data were retained, and thus an extrapolation based on a noisy slope estimate would induce an anti-correlated noise. Clearly the error estimates are large and sampling is insufficient at least in some of the seasons in a large part of the cases.

3.4 Seasonal cycle in the regressions

Thus, in order to retrieve a significant seasonal signature on the slopes and 0-intercepts, one needs to reduce the impact of sampling on the seasonal uncertainties. To do that, we will group ~~regional the regressions for individual boxes~~ estimates into larger domains, based on commonality in the sources of water, and whether evaporation dominates over precipitation and other meteoric waters or the opposite. We in particular identify one North Atlantic subpolar gyre ~~domain region A~~ (4 regions {grouping boxes 1 to 4 from Table 1; yellow boxes in Fig. 1), with usually excess precipitation, a freshwater input from the Arctic and potential contribution of sea ice melt), one east Atlantic ~~domain region B~~ (from mid-latitudes to northern tropics) encompassing {5 ~~regions-boxes~~ (6, 10 to 13 from Table 1; red on Fig. 1), with usually excess evaporation and limited local continental ~~Meteoric-meteoric~~ inputs}, and one equatorial and south-equatorial ~~domain region C with~~ {3 ~~regions-boxes~~ (15 to ~~17~~ from Table 1; green on Fig. 1), with an upwelling season, seasonal excess precipitation, usually excess precipitation, and possible influence of continental ~~m~~Meteoric water}. Based on that, we also attempted to create a western subtropical Atlantic ~~domain region~~, and in the eastern Atlantic to separate the northern tropics from the more northern latitudes. However, the uncertainties were then too large in some seasons for these ~~more-restricted~~ smaller domains regions, so that we discarded such groupings.

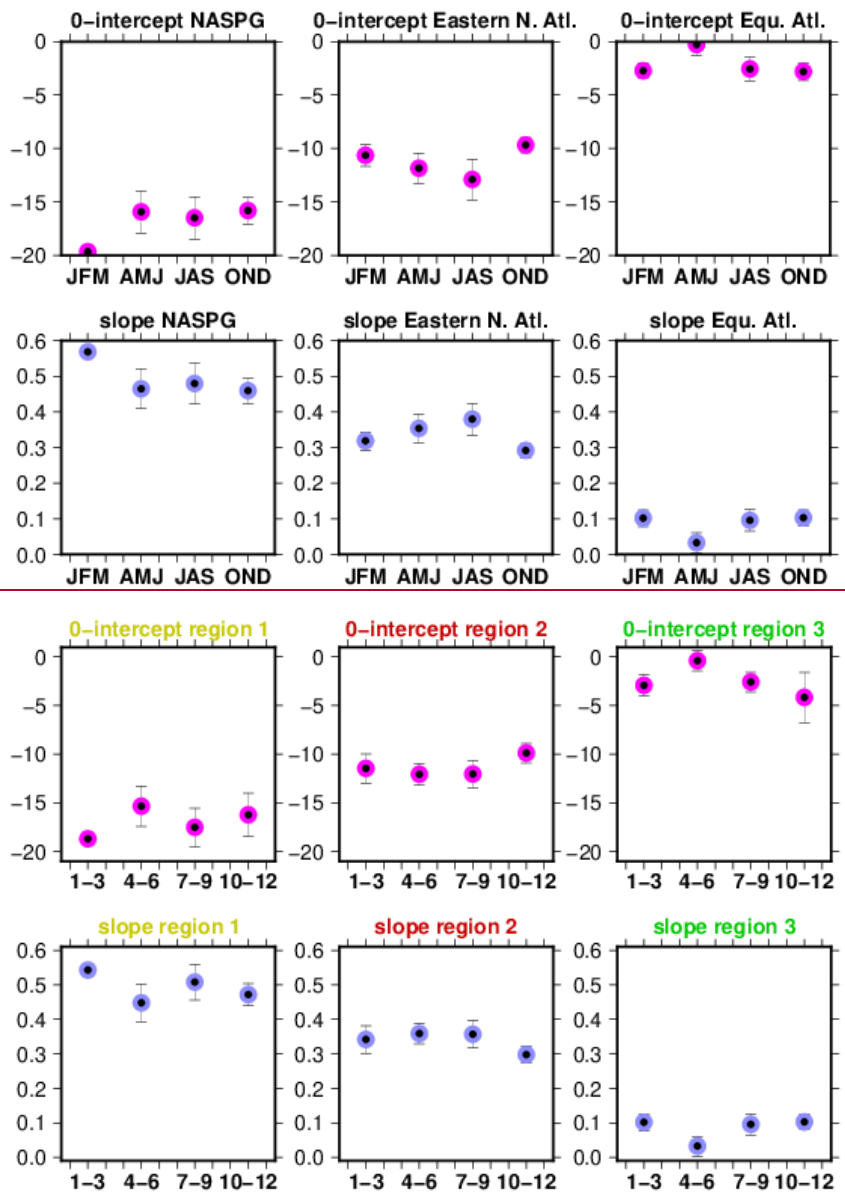


Figure 53: For the larger grouping of regions from table 2), slopes (upper panels, in ‰/psu) and 0-intercepts (lower panel, in ‰) from the seasonal regressions of $\delta^{18}\text{O}$ versus SSS for the four seasons in the three regions reported in table 1. The dots are the averages with added standard error bars. –From left to right: domain region 1-A for the

North Atlantic Sub-polar gyre (yellow boxes in Fig. 1), region 2-B for the eastern North Atlantic (red on Fig. 1), and region 3-C for the equatorial Atlantic south of 5°N (green in Fig. 1).

Table 21: domain-averaged slopes and 0-intercepts (standard error in parenthesis) from the regressions of $\delta^{18}\text{O}$ versus SSS for the four seasons (domain-region A4 for the North Atlantic Sub-polar gyre, 2-B for the eastern North Atlantic, and C3 for the equatorial Atlantic south of 5°N. Seasons 1 to 4 are respectively for months JFM1-3, AMJ4-6, JAS7-9, and OND10-12. Note that there are missing season 3 and 4 data for region 3, season 3, there are not enough data points in $\delta^2\text{H}$ and d-excess, and in season 4, region 17 is not included in the average values for $\delta^2\text{H}$ and d-excess in some of the regions and parameters of domain 3). Units are ‰/pss for all slopes and per mil for all intercepts.

Domain season	$\delta^{18}\text{O}$		$\delta^2\text{H}$		d-excess	
	Slope	0-intercept	slope	0-intercept	slope	0-intercept
1 — 1	0.569 (0.012)	-19.65 (0.45)	4.01 (0.20)	-138.9 (7.0)	-0.35 (0.09)	11.3 (3.8)
1 — 2	0.465 (0.055)	-15.93 (1.95)	3.44 (0.58)	-118.9 (20.5)	-0.54 (0.10)	18.0 (3.4)
1 — 3	0.480 (0.056)	-16.49 (1.98)	3.51 (0.45)	-121.4 (15.7)	-0.32 (0.06)	10.2 (1.9)
1 — 4	0.460 (0.035)	-15.81 (1.27)	3.66 (0.27)	-126.6 (18.8)	-0.24 (0.07)	7.5 (2.4)
2 — 1	0.319 (0.025)	-10.64 (1.00)	2.14 (0.15)	-71.6 (5.5)	-0.47 (0.09)	15.5 (3.6)
2 — 2	0.354 (0.043)	-11.85 (1.50)	2.44 (0.37)	-82.3 (13.0)	-0.47 (0.06)	15.7 (2.3)
2 — 3	0.380 (0.050)	-12.90 (1.90)	2.49 (0.32)	-84.6 (13.5)	-0.55 (0.18)	18.5 (6.5)
2 — 4	0.292 (0.020)	-9.67 (0.70)	1.97 (0.16)	-65.8 (5.5)	-0.54 (0.20)	17.9 (7.4)
3 — 1	0.102 (0.024)	-2.72 (0.84)	0.45 (0.13)	-9.8 (4.3)	-0.35 (0.07)	11.2 (2.9)
3 — 2	0.033 (0.032)	-0.26 (1.00)	0.40 (0.15)	-8.2 (5.5)	-0.45 (0.04)	14.8 (1.9)
3 — 3	0.096 (0.034)	-2.55 (1.12)				
3 — 4	0.103 (0.022)	-2.80 (0.85)	0.28 (0.14)	-4.4 (3.9)	1	0.543 (0.017) -18.69 (0.65) 3.97 (0.24) -137.7 (8.5) -0.37 (0.13) 11.9 (4.8)
A — 2	0.448 (0.055)	-15.34 (2.07)	3.31 (0.54)	-114.3 (19.1)	-0.46 (0.14)	14.9 (3.1)
A — 3	0.508 (0.052)	-17.51 (1.99)	3.66 (0.45)	-126.7 (15.9)	-0.35 (0.08)	11.2 (2.3)
A — 4	0.472 (0.032)	-16.22 (2.22)	3.77 (0.25)	-130.6 (8.9)	-0.16 (0.11)	4.4 (4.0)
B — 1	0.342 (0.052)	-11.40 (1.90)	2.26 (0.38)	-75.8 (13.4)	-0.46 (0.11)	15.2 (3.9)
B — 2	0.359 (0.034)	-12.07 (1.25)	2.49 (0.27)	-83.8 (12.1)	-0.45 (0.10)	14.7 (3.7)
B — 3	0.354 (0.047)	-11.92 (1.70)	2.30 (0.34)	-77.9 (12.5)	-0.52 (0.19)	17.4 (7.0)
B — 4	0.295 (0.031)	-09.75 (1.15)	1.98 (0.21)	-66.1 (7.6)	-0.42 (0.12)	13.4 (4.5)
C — 1	0.107 (0.033)	-2.91 (1.20)	0.46 (0.17)	-10.5 (5.6)	-0.40 (0.16)	13.2 (5.5)
C — 2	0.036 (0.030)	-0.38 (1.03)	0.44 (0.15)	-9.3 (5.1)	-0.46 (0.06)	14.0 (1.8)
C — 3	0.097 (0.028)	-2.57 (1.02)				
C — 4	0.140 (0.065)	-4.15 (2.30)	0.28 (0.14)	-4.4 (3.9)	-0.51 (0.18)	16.6 (6.5)

These domain-regional composites (Fig. 53, Table 12) do not often show very large/significant seasonal variability of the slopes/intercepts, but there are some hints

Mis en forme : Police :12 pt

Mis en forme : Police :12 pt

Mis en forme : Police :12 pt

Mis en forme : Police :12 pt

Mis en forme : Taquets de tabulation : 1,25 cm, Gauche + 2,5 cm, Gauche + 3,75 cm, Gauche + 4,99 cm, Gauche + 6,24 cm, Gauche + 7,49 cm, Gauche + 8,74 cm, Gauche + 9,99 cm, Gauche + 11,34 cm, Gauche

of signals. For example, ~~for in the NASPG composite (domain 1, regions 1 to 4 of~~
~~table~~ northern region A (yellow on Fig. 1) for $\delta^{18}\text{O}$, the slope is ~~slightly high~~ larger and
intercept more negative in ~~JFM~~ season 1 than during the rest of the year, albeit the
difference is ~~barely-not~~ significant at the 90% confidence level (Fig. 54). On the other
hand, table ~~12 (domain 1)~~ for d-excess in domain A indicates a more negative slope and
a larger 0- intercept ~~larger~~ in AMJ season 2, and the opposite for season 4 ~~(thus for a~~
~~different season that the winter deviation reported for $\delta^{18}\text{O}$)~~, but these differences are
~~not also barely~~ significant at the 90% confidence level. Furthermore, these extrema
happen for different seasons than for $\delta^{18}\text{O}$.

In ~~the equatorial domain~~ region C (equatorial Atlantic south of 5°N, green in Fig. 1), the
slope for $\delta^{18}\text{O}$ is lower and the 0-intercept less negative in AMJ season 2 than during the
other seasons, ~~a-with the opposite in season 4~~, differences also barely significant at the
90% confidence level. On the other hand, there is no significant seasonal signal in the
composite for region B (eastern Atlantic, red in Fig. 1).

3.3 Average linear regressions

For more than half of the boxes, the uncertainties on the individual seasonal slopes (and
0-intersects estimated by the extrapolation of the regression line to S=0) are larger than
the differences between these seasonal values. To portray the spatial variability in the
slopes and 0-intersects, the individual seasonal slopes (0-intersects) are first averaged.
In order not to bias the average between seasons, we accept a seasonal estimate, if it is
based on at least 4 data points in at least 3 separate years. On the other hand, this choice
to retain seasons with very small sampling can induce large standard errors on the
coefficients when there is one season with this very low sampling (such as in regional
boxes 5, 8, 9, 17, cf App. A). In boxes 19 and 20 (SW Indian Ocean), these annual
averages are obtained by averaging only three ($\delta^{18}\text{O}$) or two ($\delta^2\text{H}$ and d-excess) seasons,
respectively, as the other seasons have no data.

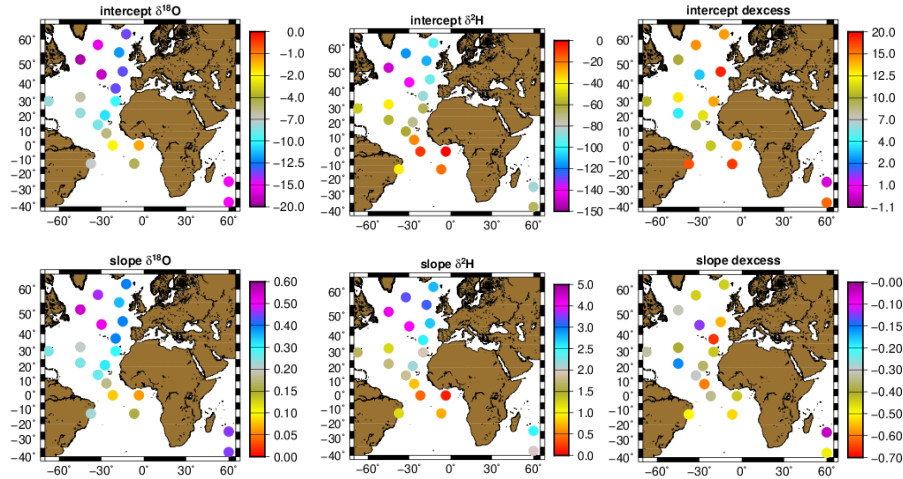


Figure 4: Annual averages of the slopes (‰/pss) and 0-intercepts (‰) of Wisotopes as a function of SSS in the different domains. Left for $\delta^{18}\text{O}$, middle for $\delta^2\text{H}$, and right for d-excess (cf table 2 for standard errors on these averages, estimated from the dispersion between the different seasonal averages).

Table 2: Annual averages of the slopes (‰/pss) and 0-intercepts (‰) of Wisotopes as a function of S in the different boxes. Left for $\delta^{18}\text{O}$, middle for $\delta^2\text{H}$, and right for d-excess. These averages (and their standard error in parentheses) are obtained by arithmetically averaging the linear regression coefficients obtained for the seasons with data. Minimum salinity threshold as well as the total number of $\delta^{18}\text{O}$ data points are also reported. Slopes and 0-intercepts are reported with more digits than their uncertainties, in order to be able to reproduce the linear fit to the distribution of data points (centered on a mean S very different from $S=0$)

Domain center	Smin	n	$\delta^{18}\text{O}$		$\delta^2\text{H}$		d-excess	
			Slope	0-intercept	slope	0-intercept	slope	0-intercept
1 45°W/52°N	34	232	0.555 (0.013)	-19.13 (0.46)	4.26(0.16)	-147.6 (5.5)	-0.27 (0.03)	8.5 (1.1)
2 32°W/58°N	34	658	0.479 (0.024)	-16.46 (0.86)	3.44 (0.13)	-119.0 (5.0)	-0.48 (0.14)	15.7 (4.8)
3 12°W/62°N	34	202	0.407 (0.051)	-13.9 (1.8)	2.87 (0.43)	-98.9 (15.3)	-0.37 (0.05)	11.9 (1.9)
4 32°W/47°N	34.5	50	0.531 (0.044)	-18.25 (1.6)	4.15 (0.24)	-143.9 (8.6)	-0.15 (0.14)	4.2 (4.8)
5 15°W/55°N	34.5	119	0.369 (0.050)	-12.6 (1.7)	3.40 (0.49)	-117.5 (17.1)	-0.14 (0.10)	3.6 (3.5)
6 15°W/47°N	35	159	0.399 (0.027)	-13.56 (1.0)	2.74 (0.24)	-93.4 (8.7)	-0.59 (0.07)	19.9 (2.5)
7 45°W/32°N	36	120	0.199 (0.010)	-6.1 (0.9)	1.17 (0.23)	-35.2 (8.6)	-0.41 (0.05)	13.3 (1.7)
8 45°W/22°N	36	53	0.256 (0.027)	-8.3 (1.0)	1.75 (0.32)	-57.4 (12.3)	-0.13 (0.25)	2.9 (9.8)

687	9 67°W/30°N	36 30	0.260 (0.031)	-8.3 (1.2)	1.76 (0.16)	-56.0 (6.6)	-0.29 (0.18)	9.0 (6.6)
688	10 20°W/37°N	35.5 159	0.399 (0.019)	-13.6 (0.7)	2.54 (0.12)	-86.1 (4.4)	-0.69 (0.18)	23.6 (6.8)
689	11 20°W/30°N	36 125	0.303 (0.028)	-10.1 (1.0)	1.98 (0.25)	-65.9 (9.2)	-0.41 (0.11)	11.3 (3.3)
690	12 27°W/21°N	36 98	0.318 (0.041)	-10.62 (1.5)	2.19 (0.33)	-73.8 (12.0)	-0.34 (0.05)	10.5 (2.2)
691	13 32°W/14°N	35 129	0.265 (0.046)	-8.6 (1.6)	1.84 (0.41)	-60.3 (14.7)	-0.28 (0.07)	8.5 (2.4)
692	14 26°W/8°N	34 93	0.197 (0.055)	-6.1 (2.0)	0.82 (0.35)	-23.2 (12.5)	-0.78 (0.22)	26.6 (7.8)
693	15 22°W/0°N	34 317	0.071 (0.006)	-1.6 (0.3)	0.39 (0.07)	-7.7 (2.7)	-0.36 (0.05)	11.4 (1.8)
694	16 4°W/0°N	34 168	0.054 (0.025)	-1.03 (0.8)	0.08 (0.11)	-0.5 (0.9)	-0.43 (0.12)	14.2 (4.4)
695	17 7°W/13°S	35.5 172	0.161 (0.045)	-4.9 (1.6)	0.65 (0.05)	-17.2 (1.8)	-0.60 (0.06)	20.3 (2.5)
696	18 35°W/13°S	36 87	0.228 (0.020)	-7.2 (0.7)	1.24 (0.27)	-38.3 (9.8)	-0.51 (0.05)	17.0 (1.9)
697	19 60°E/25°S	34.5 130	0.486 (0.008)	-16.56 (2.9)	2.43 (0.29)	-81.8 (8.7)	-0.06 (0.12)	0.8 (4.6)
698	20 60°E/37°S	35 350	0.440 (0.033)	-14.9 (1.2)	1.94 (1.20)	-64.5 (42.6)	-0.49 (0.21)	16.2 (7.4)
699								

The average slopes of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ as a function of S (Fig.4, Table 2) are largest in the North Atlantic subpolar Gyre boxes 1 and 2, as well as in the South Indian Ocean (boxes 19 and 20), whereas they are smallest near the equator. Slopes are usually also a little smaller in the western subtropical North Atlantic than in the eastern part. The 0-intercepts patterns mirror the patterns of the slope (with the opposite dependency: most negative in northern NASPG and south Indian Ocean).

For d-excess, the pattern is less clear with larger relative errors, in particular for some of the extreme values (Table 2). Negative slopes are however largest in the two tropical Atlantic boxes south of the equator (17 and 18) and are smaller just south of the NASPG (boxes 4 and 5), as well as in box 19 in the subtropical South Indian Ocean near 25°S. The d-excess 0-intercepts present a similar pattern to the one in the slope, with large positive 0-intercepts for large negative slope values. The overall anti-correlation of slope and 0-intercept could be real, but could also arise to some extent from noise, as $S=0$ is very far from the range of salinity in which data were retained, and thus an extrapolation based on a noisy slope estimate would induce an anti-correlated noise in 0-intercepts.

3.4 Monthly time series

In most regions, monthly standard errors are often large compared to the amplitude of the seasonal cycle, and part of the apparent seasonal cycle might result from sampling biases. However, there are a few regions where there is sufficient agreement in the salinity seasonal cycle with the climatology, such as in the central and western parts of the NASPG, or in the central Atlantic box 14 (5-12°N). The sampled S seasonal cycle in this last region presents a good agreement with monthly salinity climatologies, whereas

in box 15 to its south (5°N-5°S) there were larger differences, in part because of large month-to-month spatial differences in the sampling (cf App. B). We chose to illustrate the average seasonal cycles for surface salinity, $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess in these two contrasting equatorial regions with the monthly means and the standard deviations of the monthly samples (Fig. 2). In addition, the smaller estimated standard errors are commented in App. B for S these two regions. For the equatorial region, we also estimate a seasonal cycle free of the spatial sampling bias, based on the expected S-bias (from App. B), and the annual S-Wisotopes linear relationships of Table 2

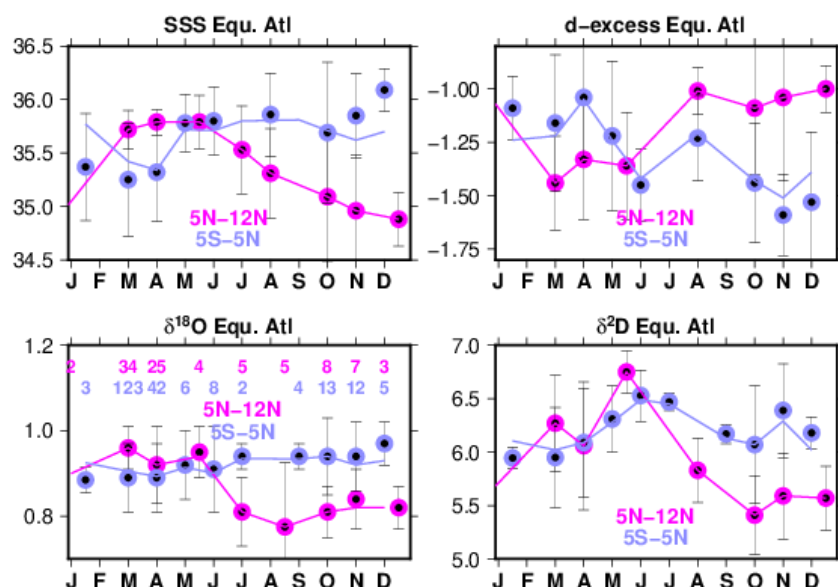


Figure 5: Average seasonal cycle in two central equatorial Atlantic boxes (14 and 15) for S (practical surface salinity, with no unit), $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess (in ‰). The averages are usually provided monthly, except in a few instances with insufficient monthly sampling, when bi-monthly sampling is retained. The standard deviation among the data points is plotted (as $\pm 1\sigma$), and the monthly data number indicated for $\delta^{18}\text{O}$ lower left panel. For the 5°S-5°N box 15, the blue curves are estimated box-average seasonal cycles with a removal of the spatial sampling bias.

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Police :Symbol

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Police :Symbol

Mis en forme : Exposant

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

The two boxes of Fig. 5 present a large contrast in their seasonal cycle. The ones in box 14 are much larger than the ones in box 15, especially after the correction applied to remove the spatial sampling bias of the box. In this box 15, the standard deviation of seasonal salinity diminishes when applying the correction from 0.27 to 0.15 for S, 0.027 to 0.015‰ in $\delta^{18}\text{O}$, from 0.19 to 0.16‰ in $\delta^2\text{H}$ and from 0.19 to 0.14‰ in d-excess. Despite the small values, the correlations between S and other isotopes variables diminish when applying the spatial bias. They remain significant at the 90% confidence interval for $\delta^{18}\text{O}$ (Pearson correlation coefficient of 0.71), but not so for $\delta^2\text{H}$ (0.60) or d-excess (-0.46), albeit it remains negative as it was before removing the spatial bias (-0.78). The lack of correlation for d-excess is not surprising with the estimated standard errors (cf App. B) and the very limited sampling in some months, e.g. in months 7-9. In box 14 (5-12°N), the correlations with S are more significant for $\delta^2\text{H}$ (Pearson correlation coefficient of 0.88) or d-excess (-0.87), but not for $\delta^{18}\text{O}$ (0.73).

This is fairly typical of what one could state in most other regions and where the seasonal cycle in salinity is rather weak, with the exception of three boxes in the North Atlantic subpolar gyre. We also present in App. C, the seasonal cycle of the BATS time series monitored over 7 years. This is somewhat different from the box averages, as there is no spatial representativity error to correct for in this case. Standard errors come from the meso-scale and interannual variability, as well as poor sample conservation and measurement uncertainties (see discussion in App. C). A comparison was done with box 9 results, but this box is the worse-sampled box and spatial representativity errors to be corrected are large. Nonetheless, at 0-order and with standard errors comparable to the observed seasonal cycle, this agrees to what is obtained at the BATS station. This suggests a limit of what can be obtained in very low sampling areas, the eastern Atlantic.

4. Discussion

Frew et al (2000) indicate for the NASPG a rather steep slope of the regression line of $\delta^{18}\text{O}$ to salinity (0.59‰/pss) and a rather low 0-intercept (-23‰) indicative of a large contribution of Arctic freshwater origin of the meteoric waters. These values are close but a little larger than the ones we find in the surface waters, but mostly in the western part of the NASPG, and with a smaller uncertainty. The surface data in Voelker et al

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

(2016) a little further south (30–40°N) in the eastern Atlantic indicate a somewhat lower slope (0.32 ‰/pss) and a higher 0 intercept, somewhat similar to what we find in collocated regions 14–16 (similar agreement found for $\delta^2\text{H}$). Benetti et al (2017a) which also includes $\delta^2\text{H}$ and d-excess data show spatial variability in the regression line of $\delta^2\text{H}$ with salinity between some of the same regions of the mid-latitude to equatorial Atlantic that are explored here with a widely extended data set. Benetti et al (2017a) explained to some extent the spatial variability from mid-latitudes to the equator through properties of precipitation and evaporative fluxes, and the relative contribution of the two, with for example the central equatorial areas dominated by precipitations not very depleted in heavy isotopes, thus inducing a small slope in the distributions of $\delta^{18}\text{O}$, $\delta^2\text{H}$, to salinity. We also found differences in the linear regressions between the eastern and western tropical/subtropical North Atlantic, albeit sampling in the CISE-LOCEAN database is certainly not very good in the western region. Based on very few data, this was also commented in Benetti et al. (2017a) who interpreted the shift of the distribution to higher $\delta^{18}\text{O}$, $\delta^2\text{H}$ as a likely result of input from western equatorial Atlantic freshwater (in particular from the Amazon-Orinoco discharge and near-equatorial rainfall) in addition to more depleted local freshwater sources. This difference in properties is also confirmed with an analysis of the BATS hydrographic station surface data in the western subtropical North Atlantic.

None of these earlier papers as well as the mapping presented in LeGrande and Schmidt (2006) discuss the seasonal variability in surface- $\delta^2\text{H}$ data in the regions investigated here away from shelves and continental slopes, because the weaker sampling and larger scatter in the earlier datasets would only allow to separate the largest scales (and only for $\delta^{18}\text{O}$).

4.1 Seasonal linear regressions

~~Here, we expanded on these results due to~~ To some extent, the qualified surface data in the updated CISE-LOCEAN database allow to make some step towards resolving the seasonal time scales. ~~To first order, the results we present are coherent with these earlier studies.~~ Sampling remains very low in large areas (Fig. 1), with often too large scatter in the data. There are some regions with large deviations in the regression to salinity: for example, west of the British Isles, the relationships are different possibly

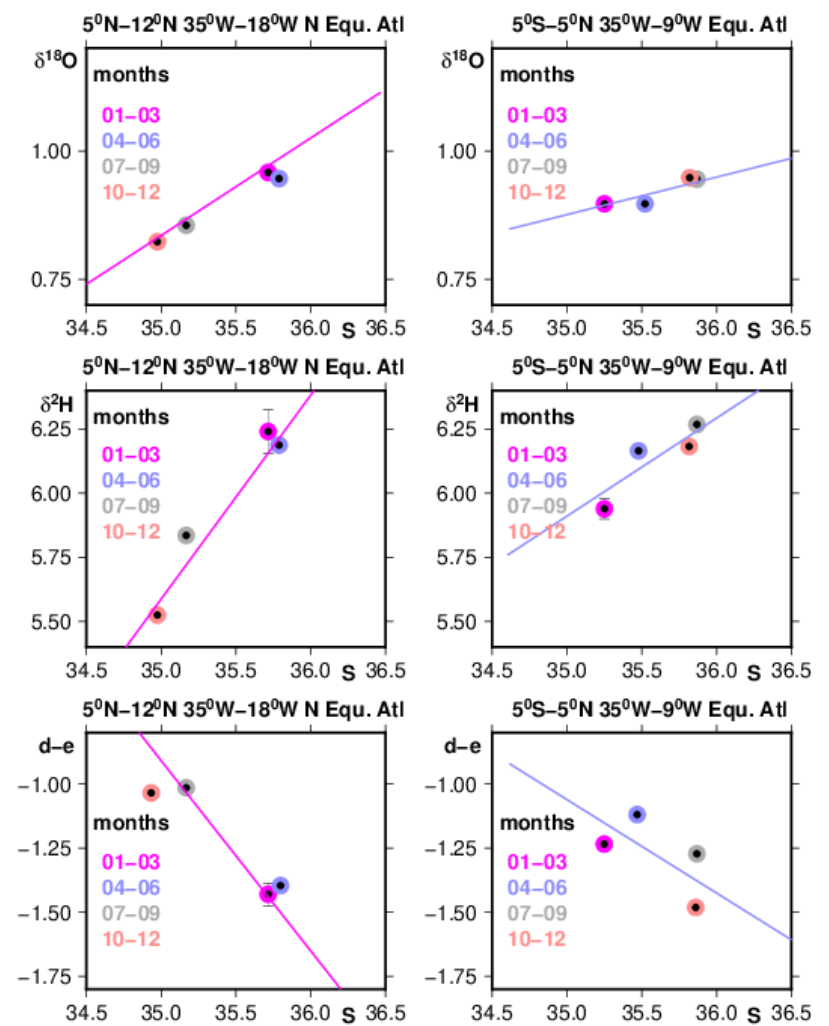
because there are multiple sources of freshwater mixing up with North Atlantic ~~Drift~~ waters from the subtropics: input from the fresher NASPG waters with an Arctic water input, on one hand, but also less depleted freshwater from rain and runoff from the British Isles and western Europe. Importantly, this region is ~~poorly~~ insufficiently sampled, so ~~part of these observations deviations~~ could also be artefacts due to insufficient sampling. Similarly, in the 20-30°S band of the southwestern Indian Ocean, we also find rather uncertain seasonal regressions (Table 2). we find some unexpected regressions. Based on Reverdin et al (2025) (and publications therein), this could originate from a combination of two main freshwater sources to this mostly excess-evaporation region, a first one very depleted which is advected from the surface sub-Antarctic austral ocean with southern latitude characteristics, and a second one less depleted carried by the Indonesian-Pacific throughflow from the western equatorial Pacific and maritime continent to the southwestern tropical Indian Ocean. In the present study, we have not tried to separate the two and could not test whether there is a seasonal dependence to it.

~~We also found differences in the linear regressions between the eastern and western tropical/subtropical North Atlantic, albeit sampling in the CISE-LOCEAN database is certainly not very good in the western region. Based on very few data, this was also commented in Benetti et al. (2017a) who interpreted the shift of the distribution to higher $\delta^{18}\text{O}$, $\delta^2\text{H}$ as a likely result of input from western equatorial Atlantic freshwater (in particular from the Amazon-Orinoco discharge and near equatorial rainfall) in addition to more depleted local freshwater sources. This difference in properties is also confirmed with an analysis of the BATS hydrographic station surface data in the western subtropical North Atlantic.~~

~~In this study, we examine the seasonal changes in Wisotopes.~~ In the central equatorial Atlantic, when comparing the Wisotopes averages in the 5°N-12°N regionbox (14) with the ones in the 5°N-5°S regionbox (15), we find well defined seasonal cycles in both regions, albeit with. ~~However, they have~~ very different timings of maxima and minima (Fig. 52). This is expected from the seasonal meridional displacements of the Inter-tropical convergence Zone (ITCZ) and associated changes of ocean circulation and equatorial upwelling, as discussed for ~~SSS~~ (Reverdin et al., 2007; da Allada et al., 2013; Yu et al., 2021). The ITCZ is furthest south in January-March, with a later initiation of the

Mis en forme : Couleur de police : Automatique

seasonal equatorial upwelling ~~at the equator~~ in May, simultaneous with the northward displacement of the ITCZ, as well as the setting of the eastward NECC later on (July-August) bringing water from the Amazon plume in addition to local precipitation to the central equatorial Atlantic further north (box 14). This ~~w~~eould have induced in season 10-12 of box 14 ~~at the higher 0 intercept fo~~changes in $\delta^{18}\text{O}$ and $\delta^2\text{H}$ ~~intercepts~~, and ~~at the~~ lower intercept for d-excess (not shown), compatible with a a signature seasonal influence of Amazon river discharge ~~characteristics in season 10-12 of the 5°N-12°N domain~~ (~~f~~For properties of water more strongly influenced by Amazon river properties, cf Benetti et al., 2017a). In each of the two regions, S, $\delta^{18}\text{O}$ and $\delta^2\text{H}$ vary mostly in phase, whereas d-excess varies in opposite phase (but not significantly in box 15) (Fig. 5), consistent with the average regression of these different properties with respect to salinity (Table 24).



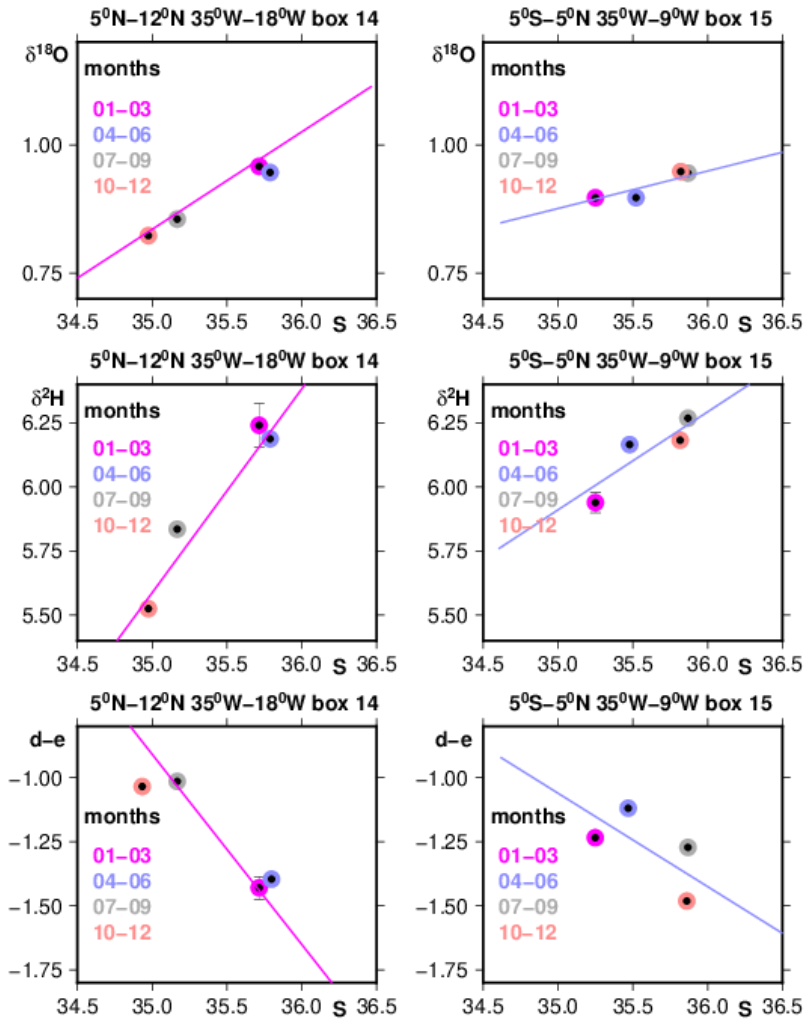


Fig. 6: Average linear regressions of Wisotopes (‰) as a function of S (the average of the four seasonal linear regressions) for two central Atlantic equatorial regions (left 5°N–12°N; right 5°S–5°N) displayed (from top to bottom for $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-e). The four average seasonal values are shown (without correction for the spatial representativity error) with their standard error bars (SE) smaller than the dots size, except in season 1-3.

Furthermore ~~As a complementary vision on what this entails,~~ the different seasonal averages for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ ~~nearly stand on~~ are plotted with the average of the seasonal regression lines (Fig. 6). ~~They nearly overlay it, suggesting it suggests~~ that the seasonal changes mostly correspond to the action of similar hydrological processes as the spatial spread during one season. This is somewhat different from the larger regional averages from Fig. 3 and Table 1, which includes box 15 and suggested a seasonal cycle in the slopes and 0-intercepts. This thus requires a little bit of caution on how spatially stable these relationships are within the larger region C, and whether what controls the spread of sampled values is the same as what controls the regional average. Notice ~~however~~ also that there is a suggestion of a seasonal difference in d-excess with points in seasons April-June 4-6 and July-September 7-9 (10-12, 1-3) to the right (left) of the average regression line for both regions (Fig. 6).

In the north-western Atlantic (1 and 2 in Fig. 1; Fig. 3, but this also holds for nearby regional boxes included in ~~regional grouping 1~~ region A of table 12), one finds suggestions of a larger winter regression slope (and more negative 0-intercepts) for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ (Fig. 2). This is reminiscent of the results in Benetti et al (2016), which they attribute to a contribution of ice formation (brine release), albeit we restrict here the data to salinities above 34, thus possibly missing ~~some a good part~~ of this signal, which is strongest at the lowest salinities and along the shelves and continental slopes. Interestingly, for d-excess, the largest anomaly happens less in winter than in spring, with the average d-excess and individual data points shifted to more negative values (at constant S), and with a larger negative slope (at least when averaged over the four ~~s 1 to 4 of seasons, as presented in~~ Table 21, grouped as ~~domain 1~~ region A of Table 12) and a larger positive 0-intercept. It would be tempting to relate that to advection of water affected by sea ice formation processes. However, what is published on sea ice properties during formation at least at equilibrium does not suggest major differences from a ratio of 8 in the fractionation for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ (Souchez et al., 1988), thus inducing no major change for d-excess in the remaining liquid sea water when freezing occurs. Thus, as salinity increases due to brine releases, this would induce a shift of the regression line to the right, thus above on Fig. 3. However, this is the opposite to what is observed. As these are areas with active winter mixing and air-sea exchange, although not to the same level in the four regional ~~domains~~ boxes, the changes might also be

attributed to those contributions. This would require a modelling approach to fully apprehend.

~~However, overall, in~~ most ~~other~~ regional ~~boxess~~ investigated, the seasonal differences observed in the regressions to salinity were not found to be significant, even after grouping (averaging) these different regions into larger domains. Thus, the seasonal variability is usually not of a similar magnitude as the large-scale spatial differences. This thus allows for the use of the surface Wisotopes data in comparisons of different databases/sets, even if these datasets have not been collected in the same season (as done in Reverdin et al., 2025). This does not preclude that some areas with large seasonal variability of advection, evaporation (E) and ~~P~~precipitation (p), and different sources of freshwater also present larger seasonal signals in the regional regression lines. This is probably what we observe in the central equatorial Atlantic in ~~the spring-~~months 03-04, and it is also likely the case in areas of the western equatorial Atlantic presenting very large seasonal (and interannual-~~variability~~) variability, due to the freshwater seasonal discharges mostly originating from the Amazon/Orinoco rivers. This has been observed ~~elsewhere~~in other oceans, such as in the western/central equatorial Pacific (Conroy et al., 2017; Horikawa et al., 2023). Higher data density would be required to establish potential seasonal and interannual variability in the regression lines. For example, in the South Indian Ocean, although there is a ~~rather good long~~ time series with often adequate spatial sampling for? austral ~~summer- (1-3)~~ conditions (Akhoudas et al., 2023), there are nearly 5 months in the year with no data in the CISE-LOCEAN database.

4.2 Average linear regressions

Although not always successful to identify seasonal variability in the relationship of Wisotopes with salinity, this dataset allows to portray rather well the spatial structure of these regressions. This usually fits well to what was earlier published seasonally. Frew et al (2000) indicate for the NASPG a rather steep slope of the regression line of $\delta^{18}\text{O}$ to salinity (0.59‰/pss) and a rather low 0-intercept (-23‰) indicative of a large contribution of Arctic freshwater origin of the meteoric waters. These values are close to

Mis en forme : Police :Italique

but a little larger to the ones we find in the surface waters, but mostly in the western part of the NASPG, and with a smaller uncertainty. Part of the difference originates in that we don't retain the lowest salinities typical of the shelves and slopes in the regressions. The surface data in Voelker et al (2016) a little further south (30-40°N) in the eastern Atlantic indicate for $\delta^{18}\text{O}$ a somewhat lower slope (0.32 ‰/pss) and a higher 0-intersect, somewhat similar to what we find in collocated regions 14-16 (similar agreement found for $\delta^2\text{H}$).

Benetti et al (2017a) who also reports on $\delta^2\text{H}$ and d-excess from a subset of the data used here show spatial variability in the regression line of isotopes with salinity between some of the same regions of the mid-latitude to equatorial Atlantic that are explored here with a widely extended dataset. Benetti et al (2017a) explained to some extent the spatial variability from mid-latitudes to the equator through properties of P and E, and the relative contribution of the two, with for example the central equatorial Atlantic dominated by precipitations not very depleted in heavy isotopes, thus inducing a small slope in the distributions of $\delta^{18}\text{O}$, $\delta^2\text{H}$, to salinity. They additionally present an east to west contrast in the slopes at the North Atlantic tropical and subtropical latitudes, which they attribute in part to the advected contribution of the Amazon-Orinoco discharge, as well as low latitudes precipitations, to surface water masses in the western Atlantic, contrasting to the high latitude North Atlantic contribution to the surface water masses in the eastern part of the basin. Our study also finds that contrast, and suggests that this is not related to seasonal differences in the sampling used in Benetti et al (2017a), albeit sampling in the western subtropical Atlantic is not large. On the other hand, the lowest regression slope for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ is also supported by the BATS hydrographic station surface data in the western subtropical North Atlantic (see App. C).

4.3 Monthly time series

When establishing box time series with limited spatial sampling, we face the issue of spatial representativeness errors in the sampling. An attempt was done to correct it by using the annually-averaged regression lines in the same box, which resulted in reduced seasonal variance in isotopes in equatorial Atlantic box 15. On the other hand, there is the caveat that the larger scale seasonal regression line (domain 3) shows a large variability, albeit marginally significant (Fig. 3). Using this modulation of regression line

in box 15 does not significantly modify the seasonal cycle in $\delta^{18}\text{O}$ and $\delta^2\text{H}$, with slightly lower resulting values for months 11-12 and higher values in months 1-3, but this could have a more noticeable impact in other boxes. On the other hand, with the limited sampling, there is also the likelihood that insufficiently sampled interannual to multi-decadal variability might result in a temporal representativity error. For example, in ~~In other areas presented here, there is less evidence for seasonal or interannual variability. In~~ the South Indian Ocean, multi-decadal trends have been described both in the southwestern subtropical region and south of the Subantarctic front, which have been ascribed to changes in the hydrographic cycle and in particular in local ~~evaporation and precipitation~~E and P (Akhoudas et al, 2023).

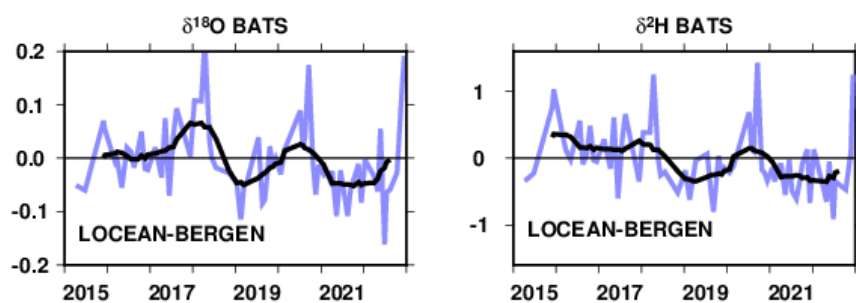
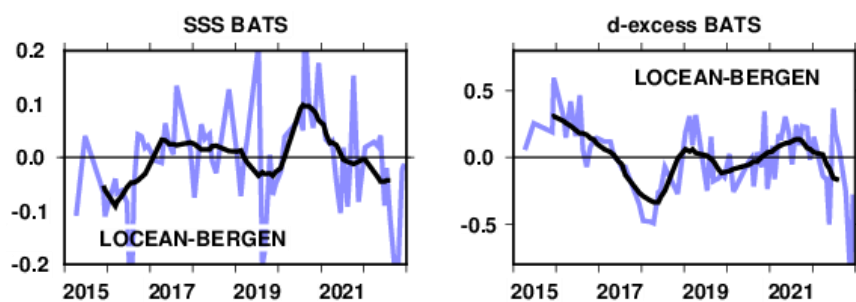
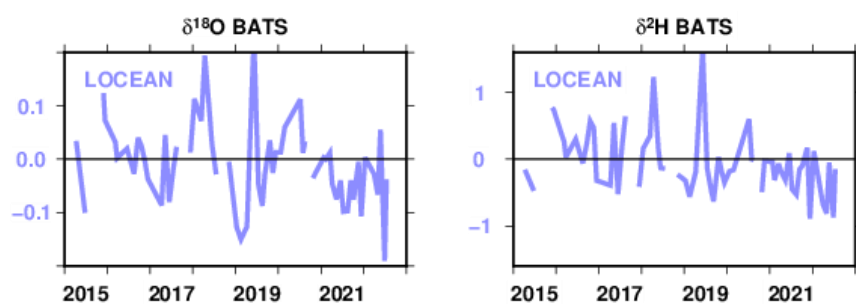
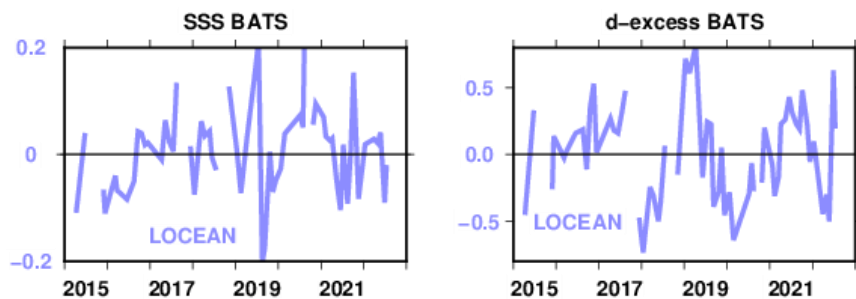


Fig. 7: Anomalized time series of ~~SSS-sea surface salinity~~ (pss), $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess (‰) at BATS (31°50'N, 64°10'W). ~~A version where averages of non-flagged LOCEAN and Bergen data is shown (see App. C). The blue curve connects individual data, whereas the black curve is a 1-year running mean of the data. Only non-flagged LOCEAN data are shown.~~

The BATS site time series in the western North Atlantic subtropics (App. ~~CB~~) is one of the rare sites where there was a concerted effort to regularly gather data over more than 7 years. It shows a clear seasonal cycle in S, $\delta^{18}\text{O}$ and $\delta^2\text{H}$, but less so in d-excess, as sampling is still too ~~limited with uncertainties in the measurements~~ (App. ~~CB~~).

Anomalizing the data by subtracting the monthly-averaged seasonal cycle removes a significant part of the variance. However, except for a few peak values ~~exceeding +/- 0.15-reaching 0.2‰ for $\delta^{18}\text{O}$, at times associated with salinity anomalies~~, the values are usually rather close to the ~~estimated standard errors, as well as -estimates and the possible magnitudes of systematic errors in the data. Even, with the one-year running mean (black curves), standard errors remain large, averaging 0.026 in S, and 0.018, 0.13, 0.06‰ on $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess, respectively.~~

~~Despite these large residual standard errors, Nonetheless, the LOCEAN version of the data-set suggests higher than normal $\delta^{18}\text{O}$ and $\delta^2\text{H}$ at times in 2016 in mid-2017 to 2018 (in particular the winter 2017-2018) and with lower values in 2019 and 2021. These do not seem to closely correspond to special anomalies in the surface salinity S variability. For example, notice in early 2016 the low salinities but no low $\delta^{18}\text{O}$ and $\delta^2\text{H}$. Data of The d-excess have an even worse signal over noise ratio, with rms errors in these analyses often exceeding 0.2‰, and further uncertainties associated with poor conservation of some of the samples. They however seem to show suggest a large contrast between very negative values in winter 2017-2018 the following winters 2017-2018 (negative) and positive values in 2021, as well as in late 2015/early 2016 and in 2019, which seem to mirror (with opposite sign) d 2018-2019 (positive), as well as the more positive values in 2021 the changes in salinity, but not so in 2022-. These short time series appear to partially co-vary with the North Atlantic Oscillation (NAO, Hurrell et al., 2003), with the winters 2014-2015, 2017-2018, and to a lesser extent early 2020, having a positive NAO index, in contrast to 2021 with a more negative NAO index~~

(positive NAO index resulting in increase in $\delta^{18}\text{O}$ and $\delta^2\text{H}$, and a decrease in d-excess). A longer times series and with smaller analytical errors would allow to further explore these possible relationships, as well as with other modes of climate variability.

Nonetheless, interpreting these the poorly resolved series-interannual variability would also require model simulations, in particular to check whether these result from separate the effect of changes in the hydrological cycle or from the effect of changes in oceanographic signals ocean circulation or surface stratification/upwelling signals (cf Ayache et al., 2024, but for another ocean region). as induced for example by modes of variability such as the North Atlantic Oscillation (NAO, Hurrell et al., 2003), with winter 2017–2018 having a positive NAO index, in contrast to 2021, having a more negative NAO index.

5. Conclusions

Although the current surface data in the CISE-LOCEAN data-base mostly collected during the last 25 years, and in particular since 2010, are still in limited number, they offer the possibility to characterize the seasonal cycle of Wisotopes and their relationship to surface salinity in part of the Atlantic Ocean and of the southwestern Indian Ocean. At 0-order in most of these regions, we suggest that the seasonal variations of the regression of Wisotopes to salinity are not very large compared to its spatial variability, and within the random and representativity errors. There are some noticeable seasonal deviations from the mean, in particular in the western North Atlantic subpolar gyre and in the equatorial region. We can better characterize the spatial distribution of this relationship than done before, which hints to small seasonal deviations from the yearly-averaged relationships both for the equatorial Atlantic and in the northwestern subpolar gyre of the North Atlantic. We also find a weak-suggestion of interannual variability at the regularly sampled BATS site in the western subtropical Atlantic.

We have not interpreted most of these deviations in a qualitative matter, relating them either to changes in ocean circulation, both horizontal or vertical mixing, or in the water fluxes between the atmosphere, the cryosphere, sea ice, and liquid sea water. In addition to the data, this would require dedicated simulations of forced isotopically-enabled

ocean models first validated by the observations (Ayache and Dutay, 2024). In the longer term, this might also require isotopically-enabled earth system models, as the isotopic signatures of these different processes (evaporation, precipitation, continental inputs) evolve in our warming world, as well as the melt of continental glaciers and ice sheets which will imprint on the seawater isotopic composition (e.g. Oppo et al., 2007). Additionally, surface seawater isotope data provide model boundary conditions to assess model performance in isotope-enabled Earth system models (e.g. Schmidt et al., 2007; Brady et al., 2019; Cauquoin et al., 2019).

The data are still very limited in individual data sets such as the CISE-LOCEAN one, the earlier GISS Global Seawater Oxygen-18 database or the CoralHydro2k Seawater $\delta^{18}\text{O}$ databases. There is an urgent need to put together efforts for better isotopically sampling the near-surface oceans, and thus better constraining processes of the hydrological cycle. This requires active intercomparison of data from different sources, with precise qualification of their differences and respective accuracy. (as currently performed on a smaller scale with the SCOR working group 171 – MASIS). There is hope with ongoing efforts of data retrieval and updating databases that this might be reachable in many parts of the world oceans, despite the obstacles of systematic differences and insufficient data coverage (Reverdin et al., 2025).

Data availability

The CISE-LOCEAN data updated to 2024 are available at <https://www.seanoe.org/data/00600/71186/>.

Code de champ modifié

Author contribution: GR initiated the study and prepared the manuscript with contributions from all coauthors. LG did a first analysis of the seasonal cycles and quality-controlled the CISE-LOCEAN data. HCSL contributed the data from the BATS site, and AV, CW, and HCSL contributed to editing the paper.

Competing interests: The authors declare that they have no conflict of interest.

Acknowledgments

1050 The LOCEAN isotopic laboratory is supported by 'Observatoire des sciences de l'Univers
 1051 Ecce Terra' of Sorbonne Université. We are thankful to Catherine Pierre and Jérôme
 1052 Demange who have set up and help run the facility, and for Aïcha Naamar, Marion
 1053 Benetti, Camille Akhoudas, and Florine Crossin to have measured some of the water
 1054 samples. We are grateful for support by INSU, by Claire Lo Monaco and Nicolas Metzl for
 1055 sampling during the OISO cruises on RV Marion Dufresne, and for the sample collection
 1056 during the French PIRATA and Ovide repeated cruises. We are also thankful to SNO SSS
 1057 (Gaël Alory) and the IRD IMAGO team (Denis Diverrès, Stéphane Jacquin, Linn Sekund),
 1058 for measuring salinity and providing some water from salinity samples, as well as to the
 1059 more than 100 ship riders on merchant and sailing vessels in the Atlantic Ocean, as well
 1060 as during hydrographic cruises who collected sea water isotopes samples. We thank the
 1061 maritime companies EIMSKIP and Royal Arctic Line, for having graciously hosted the
 1062 ship riders on their vessels, and to have facilitated their work on board. IPSL also
 1063 repeatedly supported the CISE-LOCEAN database collection and its qualification. The
 1064 surface salinity climatology is based on 14 years of satellite salinity from product CCI
 1065 V4.4, and we thank our colleagues Jacqueline Boutin, Anna Parracho and Fabrice
 1066 Bonjean for their support. The paper also benefited from relevant comments by the
 1067 two reviewers. This study has been supported by ESA CCI + salinity and CNES SMOS
 1068 projects. AV, through the Centro de Ciências do Mar do Algarve (CCMAR), acknowledges
 1069 receiving national funds from FCT- Fundação para a Ciência e a Tecnologia through
 1070 contracts UID/04326/2025 (DOI <https://doi.org/10.54499/UID/04326/2025>),
 1071 UID/PRR/04326/2025 (DOI <https://doi.org/10.54499/UID/PRR/04326/2025>) and
 1072 LA/P/0101/2020 (DOI:10.54499/LA/P/0101/2020). AV also acknowledges financial
 1073 support by Fundação para a Ciência e a Tecnologia (FCT) through projects Centro de
 1074 Ciências do Mar do Algarve (CCMAR) basic funding UIDB/04326/2020
 1075 (<https://doi.org/10.54499/UIDB/04326/2020>) and programmatic funding
 1076 UIDP/04326/2020 (<https://doi.org/10.54499/UIDP/04326/2020>) and the CIMAR
 1077 associated laboratory funding LA/P/0101/2020
 1078 (<https://doi.org/10.54499/LA/P/0101/2020>). The paper also contributes to the efforts
 1079 of SCOR working group 171, MASIS.

1080
 1081 References

Code de champ modifié

Code de champ modifié

- 1082 Akhoudas, C., Sallée, J.-B., Reverdin, G., Haumann, A., Pauthenet, E., Chapman, C. C.,
 1083 Margirier, F., Lo Monaco, C., Metzl, N., Meilland, J., and Stranne, C.: Isotopic evidence for
 1084 an intensified hydrological cycle in the Indian sector of the Southern Ocean, Nature
 1085 Comm., <https://doi.org/10.1038/s41467-023-38425-5>, 2023~~32~~.
 1086 Alory, G., Delcroix, T., Téchiné, P., Diverrès, D., Varillon, D., Cravatte, S., Gouriou, Y.,
 1087 Grelet, J., Jacquin, S., Kestenare, E., Maes, C., Morrow, R.M., Perrier, J., Reverdin, G.,
 1088 Roubaud, F.: The French contribution to the Voluntary Observing Ships Network of Sea
 1089 Surface Salinity, Deep Sea Res. I, 105, 1-18, doi:10.1016/j.DSR.2015.08.005, 2015.
 1090 [Aoki, S., Yamazaki, K., Hirano, D., Nomura, D., Murase, H.: Distribution of stable oxygen](#)
 1091 [isotope in seawater and implication on freshwater cycle off the coast from Wilkes to](#)
 1092 [George V Land, East Antarctica. Progress in Oceanography 217, 103101, doi:](#)
 1093 [10.1016/j.pocean.2023.103101, 2023.](#)
 1094 Atwood, A. R., Moore, A.L., Long, S., Pauly, R., DeLong, K., Wagner, A., and Hargreaves, J.A.:
 1095 The CoralHydro2k Seawater $\delta^{18}\text{O}$ Database, Past Global Changes Magazine 32,59, doi:
 1096 10.22498/pages.32.1.59, ~~2024~~2025.
 1097 [Atwood, A. R., Moore, A. L., DeLong, K. L., Long, S. E., Sanchez, S. C., Hargreaves, J. A.,](#)
 1098 [Morris, C. A., Pauly, R. E., Dassié, É. P., Felis, T., Voelker, A. H. L., Murty, S. A., and Cobb,](#)
 1099 [K. M.: The PAGES CoralHydro2k Seawater \$\delta^{18}\text{O}\$ Database: a FAIR-aligned compilation of](#)
 1100 [seawater \$\delta^{18}\text{O}\$ data to uncover “hidden” insights from the global ocean, Earth Syst. Sci. Data,](#)
 1101 [18, 1921–1941, 10.5194/essd-18-1921-2026, 2026.](#)
 1102 Ayache, M., and Dutay, J.-C., Mouchet, A., Tachikawa, K., Risi, C., and Ramstein, G.:
 1103 Modelling the water isotope distribution in the Mediterranean using a high-resolution
 1104 oceanic model (NEMO-MED12-watiso v1.0): evaluation of model results against in situ
 1105 observations, Geosci. Mod. Develop., 17, 6627-6655, [https://doi.org/10.5194/gmd-17-](https://doi.org/10.5194/gmd-17-6627-2024)
 1106 6627-2024, 2024
 1107 [Benetti, M., Reverdin, G., Pierre, C., Merlivat, L., Risi, C., Steen-Larsen, H. C., and Vimeux,](#)
 1108 [F.: Deuterium excess in marine water vapor: dependency on relative humidity and](#)
 1109 [surface wind speed during evaporation. J. Geophys. Res. Atmos., 119, 584-593,](#)
 1110 [doi:10.1002/2013JD020535, 2014.](#)
 1111 Benetti, M., Reverdin, G., Pierre, C., Khaliwala, S., Tournadre, B., Olafsdottir, S., and
 1112 Naamar, A.: Variability of sea ice melt and meteoric water input in the surface Labrador
 1113 Current off Newfoundland, J. Geophys. Res. Oceans, 121(4), 2841–2855, 2016.

Code de champ modifié

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

- Benetti, M., Reverdin, G., Aloisi, G., and Sveinbjörnsdóttir, A.: Stable isotopes in surface waters of the Atlantic Ocean: indicators of ocean-atmosphere water fluxes and oceanic mixing processes, *J. Geophys. Res. Oceans*, 122, doi:10.1002/2017JC012712, 2017a.
- Benetti, M., Sveinbjörnsdóttir, A. E., Ólafsdóttir, R., Leng, M. J., Arrowsmith, C., Debondt, K., Fripiat, F., and Aloisi, G.: Inter-comparison of salt effect correction for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ measurements in seawater by CRDS and IRMS using the gas-H₂O equilibration method, *Marine chemistry*, doi:10.1016/j.marchem.2017.05.010, 2017b.
- Benway, H. M., and Mix, A. C.: Oxygen isotopes, upper-ocean salinity, and precipitation sources in the eastern tropical Pacific. *Earth and Planet. Sci. Lett.*, 224, 493-507, doi:10.1016/j.epsl.2003.05.014. 2004.
- Bingham, F. M., Howden, S. D., and Koblinsky, C. J.: Sea surface salinity measurements in the historical database, *J. Geophys. Res. Oceans*, 107, 10.1029/2000JC000767, 2002.
- Boutin, J., Chao, Y., Asher, W. E., Delcroix, T., Drucker, D., et al.: Satellite and In Situ Salinity: Understanding Near-Surface Stratification and Subfootprint Variability, *Bull. of the Amer. Meteor. Soc.*, 97 (8), pp.1391-1407. 10.1175/BAMS-D-15-00032.1, 2016.
- Boutin, J., Reul, N., Koehler, J., Martin, A., Catany, E., Guimbard, S., Rouffi, F., Vergely, J.-L., Arias, M., Chakroun, M., Corato, G., Estella-Perez, V., Hasson, A., Josey, S., Khvorostyanov, D., Kolodziejczyk, N., Mignot, J., Olivier, L., Reverdin, G., Stammer, D., et al: Satellite-Based Sea Surface Salinity Designed for Ocean and Climate Studies, *J. Geophys. Res. Oceans*, 126, 11, <https://doi.org/10.1029/2021JC017676>, 2021.
- Brady, E., Stevenson, S., Bailey, D., Liu, Z., Noone, D., Nusbaumer, J., Otto-Bliesner, B. L., Tabor, C., Thomas, R., Wong, T., Zhang, J., Zhu, J.: The connected isotopic water cycle in the Community Earth System Model Version 1, *J. Adv. Model. Earth Syst.*, 11, 8, <https://doi.org/10.1029/2019MS001663>, 2019.
- Cauquoin, A., Werner, M., Lohmann, G.: Water isotopes – climate relationships for the mid-holocene and preindustrial period simulated with an isotope-enabled version of MPI-ESM, *Clim. Past* 15, 1913-1937, <https://doi.org/10.5194/cp-15-1913-2019>, 2019.
- Conroy, J. L., Cobb, K. M., Lynch-Stieglitz, J., and Polissar, P. I.: Constraints on the salinity-oxygen isotope relationship in the central tropical Pacific Ocean. *Mar. Chem.*, 161, 26-33, <https://doi.org/10.1016/j.marchem.2014.02.001>. 2014
- Conroy, J. L., Thompson, D. M., Cobb, K. M., Noone, D., Rea, S., and LeGrande, A. N.: Spatiotemporal variability in the $\delta^{18}\text{O}$ -salinity relationship of seawater across the

Mis en forme : Ne pas ajuster l'espace entre le texte latin et asiatique, Ne pas ajuster l'espace entre le texte et les nombres asiatiques

Mis en forme : Anglais (États-Unis)

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

Mis en forme : Anglais (États-Unis)

Mis en forme : Anglais (États-Unis)

- 1146 tropical Pacific, *Paleoceanography*, 32, 484–497,
1147 <https://doi.org/10.1002/2016PA003073>, 2017.
- 1148 [Craig, H., and Gordon, L.: Deuterium and oxygen 18 variations in the ocean and the](#)
1149 [marine atmosphere. In: Presented at the Proceedings of the third Spoleto conference.](#)
1150 [Spoleto, Italy, pp 9-130, 1965.](#)
- 1151 Da-Allada, C. Y., Alory, G., du Penhoat, Y., Kestenare, E., Durand, F., and Hounkonnou, N.
1152 M.: Seasonal mixed-layer salinity balance in the tropical Atlantic Ocean: mean state and
1153 seasonal cycle, *J. Geophys. Res. Oceans*, 118, 332-345 ISSN 0148-0227,
1154 <https://doi.org/10.1029/2012JC008357>, 2013.
- 1155 [Dassié, E. P., Hasson, A., Khodri, M., Lebas, N., and Linsley, B. K.: Spatiotemporal](#)
1156 [variability in the South Pacific convergence zone fresh pool eastern front from coral-](#)
1157 [derived surface salinity data. J. Clim., 31, 3265-3288, https://doi.org/10.1175/jcli-d-17-](#)
1158 [0071.1, 2018.](#)
- 1159 [Delaygue, G., Jouzel, J., and Dutay, J.-C.: Oxygen 18-salinity relationship simulated by an](#)
1160 [oceanic general circulation model. Earth Planet. Sci. Let., 178, 113-123, 2000.](#)
- 1161
- 1162 [Dissard, D., Reichert, G. J., Menkes, C., Mangeas, M., Frickenhaus, S., and Bijma, J.: Mg/Ca,](#)
1163 [Sr/Ca and stable isotopes from the planktonic foraminifera *T. sacculifer*: testing a](#)
1164 [multi-proxy approach for inferring paleotemperature and paleosalinity.](#)
1165 [Biogeosciences, 18, 423–439, 10.5194/bg-18-423-2021, 2021.](#)
- 1166 [Dong, B., Kumar, R., Shirai, K., Drysdale, R. N., and Prendergast, A. L.: Advances in giant](#)
1167 [clam \(*Tridacnidae* spp.\) sclerochronology and sclerochemistry as a high-resolution](#)
1168 [palaeoenvironmental archive. Palaeogeography, Palaeoclimatology, Palaeoecology,](#)
1169 [661, 112576, https://doi.org/10.1016/j.palaeo.2024.112576, 2025.](#)
- 1170 Frew, R. D., Dennis, P. F., Heywood, K. J., Meredith, M. P., and Boswell, S. M.: The oxygen
1171 isotope composition of water masses in the northern North Atlantic, *Deep-Sea Res. I*, 47,
1172 2265-2286, 2000.
- 1173 Glaubke, R. H., Wagner, A., and Sikes, E. L.: Characterizing the stable oxygen isotopic
1174 composition of the southeast Indian Ocean, *Mar. Chem.*, 262, 177–185,
1175 <https://doi.org/10.1016/j.marchem.2024.104397>, 2024.
- 1176 Hennig, A., Mucciarone, D. A., Jacobs, S. S., Mortlock, R. A., and Dunbar, R. B.: Meteoric
1177 water and glacial meltwater in the southeastern Amundsen Sea: a time series from 1994
1178 to 2020, *The cryosphere*, 18, 791-818, <https://doi.org/10.519/tc-18-791-2024>, 2024.

Code de champ modifié

Code de champ modifié

Mis en forme : Anglais (États-Unis)

Code de champ modifié

Code de champ modifié

- 1179 Hilaire-Marcel, C., Kim, S. T., Landais, A., Ghosh, P., Assonov, S., Lécuyer, C., Blanchard, M.,
 1180 Meijer, H. A., and Steen-Larsen, H. C.: A stable isotope toolbox for water and inorganic
 1181 carbon cycle studies, *Nature Reviews Earth & Environment* 2 (10), 699-719, 2021.
 1182 Horikawa, K., Kodaira, T., Zhang, J., and Obata, H.: Salinity-oxygen isotope relationship
 1183 during an El Niño (2014-2015) in the southwestern Pacific and comparisons with
 1184 GEOSECS data (La Niña, 1973-1974), *J. Mar. Chem.*, 249,
 1185 <https://doi.org/10.1016/j.marchem.2023.104222>, 2023.
- 1186 Hurrell, J. W., Kushnir, Y., Ottersen, G., and Visbeck, M.: An overview of the North Atlantic
 1187 Oscillation. Edited by: J.W. Hurrell, Y. Kushnir, G. Ottersen and M. Visbeck, *Geophys.*
 1188 *Monog. Ser.*, 134, 35 pp, <https://doi.org/10.1029/134GM01>, 2005.
- 1189 Konecky, B. L. et al.: The Iso2k database: a global compilation of paleo- $\delta^{18}\text{O}$ and $\delta^2\text{H}$
 1190 records to aid understanding of Common Era climate, *Earth Syst. Sci. Data*, 12, 2261–
 1191 2288, <https://doi.org/10.5194/essd-12-2261-2020>, 2020.
- 1192 [Leduc, G., Sachs, J. P., Kawka, O. E., and Schneider, R. R.: Holocene changes in eastern](#)
 1193 [equatorial Atlantic salinity as estimated by water isotopologues, *Earth and Planetary*](#)
 1194 [Science Letters, 362, 151–162, <https://doi.org/10.1016/j.epsl.2012.12.003>, 2013.](#)
- 1195 LeGrande, A. N. and Schmidt, G. A.: Global gridded data set of the oxygen isotopic
 1196 composition in seawater, *Geophys. Res. Lett.* 33,
 1197 <https://doi.org/10.1029/2006gl026011>, 2006.
- 1198 [Mulitza, S., Boltovskoy, D., Donner, B., Meggers, H., Paul, A., and Wefer, G.: Temperature:](#)
 1199 [\$\delta^{18}\text{O}\$ relationships of planktonic foraminifera collected from surface water.](#)
 1200 [Palaeogeogr Palaeocl. 202, 143–152, \[https://doi.org/10.1016/S0031-\]\(https://doi.org/10.1016/S0031-0182\(03\)00633-3\)](#)
 1201 [0182\(03\)00633-3, 2003.](#)
- 1202 Oppo, D. W., Schmidt, G. A., and LeGrande, A. N.: Seawater isotope constraints on tropical
 1203 hydrology during the Holocene, *Geophys. Res. Lett.* 34, L13701,
 1204 <https://doi.org/10.1029/2007GL030017>, 2007.
- 1205 Phillips, H. E., and Joyce, T. M.: Bermuda's tale of two time series: Hydrostation S and
 1206 BATS, *J. Phys. Oceanogr.*, 37, 554-571, <https://doi.org/10.1175/JPO2997.1>, 2007.
- 1207 [Rebotim, A., Voelker, A. H. L., Jonkers, L., Waniek, J. J., Schulz, M., and Kucera, M.:](#)
 1208 [Calcification depth of deep-dwelling planktonic foraminifera from the eastern North](#)
 1209 [Atlantic constrained by stable oxygen isotope ratios of shells from stratified plankton tows,](#)
 1210 [J. Micropalaeontol., 38, 113–131, \[10.5194/jm-38-113-2019\]\(https://doi.org/10.5194/jm-38-113-2019\), 2019.](#)

Code de champ modifié

Code de champ modifié

Code de champ modifié

Code de champ modifié

- 1211 Reverdin, G., Kestenare, E., Frankignoul, C., and Delcroix, T.: In situ surface salinity in the
 1212 tropical and subtropical Atlantic Ocean. Part I. Large scale variability, Progress in Oceanogr.,
 1213 73, 3, 311-340, <https://doi.org/10.1016/j.pocean.2006.11.004>, 2007.
- 1214 Reverdin, G., Metzl, N., Olafsdottir, S., Racapé, V., Takahashi, T., Benetti, M., Valdimarsson,
 1215 H., Benoit-Cattin, A., Danielsen, M., Fin, J., Naamar, A., Pierrot, D., Sullivan, K., Bringas, F.,
 1216 and Goni, G., SURATLANT: a 1993-2017 surface sampling in the central part of the North
 1217 Atlantic subpolar gyre, Earth Sci. Sys. Data, 10, 1901-1924,
 1218 <https://doi.org/10.5194/essd-10-1901-2018>, 2018.
- 1219 Reverdin, G., et al.: The CISE-LOCEAN sea water isotopic database (1998-2021), Earth
 1220 Sci. Sys. Data, <https://doi.org/10.5194/essd-2022-34>, 2022.
- 1221 Reverdin, G., Waelbroeck, C. Voelker, A. H., and Meyer, G.: Technical note: Large offsets
 1222 between different datasets of seawater isotopic composition: an illustration of the need
 1223 to reinforce intercalibration efforts, Ocean Sci., 21, 567-575,
 1224 <https://doi.org/10.5194/os-21-567-2025>, 2025.
- 1225 Roche, D. M. and Caley, T.: $\delta^{18}\text{O}$ water isotope in the iLOVECLIM model (version 1.0) –
 1226 Part 2: Evaluation of model results against observed $\delta^{18}\text{O}$ in water samples, Geosci.
 1227 Model Dev., 6, 1493–1504, doi:10.5194/gmd-6-1493-2013, 2013.
- 1228 Salgueiro, E., Voelker, A. H. L., Martin, P. A., Rodrigues, T., Zúñiga, D., Froján, M., de la
 1229 Granda, F., Villaceros-Robineau, N., Alonso-Pérez, F., Alberto, A., Rebotim, A.,
 1230 González-Álvarez, R., Castro, C. G., and Abrantes, F.: $\delta^{18}\text{O}$ and Mg/Ca Thermometry in
 1231 Planktonic Foraminifera: A Multiproxy Approach Toward Tracing Coastal Upwelling
 1232 Dynamics, Paleoclimatology and Paleoclimatology, 35, e2019PA003726,
 1233 10.1029/2019PA003726, 2020.
- 1234 Schefuß, E., Schouten, S., and Schneider, R. R.: Climatic controls on central African
 1235 hydrology during the past 20,000 years, Nature, 437, 1003–1006,
 1236 10.1038/nature03945, 2005.
- 1237 Schmidt, G. A., LeGrande, A. N., and Hoffmann, G.: Water isotope expressions of intrinsic
 1238 and forced variability in a coupled ocean-atmosphere model, J. Geophys. Res. 112,
 1239 D10103, <https://doi.org/10.1029/2006jd007781>, 2007.
- 1240 Schouten, S., Ossebaer, J., Schreiber, K., Kienhuis, M. V. M., Langer, G., Benthien, A., and
 1241 Bijma, J.: The effect of temperature, salinity and growth rate on the stable hydrogen
 1242 isotopic composition of long chain alkenones produced by *Emiliania huxleyi* and
 1243 *Gephyrocapsa oceanica*, Biogeosciences, 3, 113–119, 2006.

Code de champ modifié

Code de champ modifié

Code de champ modifié

1244 Steen-Larsen, H.-C., and Zannoni, D.: A versatile water vapor generation module for
1245 vapor isotope calibration and liquid isotope measurements, Atmos. Measur. Tech., 17,
1246 14, <https://amt.copernicus.org/articles/17/4391/2024/2024>, 2024.

1247 Voelker, A., Colman, A., Olack, G., Waniek, J. J., and Hodell, D.: Oxygen and hydrogen
1248 isotope signatures of Northeast Atlantic water masses, Deep-Sea Res. II, 116, 89-106.
1249 <https://doi.org/10.1016/j.dsr2.2014.11.006>, 2015.

1250 [Wilke, L., Meggers, H., and Bickert, T.: Depth habitats and seasonal distributions of recent](#)
1251 [planktic foraminifers in the Canary Islands region \(29°N\) based on oxygen isotopes.](#)
1252 [Deep Sea Research Part I: Oceanographic Research Papers. 56, 89–106,](#)
1253 <https://doi.org/10.1016/j.dsr.2008.08.001>, 2009.

1254 Yu, L., Bingham, F. M., Dinnat, E. P., Fournier, S., Melnichenko, O., Tang, W., and Yueh, S.
1255 H.: Revisiting the global patterns of seasonal cycle in sea surface salinity, J. Geophys. Res.
1256 Oceans, 126, <https://doi.org/10.1029/2020JC016789>, 2021.

1257

Code de champ modifié

Code de champ modifié

Code de champ modifié

Appendix A Monthly data distributions (or bi-monthly...)

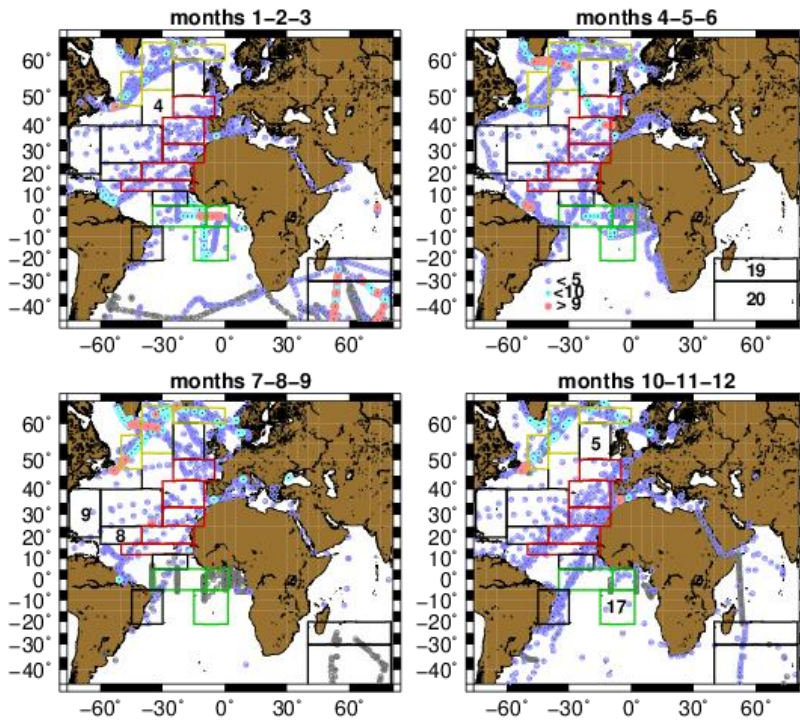


Figure A1: seasonal data distribution with box overlaid (color contours corresponding to the regions 1 - yellow, 2 - red, 3 - green). The grey dots are for IRMS data, whereas the color ones indicate CRDS data with a color scheme corresponding to data density within 25 km of the data (blue, less than 5 nearby data, cyan with 5 to 9 data, red, 10 or more data). For each season panel, the box numbers are indicated when there are 4 or less $\delta^{18}\text{O}$ data in the season.

The maps of data distribution (fig. A1 for $\delta^{18}\text{O}$, but except for IRMs data in grey in Fig. A1, maps for $\delta^2\text{H}$ and d-excess are rather similar) indicates strong inter-seasonal variability. One sees that mostly in regions with cruises repeated over many years in specific seasons (recent tropical Atlantic PIRATA cruises in months 3-4, Ovide NASPG cruises close to month 6, and most recent OISO SW Indian cruises in months 1-2). Except for the OISO cruises and a few mostly summer-time sail-boat samples (months 10-3), there are no data in the SW Indian Ocean (boxes 19, 20). In other areas, this is more random, and those boxes/seasons with virtually no data are specifically indicated on the

Mis en forme : Police :Symbol

Mis en forme : Exposant

1274 maps (4, 5, 8, 9, 17). Data density is high (here estimated as within a 25 km radius) along
1275 repeated cruises, and where there was a long-time concerted effort to collect water
1276 samples from ships of opportunity in the NASPG (SURATLANT project).
1277
1278
1279

Appendix A-B: Comparison of salinity box averages

We compare the box averages salinity seasonal cycle estimated from the LOCEAN data to climatological monthly fields of in situ salinity from Reverdin et al. (2007) updated to 1977-2016 and from the satellite-based CCI V4.4 SSS product averaged over 2011-2024 (Boutin et al., 2021) ([later referred to as CCI](#)). These two climatologies provide information on the effect of the limited spatial sampling in the boxes on the monthly box averages. The results with these two climatologies ~~that~~^{which} differently span most of the period in which data were collected are rather similar. In most regions and for a large part of the seasonal cycle, we find that the spatial standard deviation of climatological monthly salinity contributes to a rather large representativeness error with the sampling of the LOCEAN data-set which exceeds 0.1 on salinity for most months and regions. The notable exception is in the better sampled boxes of the North Atlantic subpolar gyre.

For example, in the central Atlantic north-equatorial (5-12°N) and equatorial (5°N-5°S) domains, the spatial standard deviation of climatological monthly salinity in the in-situ climatology varies between 0.2 (north-equatorial in months 1-4) to 0.4 (north-equatorial, months 6-11) or even 0.5 (equatorial, months 10-11-12-1), and this both for the in situ and the CCI [monthly](#) climatologies. Assuming random spatial sampling in the current LOCEAN dataset, this would contribute to a representativeness error which exceeds 0.1 on salinity for most months.

However, there are other contributions to the standard error on the monthly averages, as a large proportion of the data can originate from a single cruise, and all data are not independent with respect to interannual, higher frequencies than monthly or spatial variability within the domain. These all contribute to the difference of S to the average climatological seasonal cycle. In the instance of the North Equatorial region on Fig. 2 (and other regions too), the LOCEAN S seasonal cycle is more irregular than in the [sea surface salinity](#) climatologies, but the differences are usually less than ~~the 1~~ ^{σ} the standard deviation between all the data. They are however larger than if one assumes that all data are independent random representation of the mean (thus with a standard error $\sigma/\sqrt{n}$, [where \$\sigma\$ been the standard deviation around the mean](#)). The compromise we adopt to estimate the standard error is to use as the number of degrees

of freedom $n/2$. We also propagate that estimate to the other parameters (Wisotopes), assuming that what is found for S is a model for the sampling issues.

Even, with that assumption, we find in regions which are irregularly spatially sampled that the differences in SSS with the two monthly climatologies are larger than the estimated standard error (Fig. B1). Part of that is associated with spatial variability of the poorly sampled mean monthly fields, as seen for the central equatorial Atlantic region 5°S - 5°N (Fig. 2), when comparing (Fig. B1) the CCI climatology averaged spatially over the box (thick black line) with the climatology sampled at the position of the data (thin black line). The thin black line there, but also to a lesser extent for the north equatorial box (5°N - 12°N) is closer to the monthly data average, with differences compatible with the standard error estimates.

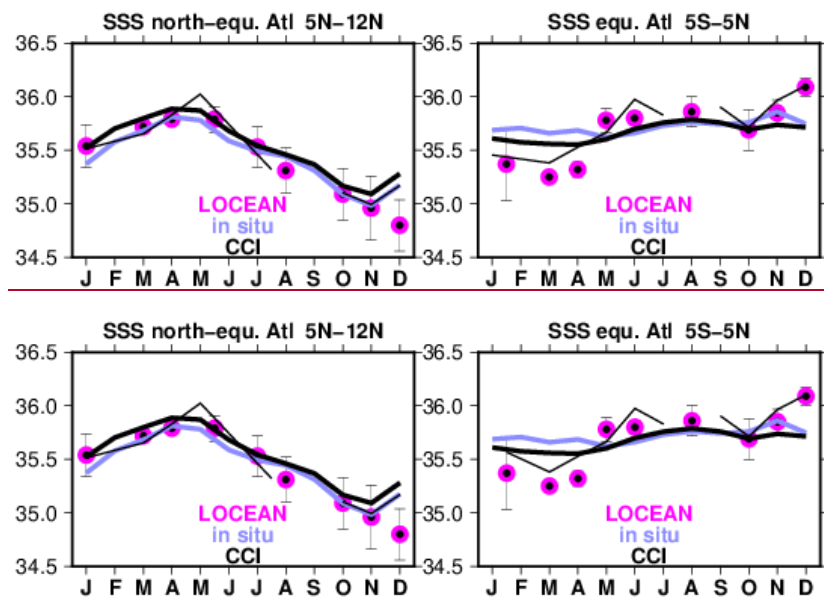


Figure A1-B1: Comparison of average seasonal cycle in the LOCEAN data-set (dots with error bars) with two climatologies (curves), the blue one for 1977-2016 (In situ; Reverdin et al., 2007), and the black ones being for the CCI V4.4 SSS for 2010-2023 (Boutin et al., 2021) (the thick curve for the box averages, and the thin curve for the average at the positions of the sampling). The standard errors plotted on the LOCEAN

1329 monthly averages, are estimated taking $n/2$ as the number of degrees of freedom, where
1330 n is the sample size.

1331 In these two boxes, in months with more than ten data points and extending that
1332 assumption to estimate standard errors for salinity to the other Wisotopes variables, the
1333 estimated standard error is small relative to the amplitude of the sampled seasonal
1334 cycle. For example, for monthly $\delta^{18}\text{O}$ values in the equatorial domain, the estimated
1335 standard error varies between 0.01 and 0.10‰ (smallest in March and largest in the
1336 poorly sampled July to September) The estimated standard error is typically 6 times
1337 larger for monthly $\delta^2\text{H}$ values and 4 times larger for d-excess values than for $\delta^{18}\text{O}$ values.

1338 Furthermore, one can also attempt to correct the sampling bias identified on salinity to
1339 the other variables, by assuming that an average linear relationship between salinity
1340 and Wisotopes holds.

Mis en forme : Interligne : 1,5 ligne

Appendix ~~BC~~: BATS seasonal variability

Water samples were collected nearly monthly from 2015 to 2022 at the BATS hydrographic site in the western subtropical Atlantic (close to (31°50'N, 64°10'W)). They were stored in plastic bottles which were frozen once ~~returning to~~ arriving at the Bermuda ~~oceanographic station~~ Institute of Ocean sciences. They were later unfrozen in Bergen (~~Norway~~) and run in three separate days on a Picarro CRDS L2130. Then, the samples were refrozen, and were unfrozen again shortly before been shipped to LOCEAN, where they were kept unfrozen at 4°C, before analysis (thus, for some, up to nearly two months).

The BATS data have been validated and flagged as usually done at LOCEAN, based on anomalous d-excess and compared to the set analyzed at the Univ. Bergen (referred to as Bergen, later on). Out of 80 samples, some were flagged as likely bad when d-excess was deemed too low (6 for the Bergen set; 15 for the LOCEAN time series) and some as likely good but suspicious if d-excess as too high (2 for Bergen, 5 for LOCEAN). Five out of the six likely bad flagged data in Bergen were also flagged as such in the LOCEAN ~~data~~ set, but not the two Bergen data flagged as ~~suspicious~~ suspicious. The larger number of flagged data at LOCEAN suggests that the plastic bottles in which the samples were stored exchanged water molecules with the outside air, after been unfrozen for shipment to LOCEAN and before analysis. The number of suspicious flags (d-excess deviating by up to +1 ‰ from usual) as well as the non-consistency for that between the two sets is suggestive of some noise in the analyses. In particular, during the LOCEAN daily runs in December 2025/January 2026 there often was a poor reproducibility of a reference sample in the runs. However, based on replicates, we expect that standard errors on the LOCEAN measurements ~~is are~~ usually less than 0.05‰ for $\delta^{18}\text{O}$.

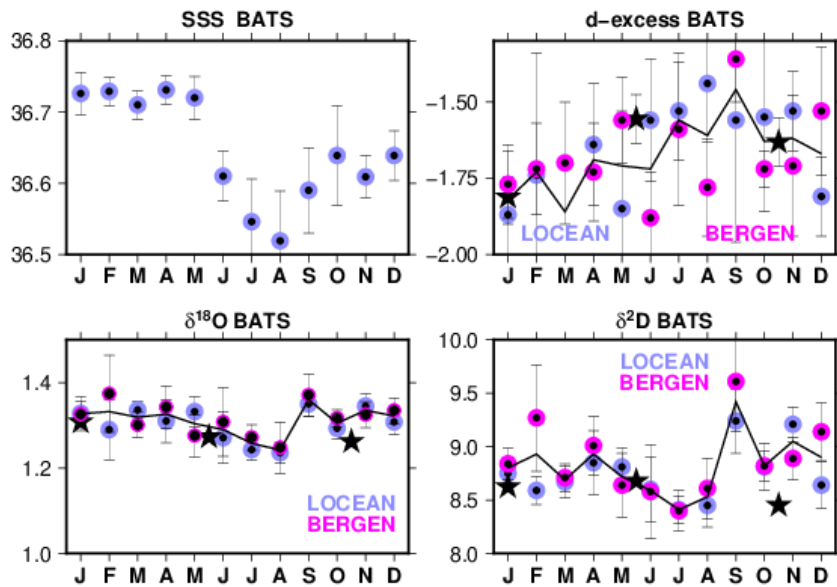
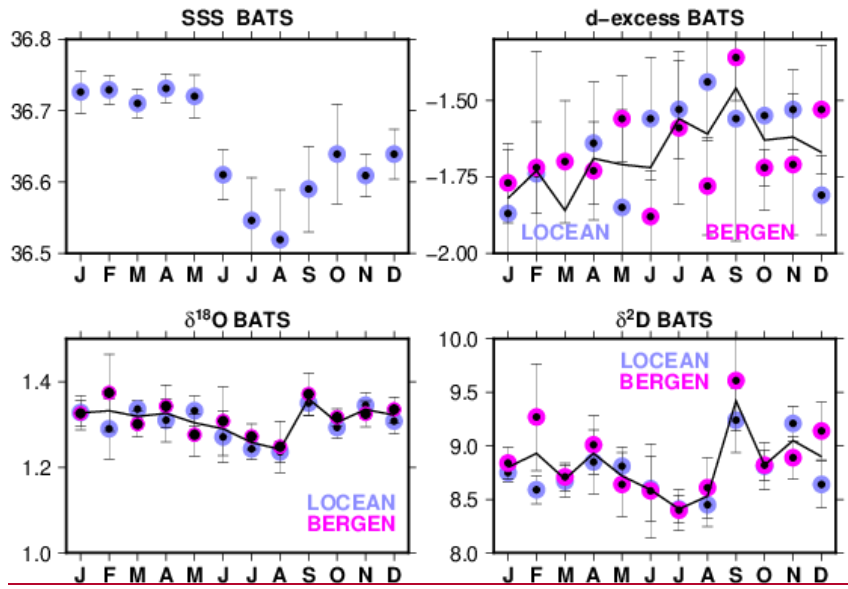


Figure B=C1. Monthly-averaged seasonal cycle at BATS of SSS, $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess (from top to bottom and left to right, in ‰) averages for the 2015 to 2022 period and their 1 σ standard errors depicted (averages and standard errors). The two colours

correspond to the LOCEAN and Bergen analyses of the samples, ~~and whereas~~ the thin line ~~is marks~~ the average of the two sets. The black stars correspond to the averages for regional box 9 (fig. 1) adjusted to the salinity at BATS (top panel left) taking the slopes of the regression to salinity reported in Table .

We also suspect systematic shifts in the Bergen set between three different days of analysis. We thus adjusted the Bergen set to the LOCEAN set by comparing unflagged data separately for each of the three days of analysis in Bergen. After that, both sets are used to estimate an average seasonal cycle (Fig. CB1). Because of the small sample size (averaging on the order of 5-6 ~~samples in for~~ each calendar month), the standard errors are large. The two sets of analyses indicate a rather coherent variability in S, $\delta^{18}\text{O}$ and $\delta^2\text{H}$ with differences within standard error bars (a mostly similar cycle is also found in the Bergen set prior to the adjustment). There are hints of a seasonal cycle with lowest S in June to September. The summer S minimum is a prominent feature in the longer BATS time series, although in the 1990-2000 decade, it shows the minimum occurred in September (Phillips and Joyce, 2007). Fig. CB1 also shows lower $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in June-August. Because of the limited sampling and the standard error bars, the large increase of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in September, with no counterpart in S, may not be a representative feature of an average seasonal cycle. However, it is tempting to associate it with the beginning of the tropical storm (hurricane) season in late summer, and possible individual storm/hurricane signatures. Because of the standard error bars in both analyses, we attribute better confidence in the average of the two sets of analyses (shown as the thin black line in Fig. CB1), with overall a variability in Wisotopes nearly following the spatial relation between S and Wisotopes in this region, except for September (region-box 9 of Table 1).

The standard errors are large on the d-excess monthly averages. However, the average of the two sets suggests a seasonal cycle with lower d-excess in months Dec.-May 12 to 5, and higher in Jul-Nov.-months 7 to 10.

On Fig. C1, we also report as stars the CISE-LOCEAN data averaged in box 9 (to have enough data, months 5 and 6 were grouped, as well as months 10 and 11). These averages have relatively small standard errors on the order of 0.025, 0.160 and 0.070

Mis en forme : Espace Avant : Automatique, Après : Automatique

1401 ‰ for $\delta^{18}\text{O}$, $\delta^2\text{H}$ and d-excess respectively. They seem to fit within the error bars to the
1402 curves of the BATS site, suggesting that the site is representative of a larger area.

Mis en forme : Police :Couleur de police : Noir