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Climate-scale changes of the semidiurnal tide over the North Atlantic coasts from 1846 to 2018

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Abstract.

We investigated the long-term changes of the principal tidal component M_2 over the North Atlantic coasts, from 1846 to 2018. We analysed 9 tide gauges with time series starting no later than 1920. The longest is Brest with 165 years of observations. We carefully processed the data, particularly to remove the 18.6-year nodal modulation. We found that M_2 variations are consistent at all the stations in the North East Atlantic (Newlyn, Brest, Cuxhaven), whereas some discrepancies appear in the North West Atlantic. The changes started long before the XXth century, and are not linear. The trends vary from a station to another; they are overall positive, up to 0.7 mm/yr. Since 1990, the trends switch from positive to negative values. Concerning the possible causes of the observed changes, the similarity between the North Atlantic Oscillation and M_2 variations in the North East Atlantic suggests a possible influence of the large-scale atmospheric circulation on the tide. We discuss a possible underlying mechanism. A different spatial distribution of water heights from one year to another, depending on the low-frequency sea-level pressure patterns, could impact the propagation of the tide in the North Atlantic basin. However, the hypothesis is at present unproven.

1 Introduction

Since the XIXth century, tides are changing due to non-astronomical factors (Haigh et al., 2020). In the North Atlantic, secular variations were observed at individual tide gauge stations, e.g. Brest (Cartwright, 1972; Pouvreau et al., 2006; Pouvreau, 2008), Newlyn (Araújo and Pugh, 2008; Bradshaw et al., 2016), Boston (Talke et al., 2018), but also at regional scale, e.g. Gulf of Maine (Doodson, 1924; Godin, 1995; Ray, 2006; Ray and Talke, 2019), North Atlantic (Müller, 2011), and at quasi-global scale (Woodworth, 2010). Long-term changes in tidal constituents are rather small, but tend to be statistically significant.

The physical causes of these changes are still poorly understood. They may have a local scale origin: changes in the local environment (e.g. harbour development, deepening of channels, dredging, siltation) or changes in the instrumentation (e.g. tide gauge technology, observatory location, instrumental errors). But they may also have a large scale origin, i.e. regional or global. Haigh et al. (2020) reported several possible large-scale mechanisms: (1) tectonics and continental drift, (2) water depth changes due to mean sea level rise or geological processes such as the Earth's surface glacial isostatic adjustment (Müller et al.,

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2011; Pickering et al., 2017; Schindelegger et al., 2018), (3) shoreline position, (4) extent of sea-ice cover (Müller et al., 2014), (5) sea-bed roughness, (6) ocean stratification which may modify the internal tide and change its surface expression (Müller, 2012), (7) non-linear interactions and (8) radiational forcing (Ray, 2009).

This paper has two main objectives. The first is to characterize the secular changes of the M_2 tide over the North Atlantic. We focus on the longest time series, i.e. starting no later than 1920. This approach is complementary to previous studies investigating M_2 changes focusing on smaller spatial scale, e.g. Brest (Pouvreau et al., 2006; Pouvreau, 2008), Gulf of Maine (Ray, 2006; Ray and Talke, 2019), or focusing on smaller temporal scale, i.e. recent decades (Woodworth, 2010; Müller, 2011). The second objective is to discuss a possible climate mechanism that can partly explain the observed changes.

The paper is organised as follows. The first section describes the data: the sea level data (i.e. tide gauges and their processing) and the atmospheric data (i.e. climate indices and sea level pressure data). The following section presents the results (i.e. M_2 variations and trends). We then discuss a possible link between the observed changes and mean sea level rise, as well as climate indices.

2 Data

2.1 Sea level data

2.1.1 Tide gauges selection

The tide gauge data were retrieved from the University of Hawaii Sea Level Center (website accessed April 2020). The dataset consists of 249 stations in the Atlantic Ocean, with hourly sea level observations. The vertical reference level differs from a station to another, which has no impact here, as we focus on tidal components. We apply harmonic analysis on a yearly basis to determine the tidal constituents, thus only a change in the reference level within a year can affect the results.

We selected the stations following three criteria: (1) time series starting before 1920, (2) time series with at least 80 years with data, (3) tidal amplitude significant enough to detect trends, i.e. M_2 amplitude larger than 10 cm. Only 15 stations among the 249 followed the two first criteria (Figure 1). They are all located in the northern hemisphere. On the east side, Stockholm, Gedser, Hornbaek and Marseille were discarded due to a too small M_2 amplitude (i.e. lower than 10 cm). These stations are located in the Baltic Sea (Stockholm, Gedser), in the strait separating the Baltic and the North Sea (Hornbaek), and in the Mediterranean Sea (Marseille). On the west side, Galveston and Cristobal were also discarded due to a too small tidal amplitude (i.e. lower than 10 cm). These stations are located in the Gulf of Mexico (Galveston) and the Caribbean Sea (Cristobal).

Finally, 9 stations followed the three criteria detailed above, and were selected for this study (see stations in bold on Figure 1). Among them, 3 are located on the North East Atlantic coasts (Newlyn, Brest, and Cuxhaven - note that Cuxhaven is located

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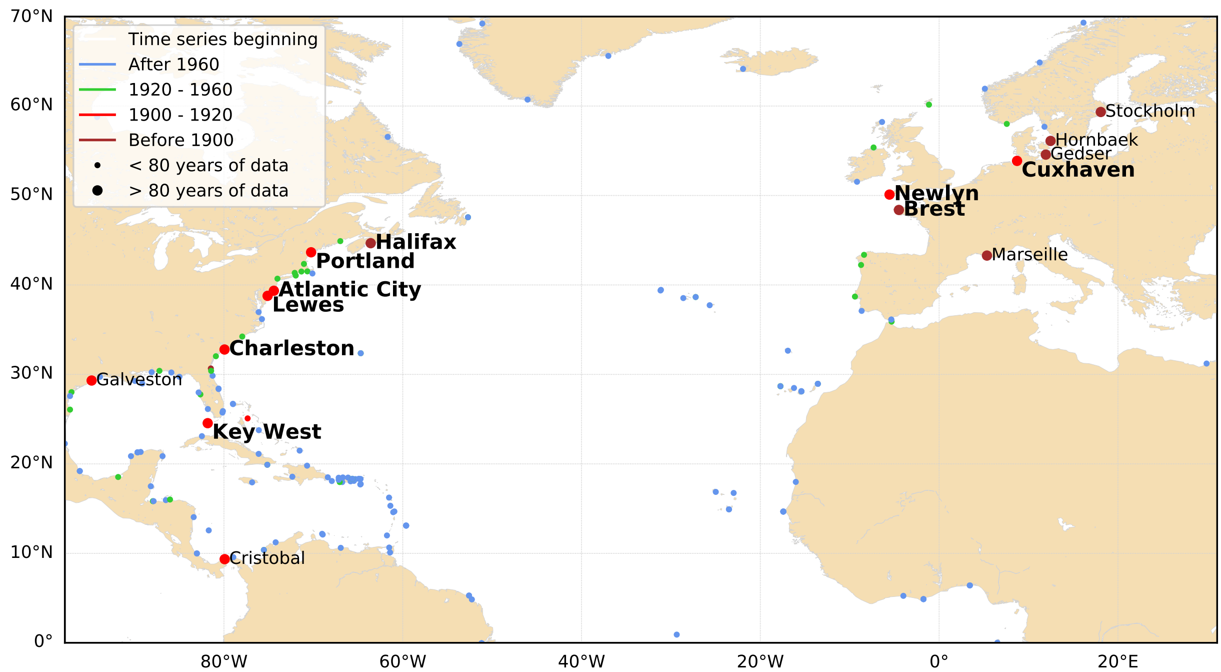

Figure 1. Tide gauges in the North Atlantic. Stations with time series starting before 1920 and longer than 80 years are labelled. Stations selected for this study are in bold.

in the North Sea) and 6 are located on the North West Atlantic coasts (Halifax, Portland, Atlantic City, Lewes, Charleston and Key West).

The main characteristics of the 9 selected stations are synthetised in Table 1. Among them, only Brest and Halifax started in the XIXth century, respectively in 1846 and 1896 (Table 1, column 2). The number of years with data for each station varies between 85 and 165 years, Brest being the longest time series (Table 1, column 3).

2.1.2 Data processing

Harmonic analysis was performed to compute the M_2 amplitude. We used the MAS program (Simon, 2007, 2013), developped by the French Hydrographic Office (SHOM). This program gives results similar to T_Tide harmonic analysis toolbox (Pawlowicz et al., 2002), largely used in the scientific community. For instance, Pouvreau et al. (2006) found non-significant differences on the yearly amplitudes of M_2 at Brest over the period 1846 to 2005 using T_Tide or MAS. Hourly time series were analysed yearly. We processed only years with at least 180 days, considering that six months was long enough to compute correctly M_2 (Pouvreau et al., 2006). This constraint resulted in excluding between 1 and 9 years, depending on the station (Table 1, columns 3 and 4). Note that M_2 is affected by a seasonal variation of a few percent (Huess and Andersen, 2001; Müller et al., 2014) ; keeping years with at least 75% of the data (instead of 50 % here) would allow to avoid this modulation,

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Table 1. Main characteristics of tide gauges selected for this study. Name of the station, timespan, number of years with data, number of years analysed (i.e. with more than 180 days), MSL average over the period 1910-2010, M_2 average amplitude and standard deviation over the period 1910-2010, M_2 nodal modulation, M_2 estimated trends since 1910 and since 1990.

Name	Timespan	Nb of yrs with data	Nb of yrs analysed	$\overline{MSL}$ (cm) [1910-2010]	$\overline{M_2}$ (cm) [1910-2010]	M_2 nod. mod. f_{nod}	M_2 trends since 1910 (mm/yr)	M_2 trends since 1990 (mm/yr)
Newlyn	1915-2016	102	100	313.5 ± 5.6	170.64 ± 0.77	3.3 %	0.15 ± 0.02	-0.28 ± 0.13
Brest	1846-2018	165	160	409.1 ± 4.8	204.54 ± 0.91	3.8 %	0.13 ± 0.02	-0.36 ± 0.12
Cuxhaven	1918-2018	102	101	507.2 ± 7.2	135.05 ± 3.68	1.8 %	0.68 ± 0.10	-0.47 ± 0.41
Halifax	1896-2013	99	96	93.8 ± 8.8	62.83 ± 0.64	3.7 %	-0.14 ± 0.02	0.33 ± 0.16
Portland	1910-2018	109	108	406.7 ± 6.1	135.07 ± 1.82	2.8 %	0.56 ± 0.03	0.73 ± 0.20
Atlantic City	1912-2018	107	104	206.0 ± 12.0	58.48 ± 0.31	3.8 %	0.00 ± 0.01	-0.18 ± 0.07
Lewes	1919-2018	85	76	147.4 ± 8.2	59.92 ± 0.43	3.1 %	-0.06 ± 0.02	-0.33 ± 0.06
Charleston	1901-2018	101	100	164.6 ± 8.7	76.40 ± 1.33	3.0 %	0.32 ± 0.03	-0.02 ± 0.08
Key West	1913-2018	106	105	159.0 ± 7.1	17.50 ± 0.36	2.9 %	0.08 ± 0.01	0.12 ± 0.02

but would lead to exclude more years.

We carefully retrieved the nodal modulation of M_2 amplitude (Simon, 2007, 2013). Here is a short description of the method. The M_2 component is subject to a 18.6-year modulation, when poorly separated from a neighboring component. Indeed, M_2 is very close in terms of frequency to another component (m_2) whose Doodson number differs only from the 5^{th} figure (255 555 and 255 545 for M_2 and m_2, respectively). This 5^{th} figure corresponds to N', the opposite mean longitude of the Moon ascending node - hence the "nodal" term - whose period is 18.6 years. Note that there is also another component close to M_2, whose Doodson number differs only from the 5^{th} figure (255 565), but it is negligible as its amplitude in the tidal potential is only 0.05% of M_2, whereas m_2 amplitude is 3.7 % of M_2 (Simon, 2007, 2013). With one year of hourly data, the two components M_2 and m_2 are not correctly separated with a harmonic analysis (at least 18.6 years are necessary). As a consequence, M_2 amplitude is modulated by m_2. However, we can estimate this modulation, and remove it. The harmonic formulation is expressed as a sum of harmonic components

$$h(t) = \sum_i a_i cos(V_i(t) - \kappa_i) \quad (1)$$

where $h(t)$ is the sea level height at time t, $V_i(t)$ is the astronomical argument (computed from Doodson number) and a_i, κ_i the amplitude and phase shift of each component. Considering that M_2 and m_2 are very close in terms of frequency, we can assume that their phase shift are similar ($\kappa_{M2} \simeq \kappa_{m2}$). As their difference of astronomic arguments is $V_{m2} - V_{M2} = N' + \pi$, the M_2 and m_2 contributions to the total water level may be expressed as

$$h_{M2}(t) + h_{m2}(t) = h_{M2}(t)[1 + f_{nod} cos(N' + \pi)] \tag{2}$$

where f_{nod}, the nodal modulation, is the ratio of the amplitude of m_2 and M_2. As M_2 and m_2 are very close in terms of frequency, f_{nod} is generally considered as close to the ratio of their amplitude in the tidal potential, A_{m2} and A_{M2}

$$f_{nod} = \frac{a_{m2}}{a_{M2}} \simeq \frac{A_{m2}}{A_{M2}} \simeq 0.037. \tag{3}$$

The opposite of the mean longitude of the Moon ascending node is simply expressed as a function of time (p . 116 in Simon (2007), p. 112 in Simon (2013))

$$N' = -N = 234.555 + 1934.1363T + 0.0021T^2 \tag{4}$$

with N' in degrees, and T the time elapsed since 2000/01/01 at 12:00, expressed in Julian centuries (36 525 days).

The tidal program we used (MAS) corrected M_2 applying the usual 3.7% nodal modulation (Eq. (3)). However, this value may vary significantly from a station to another; Ray (2006) reported values ranging from 2.3 % to 3.6 % in the Gulf of Maine. Here, we computed directly f_{nod} from the observed data, proceeding as follows. (1) We added default nodal correction $1 + 0.037cos(N' + \pi)$ to the M_2 variations. (2) We detrended the obtained signal removing the last Intrinsic Mode Function (IMF) of an Empirical Mode Decomposition (EMD) (Huang et al., 1998); note that the EMD is an analysis tool which partitions a series into 'modes' (i.e. IMFs), the last one being the trend of the signal. (3) We fitted a function $a_{m2}cos(N' + \pi)$ on this detrended signal to estimate a_{m2}, N' being expressed as in Eq. (4). (4) We finally computed f_{nod} as the ratio between m_2 and M_2 amplitudes (Eq. (3)). Figure 2 (a) shows an example of estimate of M_2 modulation at Newlyn: the fit leads to a nodal modulation of 3.3 %. Note that this value is consistent with Woodworth (2010) (3.2 %), whereas Woodworth et al. (1991) gave a slightly different value (2.8 %). Figure 2 (b) shows the impact of this value rather than the default one: oscillations of 18.6 years are clearly reduced. Note that in this study, the m_2 amplitude - and then the nodal correction - could have been computed from the full time series harmonic analysis, as records are longer than 18.6 years. However, the method presented here to compute the nodal correction, can be applied even for time series shorter than 18.6 years.

The computed nodal modulations are synthetised in Table 1 (column 7). They vary from 1.8 to 3.8 %. Note that these values are consistent with those obtained by previous authors (Ray, 2006; Müller, 2011; Woodworth, 2010; Ray and Talke, 2019). Only the value at Charleston differs significantly - 3.0 % in our study compared to 3.7% in Müller (2011).

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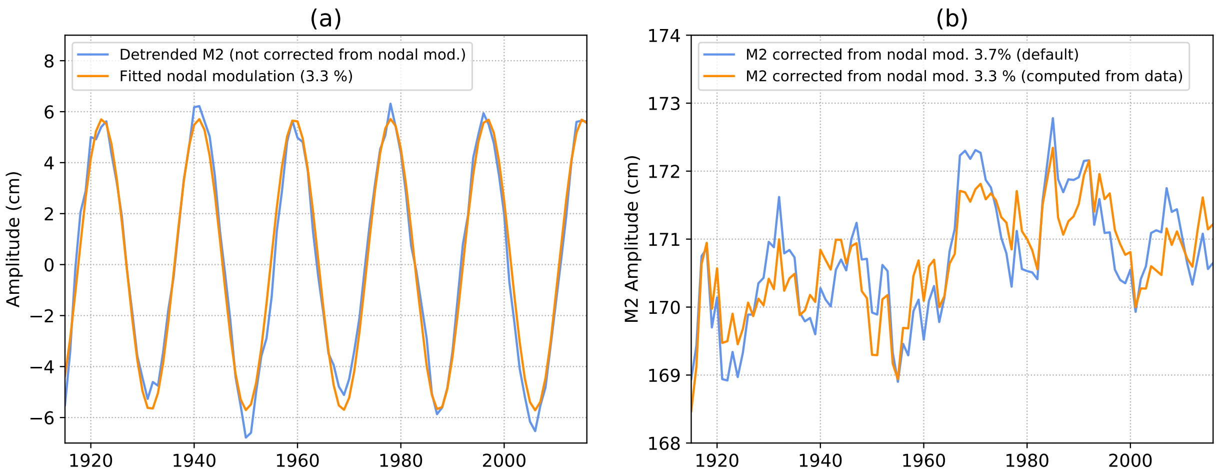

Figure 2. (a) Estimation of M_2 nodal modulation at Newlyn (b) Impact of M_2 nodal modulation correction at Newlyn

At all the stations, we computed the normalized M_2 amplitude, removing the average and dividing by the standard deviation over the period 1910-2010

$$Normalized\ M_2(t) = \frac{M_2(t) - \overline{M_2}_{[1910,2010]}}{\sigma_{M_2[1910,2010]}} \tag{5}$$

the average $\overline{M_2}$ and standard deviation σ_{M_2} over the 1910-2010 period being given in Table 1 (column 6). The idea is to scale the data, in order to compare all the stations together.

2.2 Atmospheric data

2.2.1 Climate indices

We investigated the correlation between secular changes in the tide and climate indices, such as the North Atlantic Oscillation (NAO) or the Arctic Oscillation (AO) - also called Northern Annular Mode (NAM). Climate indices are related to the distribution of atmospheric masses. They are based on the difference of average sea-level pressure between two center of actions (i.e. stations), at large time scale (e.g. monthly, seasonal, annual).

The NAO is the major pattern of weather and climate variability over the Northern Hemisphere (Hurrell, 1995; Hurrell and Deser, 2009). Variations of NAO are essential, as they drive the climate variability over Europe and North America (Hurrell et al., 2003). We used the wintertime (December to March) Hurrell station-based NAO Index (retrieved from https://climatedataguide.ucar.edu/climate-data/hurrell-north-atlantic-oscillation-nao-index-station-based). It is based on the difference of normalized average winter sea-level pressure between Lisbon (Portugal) and Stykkisholmur/Reykjavik (Iceland). The normalization consists of removing the long-term mean (1864–1983) and dividing by the long-term standard deviation.

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The NAO index covers the period 1864-2019, with yearly values.

The Artic Oscillation (AO) is another index which resembles to NAO index. It is defined as the first EOF of northern hemisphere winter sea-level pressure data (Thompson and Wallace, 1998, 2000; Thompson et al., 2000). The AO index is highly correlated with the NAO. We used the wintertime Hurrell AO index (retrieved from https://climatedataguide.ucar.edu/climate-data/hurrell-wintertime-slp-based-northern-annular-mode-nam-index). The AO index covers the period 1899-2019.

To remove the interanual variability and estimate low frequency variations, climate indices were filtered with a 9-year median filter.

2.2.2 Sea level pressure

We explored the gridded seasonal sea-level pressure reconstruction from 1750 to 2002, covering eastern North Atlantic, Europe and the Mediterranean area (Küttel et al. (2009), https://www.ncdc.noaa.gov/data-access/paleoclimatology-data). This 5°X5° gridded dataset is based on ship logs and instrumental pressure series. We computed the mean winter (December to February) sea-level pressure over the period 1850-2002. We averaged from 1850 rather than 1750 to be consistent with tide gauges temporal coverage. We also computed yearly anomalies, i.e. removing the average sea-level pressure.

3 Results

3.1 M_2 variations

For the North East Atlantic, the variations of normalized M_2 amplitude are presented Figure 3 (a). The first result is that the variations between Newlyn, Brest and Cuxhaven are very similar. This suggests that these changes are probably due to large-scale processes, rather than local effects due to changes in the environment (e.g. harbor development, dredging, siltation) or instrumentation errors. The high correlation between Brest/Newlyn and Cuxhaven may be surprising, as Cuxhaven is located in the North Sea (and not in the open Atlantic Ocean), and far away from Brest (around 1300 km from Brest, compared to 200 km between Brest and Newlyn). This indicates that the spatial scale of the processes responsible for these changes is probably at least as large as the North East Atlantic. The second result, is that there is no linear trends in M_2 variations, but rather break or change points, M_2 increasing and then decreasing, depending on the periods considered. Overall, M_2 increases before 1880, then decreases until 1960, increases again until 1980-1990, to finally decrease since 1990; note that the curve is flattening between 1920 and 1940. Pouvreau et al. (2006) yet noticed these variations at Brest and Newlyn, and suggested a long-period oscillation of around 140 years, rather than a steady secular trend. A careful analysis of the harmonic development of tidal potential showed that no tidal component could explain this oscillation. Similarly, no linear combination of tidal harmonic components could explain it (Pouvreau et al., 2006). This suggests that these variations are not due to an astronomical component, but rather linked with changes in the solid Earth-ocean-atmosphere coupling system. Unfortunately, Newlyn and

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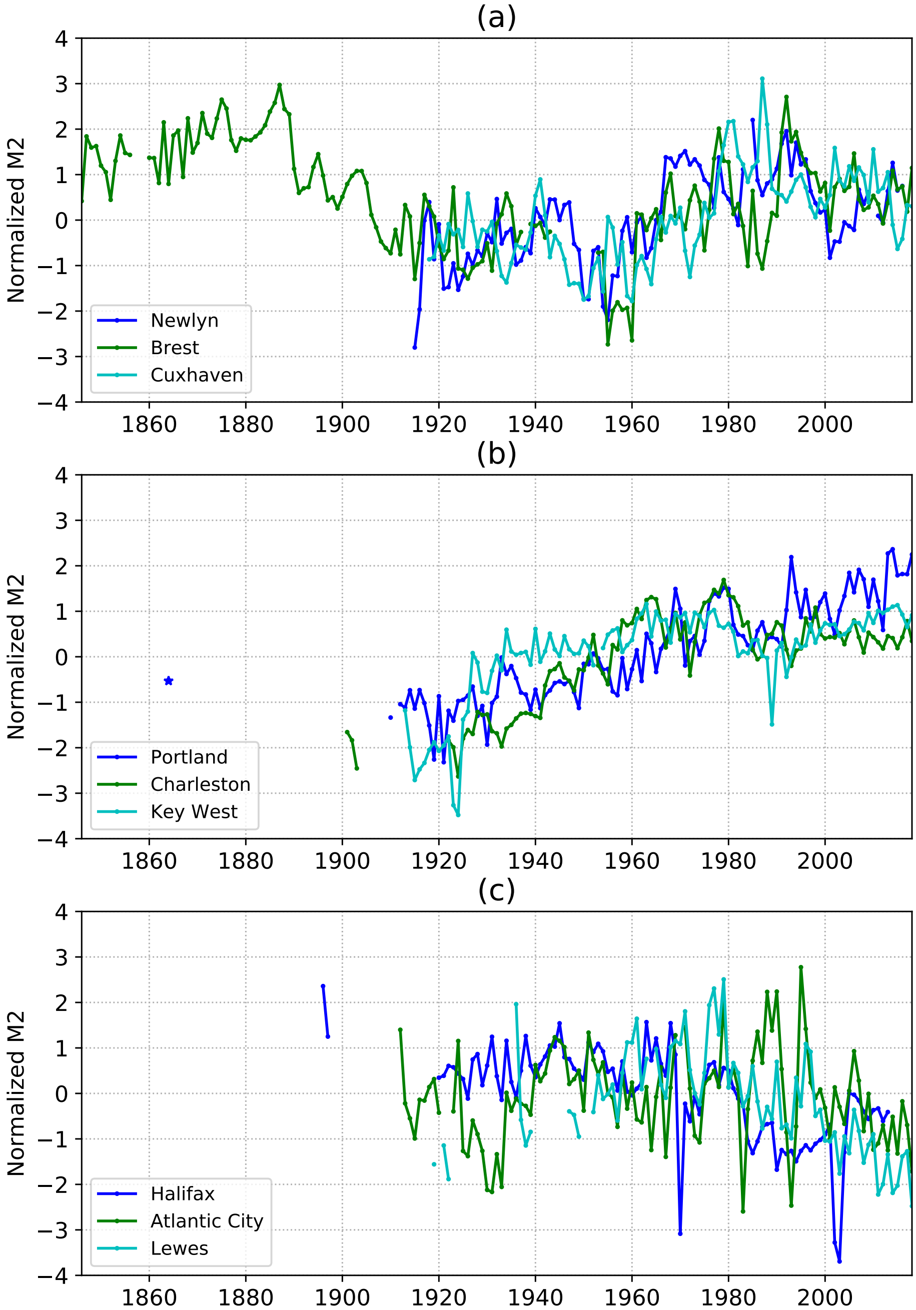

Figure 3. Normalized M2 amplitude (a) in the North East Atlantic (Newlyn, Brest, Cuxhaven) (b) in the North West Atlantic, stations with positive trends (Portland, Charleston, Key West) (c) in the North West Atlantic, stations with negative or no trend (Halifax, Atlantic City, Lewes). The blue star on (b) corresponds to M_2 amplitude at Portland from Ray and Talke (2019), after normalization (Eq. (5)).

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Cuxhaven time series starting only in 1915 and 1918, respectively, do not allow to confirm at large-scale the decrease observed at Brest from 1880 to 1920. This underlines the importance of sea level data archaelogy, for research studies related to long-term changes (Woodworth et al., 2010; Ray and Talke, 2019; Bradshaw et al., 2015, 2020). The third result is that changes in M_2 have not the same order of magnitude at each station, even if trends are similar. Note that Figure 3 represents normalized M_2, i.e. removing the average and dividing by the standard deviation. The order of magnitude of (not normalized) M_2 changes are roughly the same at Brest and Newlyn (standard deviations of 0.9 and 0.8 cm, Table 1, column 6), but more than three times larger at Cuxhaven (standard deviation of 3.7 cm). This suggests that Cuxhaven may be more sensitive to the processes responsible for these changes and/or that the environmental setting of Cuxhaven in a semi-closed basin could introduce some amplification (e.g. resonance effects, propagation in shallow waters).

For the North West Atlantic, the variations of normalized M_2 amplitude are presented on Figure 3 (b) and (c). We split the stations in two groups, in order to facilitate the detection of patterns. The first feature is that M_2 amplitude varies differently in the North West and in the North East Atlantic. The second is that there are discrepancies between stations, even when close to each other (e.g. Atlantic City and Lewes). We split the stations in two groups, each being consistent in terms of trends: one with globally positive trend, the other one with globally negative or no trend.

The first group in the North West Atlantic consists of Portland, Charleston and Key West (Figure 3 (b)). Three outcomes can be highlighted. The first is that M_2 amplitude globally increases since 1900. However, between 1980 and 1990, the three stations slightly decrease and since 1990, only Portland is still increasing significantly. The second outcome is that the rate of increase is very different from a station to another: Portland is increasing 1.4 times faster than Charleston (standard deviations being respectively of 1.82 and 1.33 cm), and 28 times faster than Key West (standard deviation being only of 0.36 cm at Key West). The very slow increase at Key West is due to a small tidal amplitude (i.e. only 17.5 cm of mean amplitude for M_2, see Table 1, column 6). The large increase in Portland may be explained by some amplification in the Gulf of Maine. Ray and Talke (2019) reported that the tides in the gulf are in resonance, with a natural resonance frequency close to the N_2 tide (Garrett, 1972; Godin, 1993). Tides may be then very sensitive to any changes in the environment (e.g. basin configuration - shape, depth - but also external forcing). The third oucome, and probably the most interesting one, is the value of M_2 at Portland in 1864-1865 (134.1 cm), estimated from Ray and Talke (2019), and represented (after normalization) as a blue star on Figure 3 (b). This value is not consistent with the positive linear trend observed at the three stations since 1900, which confirms the hypothesis formulated from Brest analysis: climate-scale variations show some breaks or change points, M_2 increasing and then decreasing, depending on the periods considered.

The second group in the North West Atlantic consists of Halifax, Charleston and Key West (Figure 3 (c)). Two points can be highlighted. The first is that M_2 globally decreases for Halifax and Lewes, particularly since 1980. This trend is less clear for Atlantic City, which is quite noisy and shows no significant trend. The second point is that at Halifax, M_2 values in 1896-1897 are higher than those after 1920. This suggests that the decrease may have started before the XXth century.

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3.2 Estimated trends

We estimated the trends for M_2 amplitude at each station, using linear regression. We computed the trends over two periods: 1910-2018, which corresponds roughly to the whole period of data (except at Brest), and 1990-2018, which corresponds to recent decades. The results are synthetized in Table 1 (columns 8 and 9) and Figures 4 and 5.

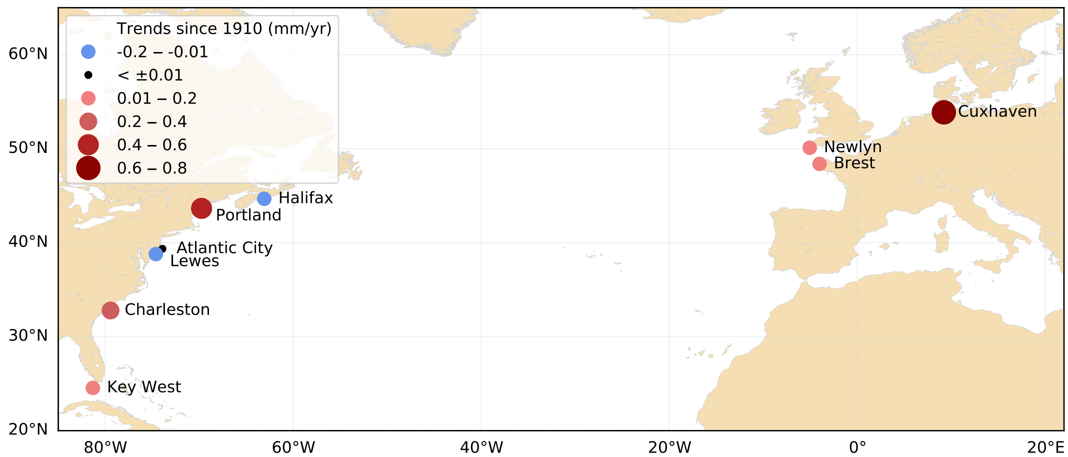

Figure 4. Estimated trends in M_2 amplitude over the period 1910-2018

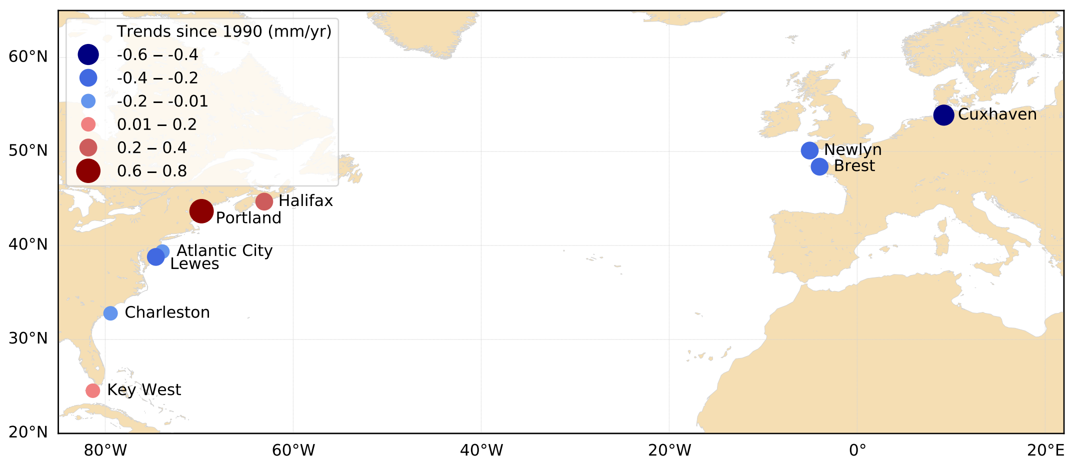

Figure 5. Estimated trends in M_2 amplitude over the period 1990-2018

The trends estimated from 1910 vary significantly from a station to another (Figure 4). They are globally positive (up to 0.7 mm/yr at Cuxhaven), which is consistent with previous findings (Araújo and Pugh, 2008; Ray, 2009; Woodworth, 2010; Müller et al., 2011; Ray and Talke, 2019). They are slightly negative at two stations (Lewes, Halifax), and one station shows no significant trend (Atlantic City). The estimates are statistically consistent with those previously found by different authors (e.g.

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0.15 ± 0.02 mm/yr at Newlyn compared to 0.19 ± 0.03 mm/yr in Araújo and Pugh (2008), 0.56 ± 0.03 mm/yr in Portland, compared to 0.59 ± 0.04 mm/yr in Ray and Talke (2019)). In the North East Atlantic, the trends are consistent, which is not surprising as the stations vary similarly (Figure 3).

The trends estimated since 1990 are quite different from those estimated since 1910 (Figures 4 and 5), with more stations with negative trends: 6 stations (Atlantic City, Lewes, Charleston, Brest, Newlyn, Cuxhaven), instead of 2 stations (Halifax, Lewes). In the North East Atlantic, they switch from positive to negative trends. This underlines (1) some recent changes in the latest decades (Müller, 2011; Ray and Talke, 2019) (2) the difficulty to estimate long-term trends from short records (i.e. less than 30 years), especially if the data are noisy (interannual variability) and the underlying processes non-linear (change points).

Note that the largest trends are observed in semi-closed basins (Cuxhaven in the North Sea, and Portland in the Gulf of Maine). This suggests a possible amplification due to resonance effects.

The trends have to be interpreted very carefully. The M_2 variations are not linear, and may increase or decrease depending on the years; as a consequence, the estimated trends depend strongly on the period considered to estimate it. The interannual variability also plays an important role, and when substantial, trends can vary depending on the computational period.

4 Discussion

4.1 Possible link with mean sea level rise

Mean sea level rise could partly explain M_2 changes, but is not sufficient to explain alone the secular changes in tide (Ray and Talke, 2019). Simulations show that mean sea level rise impact M_2 up to $\pm 10\%$ of the rise (Pickering et al., 2017; Idier et al., 2017). Changes are often of the same sign than mean sea level rise, but sometimes opposite. Figure 6 shows the annual mean sea levels at all the stations, after removing the average over the period 1910-2010 (Table 1, column 5). Mean sea level is rising steadily over all the XXth century, which is not always in line with the changes observed in M_2 amplitude, particularly in the North East Atlantic (Figure 3 (a)). Moreover, global simulations with mean sea level rise suggest that M_2 could increase in the western part of the English Channel (i.e. Brest and Newlyn), and decrease in the southern part of the North Sea (i.e. Cuxhaven) (Pickering et al., 2017). Once again, this is not not supported by our observational results, as M_2 varies the same way at these three stations.

Note that mean sea levels obtained from tide gauges include a solid Earth component as they are referenced to the land. Consequently, if the land is subsiding, mean sea level as observed with a tide gauge will increase (Wöppelmann and Marcos, 2016). Estimates of vertical land motion from SONEL (www.sonel.org, Santamaría-Gómez et al. (2017)) show that the stations considered here are quite stable in the North East Atlantic (i.e. vertical land movements smaller than 0.5 mm/yr), but slightly falling in the North West Atlantic (i.e. trends between -1 and -2 mm/yr), with an exception in the Gulf of Maine, where

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land tends slightly to rise. Note that these trends are computed on relatively short periods (i.e. generally < 15 years), making it difficult to infer robust trends over the last century.

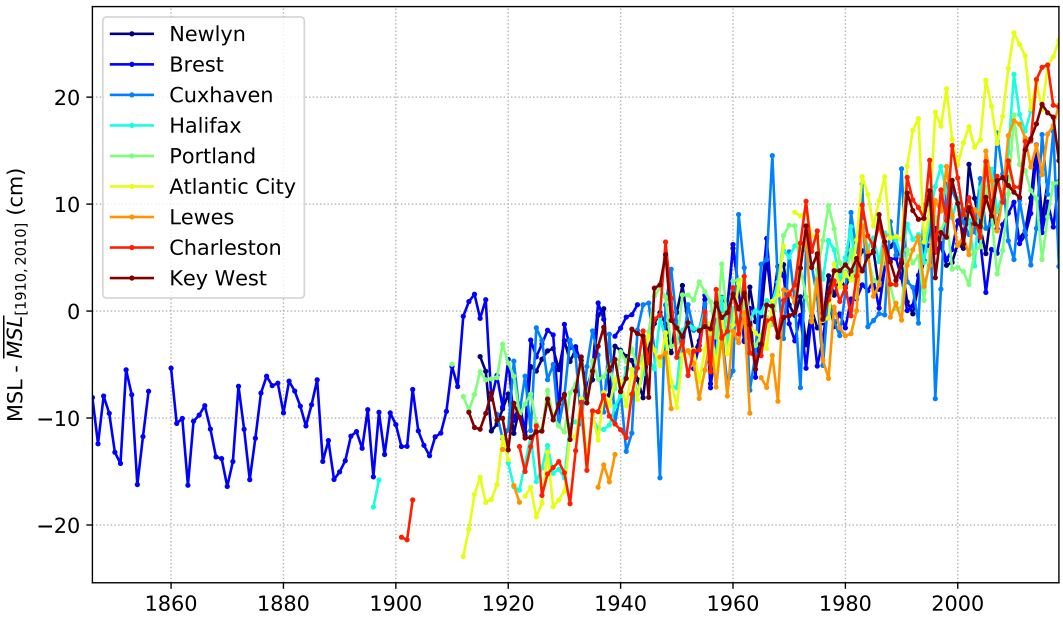

Figure 6. Annual mean sea levels, after removing the average over the period 1910-2010 (see Table 1, column 5)

4.2 Possible link with climates indices

Other processes than mean sea level rise may impact the tide (see section 1). Here, we focus on atmospheric circulation and ocean stratification. Ocean and atmosphere are fully coupled, and air-sea fluxes are responsible for the exchange of momentum and heat at their interface. Two mechanisms can modify the tide. (1) The momentum flux (wind stress) and the gradient of sea level pressure impact directly the water height; significant change in their low frequency variability can impact the tide. Huess and Andersen (2001) showed that simulations better catch the seasonal variability of M_2, when they are forced with a meteorological field. (2) The heat fluxes affects directly the ocean stratification. Any change in the stratification could impact the tide, in two different ways. The first is the internal tide generation which transfers energy from barotropic to baroclinic motion (Kang et al., 2002). The second is that stratification acts on the eddy viscosity profile and bottom drag over continental shelf and then modifies the M_2 surface expression (Müller, 2012; Katavouta et al., 2016). Ray and Talke (2019) suggest a possible role of stratification by long-term warming of the Gulf of Maine waters. To investigate the relationship between these processes and the observed M_2 changes, we used climate indices that are relevant to represent them: NAO/AO indices are representative of the atmospheric circulation, and Atlantic Multidecal Oscillation (AMO) index is representative of the sea

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surface temperature in the North Atlantic.

The NAO index represents the difference of normalized sea level pressure between the Azores high pressure system and the Iceland low pressure one (Hurrell, 1995). It indicates the redistribution of atmospheric masses between the Subtropical Atlantic and the Arctic (Hurrell and Deser, 2009). As the AO is highly correlated with the NAO (Figure 7), in the following, we focus only on the NAO index.

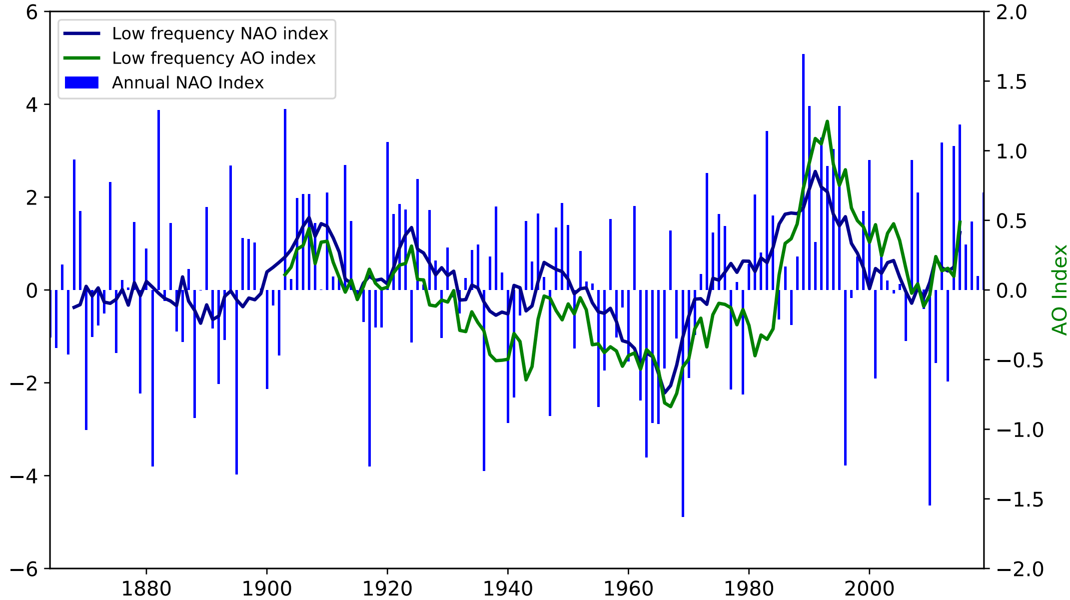

Figure 7. NAO and AO indices. Blue bars correspond to annual values of NAO index. Blue and green lines correspond to low frequency variations of NAO and AO, obtained with a 9-year median filter.

In the North East Atlantic, the similarity between the variations of the low-frequency winter NAO index (Figure 7) and those of M_2 (Figure 3 (a)) suggests a possible impact of large-scale atmospheric circulation on tide. The NAO index varies from positive to negative phases. Filtering the interannual variability, NAO tends globally to decrease between 1910 and 1970, then increase until 1990, and once again decrease. The same way, M_2 amplitude tends to decrease up to 1960, then increase until 1990, and once again decrease. These similar patterns raise a possible connection between NAO and M_2 variations. Yet, this hypothesis is at present unproven. It was tentatively proposed by Müller (2011), without providing any description of the physical mechanism, however. In the following, we develop further this idea.

The underlying mechanism could be the difference of spatial distribution of water heights, depending on the NAO index. Figure 8 (a) shows the average sea-level pressure during the period 1850-2002, derived from a reconstructed sea-level pressure, from ship logs and measurements (Küttel et al., 2009). A positive NAO year (e.g. 1989) corresponds to a situation with a

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stronger gradient pressure than average, between the two pressure systems of Azores and Iceland (Figure 8 (c)). By contrast, a negative NAO year (e.g. 1969) corresponds to a weaker gradient pressure than usually (Figure 8 (b)). This way, from one year to another, the large-scale atmospheric masses are differently distributed, and as a consequence, the water volumes are also differenly distributed in the Northern Atlantic. In a situation of NAO+, the waters are pushed southern, moving from Iceland to the European coasts of France, Spain and Portugal. Figure 9 shows the redistribution of the water volumes, between two years with high and low NAO indices (here 1989 and 1969). Note that this is an extreme situation, as these years have strong positive and negative indices. The impact in terms of water height may vary from -21 cm to 12 cm. This variation of a few tens of cm is probably negligible offshore, but may have some impact on tide propagation along the continental shelves and in shallow waters. It could also shift slightly the amphidromic points. Assuming that these changes have a similar impact (in terms of magnitude) on M_2 as mean sea level changes, that is, $\pm$ 10% according to recent simulations (Pickering et al., 2017; Idier et al., 2017), we find that they can yield changes in M_2 amplitude up to a few centimeters. In other words, their order of magnitude is in agreement with the changes observed in M_2 (Table 1). The assumption is reasonable, but dedicated simulations should be conducted to confirm or discard the water volumes redistribution hypothesis. Finally, note that NAO variability results not only in sea-level pressure change, but also wind stress, air surface temperature and precipitations (Visbeck et al., 2001). Large changes in winds at the scale of the Atlantic could also play a role.

In the North West Atlantic, there is no clear similarity between the NAO index and the variations of M_2. Only the decrease of M_2 since 1990 at Halifax and Atlantic City may reveal a potential link with the NAO, as this index decreases since 1990.

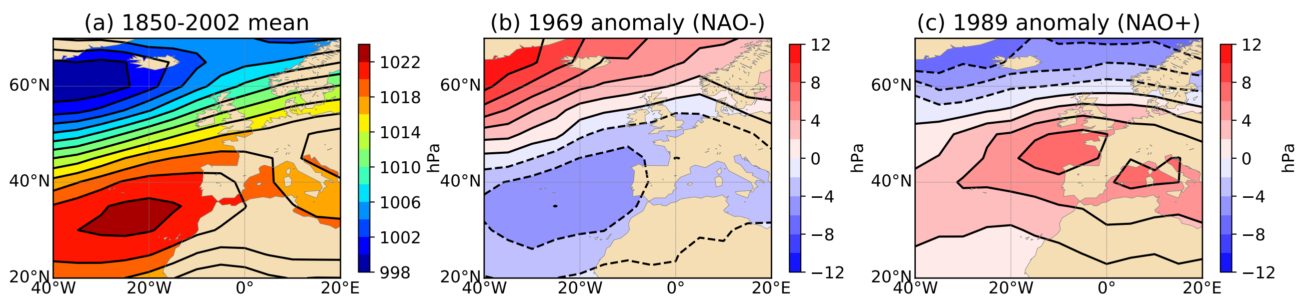

Figure 8. Winter sea-level pressure (a) average over 1850-2002 (b) anomaly in 1969 (NAO-) (c) anomaly in 1989 (NAO+)

Finally, we investigated the link between M_2 variations and AMO. The AMO index is defined as the average sea surface temperature in the North Atlantic, detrended to isolate the natural variability (Enfield et al., 2001). However, we did not find any clear relationship. This index shows an oscillation with a period of around 70 years (Schlesinger and Ramankutty, 1994; Enfield et al., 2001). Since 1856, the lowest indices (i.e. the coldest sea surface temperature periods) were observed in 1900-1920 and 1970-1990, which is not consistent with the observed M_2 variations.

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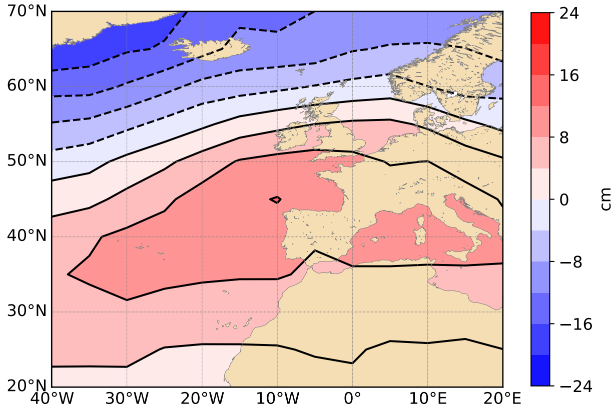

Figure 9. Impact on mean sea level of the difference of winter sea-level pressure between 1989 (NAO+) and 1969 (NAO-)

5 Conclusions

We investigated the long-term changes of the principal tidal component M_2 over the North Atlantic coasts. We analysed 9 tide gauges with time series starting no later than 1920. The longest is Brest with 165 years of data. We carefully processed the data, particularly to remove the 18.6-year nodal modulation.

We found that M_2 variations were consistent at all the stations in the North East Atlantic (Newlyn, Brest, Cuxhaven), whereas some discrepancies appear in the North West Atlantic. The changes started long before the XXth century, and are not linear. The trends vary significantly from a station to another; they are overall positive, up to 0.7 mm/yr, or slightly negative. Since 1990, in many stations, the trends switch from positive to negative values. The significant difference between the trends since 1910 and 1990 calls for caution when interpreting trends based on short records, i.e. less than 30 years, especially if the data are noisy (interannual variability) and the underlying processes non-linear (change points).

Concerning the causes of the observed changes, the mean sea level rise is not sufficient to explain alone the variations. The similarity between the North Atlantic Oscillation and M_2 variations in the North East Atlantic suggests a possible influence of the large-scale atmospheric circulation on the tide. The underlying mechanism would be a different spatial distribution of water heights from one year to another, depending on the low-frequency sea-level pressure patterns, and impacting the propagation of the tide in the North Atlantic basin. In the future, dedicated modelling studies should be undertaken to confirm or discard this hypothesis.

In this study, we focused only on M_2 amplitude. A similar analysis on the phase would draw a more complete picture of the M_2 variations (Müller, 2011; Woodworth, 2010; Ray and Talke, 2019). Other constituents are also affected. Results show that S_2 amplitude decreases at all the stations located in the North West Atlantic, and in contrast, tend to increase in the North East

Atlantic (not shown). The large-scale decrease of S_2 observed in the North West Atlantic is consistent with previous studies, e.g. Ray (2006) in the Gulf of Maine. Further investigations should be definitely conducted to extend this work to more constituents.

One of the major finding of this work is that the changes started long before the XXth century. This conclusion would not have been possible without the huge work of data rescue undertaken over the past decades (e.g. Pouvreau et al., 2006; Pouvreau, 2008; Bradshaw et al., 2016). This underlines the great importance of sea level data archaeology, which allows to extend and improve historical datasets (Woodworth et al., 2010; Bradshaw et al., 2015, 2020; Ray and Talke, 2019; Haigh et al., 2020). This is essential for studies related to climate change.

Finally, we should mention several limitations and perspectives in this study. (1) We considered years with at least 50% of data. However, M_2 is affected by a seasonal variation of a few percent (Müller et al., 2014). Keeping years with at least 75% of the data would allow to avoid this modulation - but would lead to exclude more years. (2) We processed the time series downloaded from the database, considering they were quality controlled. A deep analysis of the data quality before processing would probably be valuable. (3) We did not investigate the history of each station. There are probably some local changes (e.g. environment or instrumentation) that may explain a part of the variability of M_2 amplitude, and some discrepancies between stations. (4) The tide gauges are located on the coast, and mainly in harbours. They are affected at the same time by local and regional/global scale changes, that are difficult to separate. Moreover, they may be not representative of changes offshore. A similar study based on satellite altimetry data would probably be of great interest, even if temporal scale for satellite data is still rather short (i.e. < 30 years) compared to climate-scale processes.

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and NOAA Paleoclimatology Program (retrieved from https://www.ncdc.noaa.gov/data-access/paleoclimatology-data accessed March 2020). The estimates of vertical land motion were retrieved from SONEL (www.sonel.org, accessed April 2020). The harmonic analysis program MAS was provided by the French Hydrographic Office (SHOM). The authors warmly thank P. Woodworth for his helpful comments.

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