Modulation of Tropical Cyclone Tracks and Rainfall by the North Atlantic Oscillation

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Abstract Understanding and quantifying the large-scale environmental control of spatio-temporal tropical cyclone (TC) variability beyond a few weeks remains a challenge with significant implications for societal impacts. This study focuses on the relationship between the North Atlantic Oscillation (NAO) and year-to-year changes in TC activity and rainfall using observational data and reanalysis products. We use Poisson regression models to show that low-frequency NAO (LF-NAO) variability is associated with a distinct pattern of TC activity, which extends across the western North Atlantic, the Caribbean Basin, and the Gulf of Mexico. The negative LF-NAO phase is associated with enhanced TC activity: an interquartile range decrease in the NAO corresponds to a 30%–40% increase in TC track density. While the NAO is known to affect the weather regimes of the mid-latitudes, we show that its low-frequency component has a strong correlation to the large-scale environment across the Main Development Region of TCs. The negative LF-NAO phase is associated with two favorable environmental conditions for TCs: significantly higher sea-surface temperature and weaker deep-tropospheric wind shear. The LF-NAO relationship to TC activity is strongest during El Niño Southern Oscillation positive or neutral conditions and during the negative Atlantic Multidecadal Oscillation phase, and it can also be detected in the basin-scale variations of TC rainfall. By developing annual TC rainfall composites from satellite data and reanalysis products, we show that TC rainfall is strongly enhanced in the Caribbean and in the Gulf of Mexico during the negative LF-NAO phase.

Plain Language Summary Better understanding high-impact weather events like tropical cyclones (TCs), in particular their variability in a changing climate and rising sealevel, is key to improving our preparedness and minimizing their impact. This study advances our knowledge of what controls TC numbers, locations, and heavy rainfall. In this study, we investigate how changes in atmospheric and oceanic conditions associated with the North Atlantic Oscillation (NAO) modulate North Atlantic TC tracks and rainfall. We show that year-to-year changes in the number of TCs across the Western Tropical Atlantic, the Caribbean basin and the Gulf of Mexico reflect slow-evolving variations in the NAO. By comparing opposite phases of the NAO we show that the number of TCs increases by 30%–40% when the NAO is in the negative phase. Observations and reanalysis products show that the negative phase of the NAO is associated with two significant changes in the environmental conditions that affect TC activity across the tropical North Atlantic: (a) above-average sea-surface temperature and (b) below-average vertical wind shear, which are favorable for TC formation and intensification. The influence of the NAO is strongest during otherwise less favorable large-scale conditions (El Niño Southern Oscillation positive-neutral and negative Atlantic Multidecadal Oscillation). Both satellite- and reanalysis-derived rainfall data show that an NAO-induced increase in TC activity and a shift of TC tracks from the western Atlantic to the Caribbean and the Gulf of Mexico have direct impacts on TC rainfall.

1. Introduction

Tropical cyclones (TCs) are extremely impactful weather phenomena, responsible for billion-dollar losses and more than 2,500 deaths in the United States from 1963 to 2012 (Rappaport, 2014). Annual TC activity varies significantly within the North Atlantic basin (Figure 1): on average, the highest number of TCs is found just offshore the southeastern US coast, while a local minimum is instead present in the southern Caribbean (Shieh & Colucci, 2010). North Atlantic TC activity also varies remarkably from year to year, especially in areas such as the tropical North Atlantic, the Northern Caribbean/Gulf of Mexico and along the East Coast (Figure 1b). Multiple modes of climate variability are known to affect the number and distribution of North Atlantic TCs on seasonal to multi-decadal timescales. The eastward propagation of the Walker circulation that characterizes El Niño events induces a significant increase of the tropospheric wind shear over the Caribbean, resulting in lower-than-average TC activity and reduced US landfalls (Boudreault et al., 2017; Camargo et al., 2007; Elsner, 2003; Goldenberg...
TC activity is also influenced by local and remote SST variability: above-average SST in the Main Development Region (MDR) supports enhanced TC activity, whereas below-average SST can significantly hamper the formation and growth of TCs. The long-term SST variability associated with the Atlantic Multidecadal Oscillation (AMO) has been linked to changes in North Atlantic TC activity, with its positive phase featuring enhanced basin-wide TC activity (Caron et al., 2015; Goldenberg et al., 2001; Jagger & Elsner, 2006). Similarly, the Atlantic Meridional Mode (AMM, Kossin & Vimont, 2007; Patricola et al., 2014; Vimont & Kossin, 2007) is positively correlated with TC activity. Due to its weakly stable nature, the AMM requires an external forcing to be excited, such as the AMO (Vimont & Kossin, 2007) and the North Atlantic Oscillation (NAO) (Chiang & Vimont, 2004; Czaja & Frankignoul, 2002; Penland & Hartten, 2014). Neither the AMO nor the AMM, however, are powerful predictors of TC landfalls along the U.S. coast (Boudreault et al., 2017). Finally, variability in the Sahel rainfall has been linked to changes in the number of Atlantic TCs (Boudreault et al., 2017; Goldenberg & Shapiro, 1996; Kozar et al., 2012), although the correlation exhibits a strong non-stationary behavior that limits its predictive power (Fink et al., 2010).

The NAO is a leading mode of climate variability across the North Atlantic. It is an oscillation in the atmospheric mass between the Arctic region and the subtropical Atlantic (J. W. Hurrell et al., 2001) that dominates the atmospheric variability in the North Atlantic and has direct influences on the weather regimes over Western Europe (van Loon & Rogers, 1978; Wallace & Gutzler, 1981). During the positive phase of the NAO, an anomalously deep Icelandic low combines with an anomalously strong subtropical high to produce enhanced westerlies over the North Atlantic and a poleward shift of the extratropical storm track. The reverse conditions characterize the negative phase, resulting in weaker-than-usual westerlies at lower latitudes. The NAO is typically measured as the sea-level pressure (SLP) difference between Iceland and the Azores or as the time series of the leading empirical orthogonal function (EOF) of SLP anomaly over the North Atlantic (J. W. Hurrell, 1995). While the NAO signal is strongest during the winter months, it is detectable over the entire year. The NAO is a source of variability across multiple time scales, from inter-annual to multi-decadal (J. W. Hurrell, 1995; J. W. Hurrell & Van Loon, 1997). The monthly NAO index power spectrum in Figure 2 shows a distinct annual cycle as well as a broad peak at lower frequencies. As shown by Woollings et al. (2015), NAO variability on different timescales is associated with distinct influences on the Western European climate. Similarly, high- and low-frequency NAO variability is associated with distinct SST patterns over the North Atlantic basin (Hoerling et al., 2001). Subtropical SSTs play an important role in the coupled atmosphere-ocean dynamics of the NAO: the summertime “horseshoe” SST pattern has been shown to promote an NAO-like atmospheric response in the following winter and thus act as a source of low-frequency variability or persistence (Cassou et al., 2004; Czaja & Frankignoul, 1999, 2002; Han et al., 2016; Kushnir et al., 2006). The subtropical SST anomaly associated with the NAO does not only have remote effects in the mid-latitudes but also locally affects the upper-tropospheric wind by inducing changes in tropical deep convection and its associated heating (Cassou et al., 2004; Sutton et al., 2000).

A number of studies have investigated the influence of the NAO on North Atlantic tropical cyclones (e.g., Aryal et al., 2018; Boudreault et al., 2017; Caron et al., 2015; Elsner, 2003; Elsner & Kocher, 2000; Kozar et al., 2012; Villarini et al., 2012; L. Zhang et al., 2022). Elsner and Kocher (2000) find evidence across a wide range of timescales that the landfall of major hurricanes along the US coastlines south of the Carolinas is favored during the negative NAO phase. In a following study, Elsner (2003) documents how zonally-oriented hurricanes are...
more likely to be observed during the negative NAO, hence increasing the chance of landfall along the southern US coast. Conversely, Kossin et al. (2010) and L. Xie et al. (2005) show that more hurricanes recurve near the East Coast during the negative NAO phase. Mei et al. (2014) found a leading mode of TC track density variability within the North Atlantic basin to be linked to the NAO, while the NAO index is shown to be a useful predictor of total North Atlantic tropical cyclones (Kozar et al., 2012) and hurricane (Jagger & Elsner, 2006) counts. Recent studies have focused on the persistent influence of winter and spring NAO on the atmosphere and ocean during the North Atlantic Hurricane season. The springtime NAO is correlated to the fraction of North Atlantic landfalling hurricanes and the occurrence of TC-related extreme rainfall events along the East Coast (Aryal et al., 2018; Caron et al., 2015; Villarini et al., 2012).

The influence of the NAO is often explained in terms of variations in the steering flow through changes in the strength and position of the subtropical high pressure (e.g., Elsner, 2003; Kossin et al., 2010). Other studies point instead to the connection between the NAO and the large-scale environment across the tropical North Atlantic (Jones et al., 2022; L. Zhang et al., 2022). A positive spring NAO is associated with negative SST anomalies across the tropical North Atlantic, which in turn support an anticyclonic circulation, with low relative humidity and enhanced vertical wind shear, thus suppressing TC genesis in the following season (L. Zhang et al., 2022). Anticyclonic wave-breaking events also influence TC activity in the North Atlantic through changes in large-scale environment during the hurricane season and forcing quasi-stationary NAO regimes (Jones et al., 2022; Wang et al., 2020; G. Zhang & Wang, 2019). The mechanisms by which the NAO influences TC activity are not yet fully understood (Boudreault et al., 2017) and existing literature provides inconsistent conclusions on the ultimate NAO effect on Atlantic TCs (G. Zhang & Wang, 2019). Different studies, in fact, question whether the NAO does affect TC activity at all: Colbert and Soden (2012) do not find an effect of the NAO on the tracks of the storms that form within the MDR and argue the characteristic track of straight-moving hurricanes is primarily due to their tendency to form in the south-western portion of the MDR. Boudreault et al. (2017) advise against including the NAO in the pool of predictors of landfalling hurricanes as it lacks predictive capability.

This study focuses on the relationship between the NAO and North Atlantic TC activity and its sub-basin variability. Previous studies approach this question either using storm track clustering methods (e.g., Boudreault et al., 2017; Elsner, 2003; Kossin et al., 2010; Kozar et al., 2012) or EOF analysis (e.g., L. Xie et al., 2005; Mei et al., 2014), conversely we statistically model annual TC track and TC genesis densities across the North Atlantic.
basin using Poisson regressions. Most existing studies focus on the spring or summertime NAO, our study investigates instead the distinction between high- and low-frequency NAO variability and emphasizes the spatial characteristics of their connection to TC activity. We compute a low-passed NAO index and compare the results with those obtained using the unfiltered NAO index. We correlate low-frequency NAO variability to changes in the large-scale environment across the tropical North Atlantic that can affect TC formation and growth. Since TC rainfall has significant implications for the hydro-climatology and the occurrence of extreme events over the southern and eastern United States (e.g., Aryal et al., 2018; Khouakhi et al., 2017; Mazza & Chen, 2021), we examine whether the NAO modulation of TC activity is already detectable in the distribution of North Atlantic TC rainfall using both reanalysis and satellite-derived precipitation estimates.

2. Data and Methods

2.1. TC Track and Genesis Density

The TC track data is obtained from the International Best Track Archive for Climate Stewardship (IBTrACS, Knapp et al., 2010). This study considers all TCs present in the IBTrACS that formed in the North Atlantic basin over the period 1900–2019 (Figure 1). The TC genesis time, latitude and longitude are taken to be those of the first data point for each storm. Given the emphasis on TC rainfall, the entire life of the storms is considered, including their post-tropical phase. Before the analysis, storm track data is linearly interpolated from 3-hourly to hourly frequency. IBTrACS storm position information is used to compute the monthly TC track density and TC genesis density on a 0.5°-resolution rectangular grid extending from 120°W to 0°E and 0°N to 50°N. TC track and genesis densities are calculated as the number of TCs that pass or form within 500 km of any grid point, respectively. The 500-km search radius is applied to reduce the innate high spatial variability in TC densities, similar to what previous studies have done (e.g., L. Zhang et al., 2022). The annual TC track and genesis densities are obtained by summing the monthly density fields over the North Atlantic hurricane season (June–November).

2.2. Climate Indices

Statistical modeling of TC counts in the North Atlantic often employs a set of covariates that represent modes of climate variability known to affect TC activity. The Niño 3.4 index is the preferred index of El Niño Southern Oscillation (ENSO) variability (Boudreault et al., 2017; Kozar et al., 2012; Mann et al., 2007). Tropical Atlantic variability is accounted for by considering the AMO (Boudreault et al., 2017) and the AMM (Colbert & Soden, 2012; Kossin et al., 2010). The NAO is also often included in the pool of predictors (Boudreault et al., 2017; Caron et al., 2015; Elsner, 2003; Kozar et al., 2012; Mann et al., 2007; Villarini et al., 2012), along with the Sahel rainfall anomaly (Boudreault et al., 2017; Kozar et al., 2012) and the MDR upper-tropospheric temperature (Boudreault et al., 2017).

This study aims to investigate the spatial variability in TC activity associated with the NAO. To do so, we select a set of climate indices such as to maximize the period covered and account for the climate variability known to be correlated with North Atlantic TC activity. For the analysis we employ three indices over the period 1900–2019: the Niño3.4 index, the unsmoothed AMO and the principal component (PC)-based NAO. The NAO power spectrum reveals that significant variability is present at different timescales: annual and at lower frequencies (Figure 2). Existing literature indicates that low-frequency NAO variability is linked to tropical and subtropical variability in SST and wind shear (Cassou et al., 2004; Czaja & Frankignoul, 1999, 2002; Hoerling et al., 2001; Jones et al., 2022; L. Zhang et al., 2022), as such we separately consider two NAO indices: (a) the monthly PC-based NAO index (Figure 3a), and (b) a low-frequency NAO (LF-NAO) index (Figure 3b). The LF-NAO index is computed by applying a second-order Butterworth filter to the monthly, PC-based NAO index. Before filtering, the NAO index is trended. The cut-off frequency in the filter is such that all the oscillations with periods shorter than 2.5 years are removed. Before the Poisson regression, all indices are averaged over the June–November period.

2.3. Poisson Regression Model

Poisson regression is the statistical tool of choice when modeling count data such as the basin-wide number of tropical cyclones (e.g., Boudreault et al., 2017; Caron et al., 2015; Elsner, 2003; Jagger & Elsner, 2006; Tippett et al., 2011; Villarini et al., 2012). In a Poisson regression model, the regression equation typically takes the
The logarithm acts as the link function between the expected value of the dependent variable \( E[Y] \) and a linear combination of \( n \) covariates \( x_1, x_2, \ldots, x_n \). The \( \beta \) coefficients that multiply each predictor express the logarithm of the expected change in TC count for a unit increase in the predictor.

\[
\log(E[Y|\mathbf{x}]) = \beta_0 + \beta_1 x_1 + \beta_2 x_2 + \cdots + \beta_n x_n + \epsilon_i
\]

We construct two separate Poisson regression models: in \( P_1 \), the TC track density is the dependent variable (Equation 2) and in \( P_2 \) the TC genesis density is the dependent variable (Equation 3). Both \( P_1 \) and \( P_2 \) employ the same set of covariates, namely the Niño 3.4, the NAO and the AMO indices (Equations 2 and 3).

\[
P_1 : \log(E[TC_{\text{den}}|\mathbf{x}]) = \beta_0 + \beta_1 \text{NAO} + \beta_2 \text{NINO3.4} + \beta_3 \text{AMO} + \epsilon_i
\]

\[
P_2 : \log(E[TC_{\text{gen}}|\mathbf{x}]) = \beta_0 + \beta_1 \text{NAO} + \beta_2 \text{NINO3.4} + \beta_3 \text{AMO} + \epsilon_i
\]

The statistical significance of the regression coefficients is calculated using a two-tailed \( t \)-test at the 95% confidence level, using the effective number of degrees of freedom (\( N^* \)) calculated at each grid point using Equation 4 following Bretherton et al. (1999), where \( N \) is the sample size (\( n = 120 \)) and \( r_1, r_2 \) are the lag-1 autocorrelation of the NAO and the TC track (or genesis) density time series.

\[
N^* = \frac{1 - r_1(\delta t)r_2(\delta t)}{1 + r_1(\delta t)r_2(\delta t)}
\]

For both the filtered and unfiltered NAO indices, the strength of the NAO-TC relationship is quantified by calculating the change in the expected number of TC (and TC genesis events) (\( Y \)) associated with an interquartile change of the NAO (i.e., from the 75th to the 25th percentile) while keeping all the other predictors constant at their mean values (Equation 5).

\[
\text{NAO}_{\text{effect}} = \frac{E[Y|\text{NAO}_{75}] - E[Y|\text{NAO}_{25}]}{E[Y|\text{NAO}_{75}]}
\]

To examine how the NAO relationship to North Atlantic TC activity is affected by ENSO and AMO, we focus on the Caribbean/Gulf of Mexico region (95°–75°W/10°–30°N). We compute the annual TC track density averaged in the region and stratify it by LF-NAO, ENSO and AMO phases. The ENSO positive, neutral and negative phases...
are defined when the average Niño 3.4 is greater than 0.4, between −0.4 and +0.4, and less than −0.4 respectively. AMO and LF-NAO variability is instead partitioned into positive and negative phases. For each AMO and ENSO phase, we compare the TC track density in the positive and negative NAO phases and test the difference using a Mann-Whitney U test under the null hypothesis that the samples are taken from the same distribution.

2.4. Large-Scale Environment

The physical mechanisms connecting the NAO to the North Atlantic TC activity are investigated by computing NAO seasonal (June–November) standardized anomalies of (a) SST from the merged Hadley-OI sea-surface temperature (SST; J. W. Hurrell et al., 2008) from 1900 to 2019, (b) deep-layer wind shear (200–850 hPa) from the ERA5 data set (Hersbach et al., 2020) from 1950 to 2019. For both variables, NAO-relative composites are constructed for the top (NAO+) and bottom (NAO−) terciles of the NAO index distribution.

The relationship between the NAO, SST and wind shear anomalies is isolated by constructing a multiple linear regression including the same set of covariates used for the Poisson models. We regress the standardized monthly SST or wind shear anomaly (\(Y_i\)) onto the NAO indices (Equation 6) and assess the significance of the regression coefficients with a two-sided Student's t-test under the null hypothesis that the coefficient is zero.

\[
y_i = \beta_0 + \beta_1 \text{NAO}_i + \beta_2 \text{NINO3}.4_i + \beta_3 \text{AMO}_i + \epsilon_i
\]  

2.5. TC Rainfall

TC rainfall is calculated from the Climate Prediction Center morphing method (CMORPH) data set and the ERA5 reanalysis over the periods 1998–2019 and 1979–2019 respectively. CMORPH is a half-hourly rainfall estimate product obtained from blending passive microwave precipitation estimates and IR-derived motion vectors (Joyce et al., 2004) available at half-hourly intervals on an 8-km global grid. The ERA5 is a state-of-the-art, high-resolution reanalysis product available at hourly frequency at a 31-km resolution from 1979 to the present (Hersbach et al., 2020). TC rainfall is calculated as the hourly accumulated rainfall within 500 km of the TC center. The chosen search radius is consistent with several prior studies on TC rainfall (e.g., Chen et al., 2006; Jiang & Zipser, 2010; Larson et al., 2005; Lonfat et al., 2004; Mazza & Chen, 2023; Prat & Nelson, 2016; Scoccimarro et al., 2014; Villarini et al., 2014). For CMORPH, the hourly estimates are obtained by averaging the 30-min rain rate estimates centered at the hour. NAO-relative composites are constructed for both CMORP and ERA5 by accumulating TC rainfall over the North Atlantic hurricane season (June–November).

3. NAO Influence on North Atlantic TC Activity

In this section, we discuss the relationship between the NAO and North Atlantic TC genesis and track density.

3.1. Modulation of TC Track Density

We start by considering TC track density. The regression coefficients for the LF-NAO and the unfiltered NAO are shown in Figures 4a and 5a, respectively. As shown in Figure 4a, the LF-NAO index captures a distinct pattern of North Atlantic TC activity variability. Large, negative coefficients extend from the western sector of the tropical North Atlantic into the Caribbean basin and the central Gulf of Mexico. Another area of negative coefficients originates in the MDR and tracks northwest of the Caribbean and further east into the open ocean. This secondary signal is likely associated with storms that recurve away from the East Coast. Although distinct, these areas of statistically significant correlation between TC track density and the LF-NAO appear to be part of a basin-wide pattern of variability. The sign of the regression coefficient indicates that, as the LF-NAO decreases, the number of TCs passing through these regions increases. These results are in stark contrast with those obtained for the unfiltered NAO index, which captures TC activity variability only in very limited sectors of the basin (Figure 5a).

To quantify the strength of the relationship between the LF-NAO and TC track density, we compute the expected change in TC track density if the LF-NAO decreases from a strongly positive regime (75th percentile) to a strongly negative one (25th percentile), while keeping all the other predictors constant at their mean values (Equation 5). Figure 4b shows that in a typical negative LF-NAO regime, the number of TCs in the Caribbean
and the Gulf of Mexico increases by 30%–40% compared to a positive one. We can readily contrast this result with that obtained for the unfiltered NAO, which is not associated with significant changes in the number of TCs across the North Atlantic, except for limited regions in the south-eastern and north-eastern sectors and eastern Canada (Figure 5b).

3.2. Modulation of TC Genesis

We investigate whether the NAO modulation of TC activity results, at least partially, from an influence on TC genesis. To do so, we examine the regression coefficient maps for model \( P_2 \) (Figures 6 and 7). The LF-NAO index not only captures changes in the TC track density but also in the TC genesis density (Figure 6a): the negative coefficients observed in the Caribbean, the eastern Gulf of Mexico and near the Lucayan Archipelago suggest that TC genesis is enhanced in these regions during the negative LF-NAO phase. A smaller area of increased TC genesis is also observed in the western MDR. The pattern is reminiscent of the one in (L. Zhang et al., 2022), obtained from the analysis of the springtime NAO. The number of TC genesis events is estimated to increase by approximately 30%–45% in a strong, negative NAO regime, compared to a positive one (Figure 6b). Conversely, the unfiltered NAO index does not capture a coherent pattern of TC genesis variability: the regression coefficient map in Figure 7a shows a rather noisy, non-significant signal, which lacks spatial organization, except for the south-eastern North Atlantic, as was the case for TC track density (Figure 5b).

4. Influence on the Large-Scale Environment

The coupling between the NAO and the North Atlantic SST has been widely investigated (e.g., Cassou et al., 2004; Czaja & Frankignoul, 1999; Czaja & Frankignoul, 2002; Seager et al., 2000): the NAO atmospheric forcing is typically associated with a tripolar SST anomaly pattern that extends from the subtropical ocean into the high latitudes.

The seasonal SST anomaly composites suggest that both the unfiltered and LF-NAO indices are associated with a tripolar SST anomaly pattern (Figures 8 and 9). During the positive NAO phase, an area of warmer SST...
Figure 5. (a) Unfiltered North Atlantic Oscillation (NAO) regression coefficient for $P_1$, the white dotted area indicates statistically significant coefficients at the 95% confidence level; (b) difference in the expected number of tropical cyclones between the low-frequency NAO 25th and 75th percentile.

Figure 6. Same as in Figure 4 but for tropical cyclone genesis density ($P_2$).
is straddled at higher and lower latitudes by two poles of negative SST anomalies; the reverse is true during the negative NAO phase. As Hoerling et al. (2001) show, low-frequency NAO variability modulates the North Atlantic SST well into the tropical region. By examining the composite fields and regression coefficients from Equation 6 (Figure 8c), we see that the LF-NAO index captures a pattern akin to the NAO SST horseshoe, which extends across the entire MDR and into the Caribbean basin. In this region, SST is negatively correlated with NAO: the negative NAO phase thus features significantly higher-than-average SST. The differences in structure between the composite fields (Figures 8a and 8b) and the coefficient pattern (Figure 8c) in the eastern portion of the North Atlantic basin are primarily due to the inclusion of the AMO in the multiple linear regression (Equation 6). The SST anomaly pattern associated with the unfiltered NAO (Figure 9) shows a fundamental difference compared to that of the LF-NAO (Figure 8). The tripole is shifted to the north, with its southern lobe located mostly in the sub-tropical belt, leaving the MDR only marginally involved.

The NAO is known to also influence upper-level winds across the subtropics (Cassou et al., 2004; Kushnir et al., 2006; Sutton et al., 2000; L. Zhang et al., 2022); similarly, TC activity (Gray, 1984) and structure (Chen et al., 2006; Rogers et al., 2003) are known to be modulated also by variations in the vertical wind shear. Hence, we investigate the NAO-related changes in vertical wind shear. The LF-NAO composites are associated with a tripolar wind shear structure spanning the entire basin: during NAO+ hurricane seasons, positive wind shear anomalies are present between 10 and 20°N and north of 50°N. Conversely, the latitudinal belt between 20 and 50°N is largely occupied by a negative anomaly. The reverse pattern is observed during the NAO− regime (Figures 10a and 10b), where wind shear is significantly weaker than average across the entire MDR and the Caribbean basin. The regression coefficient map Figure 10c confirms that deep-layer wind shear is negatively correlated with the LF-NAO. While the unfiltered NAO composites feature similar tripolar wind shear anomalies, they differ significantly from the LF-NAO ones (Figures 11a and 11b): the wind shear anomalies are larger in magnitude but they mostly affect the subtropics and the mid-latitude, leaving regions of climatologically high TC activity such as the MDR and the Caribbean basin largely unaffected (Figure 11c).

Most tropical disturbances that later become TCs in the North Atlantic travel across the MDR, in particular during the bulk of the hurricane season. The LF-NAO variability captures changes in SST and wind shear across the entire MDR (Figures 8 and 10) that can explain the observed downstream modulation of TC track density
and genesis (Figure 4). Our results agree well with the long-established influences of SST and shear on TCs: strong wind shear does not only prevent TC genesis but also can effectively disrupt an existing TC (e.g., Nolan et al., 2007). Similarly, cooler SST limits the transfer of energy from the upper-ocean to disturbances and mature storms alike, preventing their organization into a TC or actively suppressing them (e.g., Nolan & Rappin, 2008). As such, SST and wind shear have been typically included in genesis potential indices (e.g., Camargo et al., 2007; Emanuel & Nolan, 2004; Tippett et al., 2011). During the negative LF-NAO phase, higher SST and weaker wind shear (Figures 8 and 10) provide favorable conditions for enhanced TC activity. The opposite is true during the positive LF-NAO regime. Other known large-scale influences on TCs (e.g., lower-tropospheric relative vorticity and mid-tropospheric relative humidity) do not display significant covariability with the LF-NAO across the North Atlantic (not shown).

5. Interaction With AMO and ENSO Variability

Modes of climate variability such as ENSO and the AMO mutually affect their respective influences on North Atlantic TCs (Caron et al., 2015). In this section, we focus on the interaction of ENSO and the AMO with the LF-NAO, and its relationship to TCs in the Caribbean/Gulf of Mexico region.

ENSO strongly affects TC activity, in particular across the western sector of the North Atlantic (see Figures S2 and S4 in Supporting Information S1). As shown in Figure 12a, TC activity in the Caribbean/Gulf of Mexico
increases from positive to negative ENSO phases during both LF-NAO+ and LF-NAO− regimes. However, it is during neutral and positive ENSO years that the LF-NAO modulation of TCs appears to be strongest and most significant (p-value\textsubscript{pos} = 0.008, p-value\textsubscript{neu} = 0.003). Conversely, during La Niña events there is no significant variation between positive and negative LF-NAO regimes.

The positive AMO phase favors enhanced TC activity across the North Atlantic (see Figures S1 and S3 in Supporting Information S1), although its influence is more limited across the Caribbean/Gulf of Mexico. Figure 12b reveals important characteristics of the LF-NAO influence on TCs. It is strongest and significant only during the negative AMO phase (p-value\textsubscript{neg} = 0.005), while no discernible relationship exists during AMO+ years. Most importantly, years characterized by negative AMO and negative LF-NAO display levels of TC activity comparable to those in the positive AMO phase, suggesting that the favorable large-scale conditions associated with the negative LF-NAO regime might effectively offset the otherwise unfavorable environment. The results presented are robust to the box extent and to the Niño 3.4 threshold value (not shown).

6. Modulation of North Atlantic TC Rainfall

While TC rainfall estimates are only available over shorter periods, we investigate whether they reflect the LF-NAO influence. Using the LF-NAO index, we construct NAO-relative annual composite TC rainfall maps
from CMORPH (Figure 13) and the ERA5 reanalysis (Figure 14) for the 1998–2019 and 1979–2019 periods respectively.

The composite analysis strongly supports the Poisson regression results: the negative LF-NAO phase is characterized by enhanced TC rainfall over the Caribbean Basin and the Gulf of Mexico compared to the positive phase. The composite difference pattern (Figures 13c and 14c) strongly resembles the regression coefficient map (Figure 4), suggesting that the increase in TC track density associated with tropical SST and wind shear variability (Figures 8c and 10c), directly translates to an enhancement of TC rainfall on interannual time scales. The signal over the Caribbean and Gulf of Mexico is robust across both time periods: the CMORPH composites (Figure 13), however, also feature an enhancement of TC rainfall north-west of the Caribbean and along the East Coast in the positive LF-NAO phase that is not accompanied by an increase in TC track density nor appears robust over a longer period (Figure 14).

7. Discussion and Conclusion

Sources of predictability of TC activity on interannual-to-decadal timescales continue to be the focus of active research for the Atlantic (e.g., Boudreault et al., 2017; L. Zhang et al., 2022) as well as the Pacific basin (e.g., Scoccimarro et al., 2021; L. Zhang et al., 2022). In the North Atlantic, ENSO, the AMO/AMM play a significant
role in modulating the dynamic and thermodynamic state across the MDR. Instead, the importance of the NAO is still debated. In this study, we address the relationship between the NAO and North Atlantic TC activity by statistically modeling TC track and genesis density. Using Poisson regressions, we show that only the low-frequency NAO variability is associated with significant changes in TC activity and that such modulation is characterized by a well-defined spatial pattern. Both TC genesis and activity are negatively correlated with the LF-NAO across the western Atlantic, the Caribbean basin and the Gulf of Mexico. The Poisson regression indicates that an interquartile decrease of the LF-NAO corresponds to a 30%–40% increase in TC track density in these regions. The pattern resembles a leading mode of sub-basin TC activity variability described by Mei et al. (2014), which according to the authors is linked to the NAO. In contrast, we show that the unfiltered NAO index does not capture substantial TC activity variability, either in terms of TC track density or TC genesis.

The observed correlation between the LF-NAO and TC activity is explained in terms of large-scale changes in the deep-layer wind shear and SST. LF-NAO variability is associated with a distinct SST anomaly that features a horseshoe structure, whose southern branch extends across the entire MDR and the Caribbean basin. In agreement with the modeling studies on the NAO-related atmosphere-ocean coupling in the North Atlantic (e.g., Cassou et al., 2004; Sutton et al., 2000), a similar tripolar pattern is exhibited by the 200–850 hPa wind shear anomalies, mainly due to changes in the upper-level winds resulting from the SST-induced changes in tropical convection and heating. As a result, significantly higher-than-average SST and lower-than-average wind shear characterize the negative LF-NAO phase across the MDR and the Caribbean basin, supporting increased TC

Figure 11. Unfiltered North Atlantic Oscillation (NAO) composite wind shear anomaly: (a) NAO+ (top tercile), (b) NAO− (bottom tercile) and (c) wind shear anomaly regression coefficient from Equation 6; the contoured-dotted area indicates statistically significant coefficients at the 95% confidence level.
activity both locally and downstream. While the unfiltered NAO index captures a similar pattern of SST and wind shear variability, its northward shift is such that it does not influence regions of high TC activity, thus explaining its lack of predictive power.

It is worth noting that the physical mechanisms behind the modulation of TC activity by the NAO are like those responsible for the influence of the AMO/AMM. As indicated by Kossin et al. (2007) and Vimont and Kossin (2007), the AMO manifests its effect by exciting the AMM, whose positive phase is associated with higher-than-average SST and lower-than-average wind shear over the MDR. Given the high correlation between the AMO and the AMM (1948–2019, \( r = 0.8 \)), in the statistical models we include the AMO as it represents a good proxy for AMO/AMM variability and allows us to consider a much longer period. For completeness, the analysis has also been performed over the 1950–2019 period using the AMM in place of the AMO and the results remain consistent with those presented in the paper (not shown). Despite acting through similar mechanisms, there are strong indications that the low-frequency NAO captures a different component of TC activity variability: (a) although significant its correlation with the AMO is rather small (\( r = -0.26, r^2 = 0.067 \)); (b) by controlling for the AMO in our statistical models we account for the influence of any covariance between the two predictors, and (c) not only the pattern of the SST and wind shear regression maps (Figures 8c and 10c) are obtained while controlling for the AMO state but they also show distinctive characteristics. Compared to the AMO/AMM pattern, the NAO SST composite displays a much more pronounced tripole structure and, very importantly, the center of action in the tropical North Atlantic is located further west than the typical AMO/AMM one (e.g., Chiang & Vimont, 2004), which can be found near the North African coast. Moreover, the SST pattern does not feature the typical cross-equatorial gradient that constitutes the primary feature of the AMM mode. While the wind shear pattern is more similar to that associated with the AMO/AMM, the largest influence of the NAO in the tropical North Atlantic is also further west compared to that of the AMM (see Figure 1 in Vimont and Kossin (2007) and Figure 3 in Kossin et al. (2007)).

The LF-NAO relationship to North Atlantic TCs is in turn modulated by AMO and ENSO variability. In the Caribbean/Gulf of Mexico region, the largest differences between positive and negative LF-NAO regimes are observed during El Niño or ENSO neutral conditions as well as during the negative AMO phase. Such results could have important implications as the large-scale changes in SST and wind shear associated with the LF-NAO could offset negative AMO/ENSO conditions and support enhanced TC activity despite an otherwise unfavorable large-scale environment.
We further show that the LF-NAO modulation of TC activity has immediate consequences for TC impact. The NAO-induced variability in TC tracks directly translates to changes in the basin-scale distribution of TC rainfall across the North Atlantic. Where TC activity increases in response to low-frequency NAO variability so does TC rainfall. The relationship exists both in reanalysis and satellite-based rainfall estimates and is robust over different periods.

Predictive skills for the NAO during summer are typically low (Dunstone et al., 2018) and the unfiltered NAO index does not appear to capture a substantial pattern of TC activity variability in the North Atlantic. Low-frequency NAO variability, instead, is not only characterized by a certain degree of persistence (lag-1 autocorrelation = 0.45) but also captures significant changes in the dynamics and thermodynamics across the MDR during the hurricane season. In our work we employ a statistical model to infer the role of the NAO, thus we cannot argue that there exists a direct causal mechanism between the LF-NAO and North Atlantic TC activity. In fact, previous literature (e.g., Cassou et al., 2004; Sutton et al., 2000) suggests that tropical SST variability might also in turn force low-frequency NAO variability. The methodology employed in this study also cannot
address the relative contribution of internal and forced climate variability in modulating TC activity. A recent framework for the detection and attribution of regional rainfall changes due to internal and forced variability has been proposed by Risser et al. (2023); given the ongoing improvements in climate models' ability to simulate TC activity, future research should directly address such a consequential question. Our results, nevertheless, reaf-

Figure 14. Same as in Figure 12 but for ERA5 (1979–2019).
Data Availability Statement

Ibtracs data can be accessed at https://www.ncdc.noaa.gov/products/international-best-track-archive (Knapp et al., 2018). CMORPH rainfall data can be downloaded from https://www.ncdc.noaa.gov/products/climate-data-records/precipitation-cmorph (P. Xie et al., 2019). ERA5 hourly rainfall data (Hersbach et al., 2018b) and pressure-level data (Hersbach et al., 2018a) can be obtained from the Copernicus Climate Data Store (https://cds.climate.copernicus.eu/). The merged Hadley-IO SST data (Shea et al., 2022) can be downloaded from the DASH repository at NCAR (https://gdex.ucar.edu/dataset/158_asphilli.html). The NAO-PC based index (J. Hurrell, 2022) can be downloaded from the NCAR Climate Data Guide repository (https://climatedataguide.ucar.edu/climate-data). The unsmoothed AMO (PSL, 2022a) and Niño 3.4 (PSL, 2022b) indices can be downloaded from the NOAA PSL website (https://psl.noaa.gov/data/climateindices/). The Poisson regression analysis is performed in Python using the statsmodels package (Seabold & Perktold, 2010) (https://www.statsmodels.org/stable/index.html).

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