On the Rivers in the Euro-Atlantic Sky

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Abstract

We study the spatiotemporal variability of Atmospheric Rivers (ARs) over Euro-Atlantic region using long-term reanalysis datasets. Winds, temperature and specific humidity at different pressure levels during 1979-2018 are used to study the water vapour transport integrated between 1000-300 hPa (IVT300) as a proxy to ARs. The intensity of ARs in the North Atlantic has been increasing in recent times (2009-2018) with large decadal variability and poleward shift (~5° towards the North) in landfall (1999-2018). Significant bias shown by different reanalysis products in IVT300 compared to ERA5 data is attributed to bias in specific humidity and winds. Different reanalysis datasets show similar spatial patterns of IVT300 in mapping ARs but has a bias of around 40-60 kgm⁻¹s⁻¹ compared to ERA5. The magnitude of winds and specific humidity in the lower atmosphere (below 750 hPa) dominates the total column water vapour and intensity of ARs in the north Atlantic. IVT300 in all reanalysis datasets in the North Atlantic show a standard deviation of 200 kgm⁻¹s⁻¹ which is around 60% of the IVT300 climatology (>300 kgm⁻¹s⁻¹). Though ARs have higher frequency of landfalling over Western Europe in winter half-year.
(WHY); the intensity of IVT300 in winter is 3% lower than the annual mean. On the other hand, lower frequency of ARs in summer half-year (SHY) shows 3% higher intensity than the annual mean. There is a significant impact of the North Atlantic Oscillation (NAO) and Scandinavian blocking on the location of landfall of ARs. Furthermore, there is a strong latitudinal dependence of the source of moisture flux in the open ocean, contributing to the formation and enhancing AR's strength.

**Keywords:** Atmospheric Rivers, North Atlantic Ocean, summer half-year, winter half-year, geopotential, surface latent heat flux, North Atlantic Oscillation and Scandinavian blocking.

**Introduction**

Tropospheric atmospheric dynamics are guided by water vapour in the lower atmosphere (Schneider et al., 1999). Particularly, heat and momentum in the lower troposphere have strong coupling with the movement of moisture in the troposphere. Hence, it is essential to study the tropospheric moisture transport to better understand the global water cycle, synoptic weather patterns and climate change due to enhanced evaporation in recent decades and global warming (Trenberth, 2011). Also, the ocean and atmosphere play a key role in transporting heat and water vapour poleward, respectively. Atmospheric general circulation plays a vital role in circulating water vapour in the lower troposphere. The large-scale land-ocean atmospheric exchange of water demonstrates the coupling of the atmospheric branch of the hydrological cycle (Hack et al., 2006). The global and continental-scale transport of water vapour has important implications for climate variability and hydrology (Brubaker et al., 1994). Hence, atmospheric scientists must consider studying climatological, meteorological and hydrological aspects of the transport of moisture in
the lower atmosphere (Gimeno 2013; Gimeno et al., 2012). It is particularly important to
understand conceptual models of moisture transport to aid research into the origin of continental
precipitation (Gimeno 2014). Also, moisture transport in mid-latitudes plays a key role in guiding
the global atmosphere and climate dynamics in various temporal and spatial scales.

Most of the meridional water vapour transported across midlatitudes (90% of the total mid-latitude
vertically integrated water vapour flux) takes place through narrow corridors in less than 10% of
the zonal circumference. These narrow filaments of poleward water vapour transport are termed
as atmospheric rivers (ARs) (Yong and Newell 1998; Ralph et al. 2004). These transient
filamentary regions occur within the warm conveyor belt of extratropical cyclones in a maritime
environment and are characterized by high water vapour content and strong low-level winds
(Ralph et al. 2004, 2005, 2006). Thus, these corridors tend to be quite narrow (<1000 km wide)
relative to their length scale (>2000 km) (Neiman et al., 2008). The warm conveyor belt transports
both sensible and latent heat, particularly the later contributes to the poleward heat transport that
occurs in the form of water vapour flux from the warm sea surface over oceanic regions serving
as a major moisture source. Most of the water vapour transport within these rivers occurs in the
lowest 2.5 km of the atmosphere due to moist-neutral stratification (Ralph et al. 2005). Hence,
these are also called tropospheric rivers due to their occurrence in the lower troposphere (Yong
and Newell 1994, 1998). The combination of lower tropospheric moist neutrality, strong horizontal
winds, large and concentrated water vapour content yields an occurrence of heavy orographic
precipitation and winds on elevated terrain, which can lead to severe and widespread flooding
(Ralph et al. 2006; Neiman et al. 2002, 2011; Ruby and Qian 2009; Lavers et al., 2011, 2012;
Waliser and Guan, 2017; De Luca et al., 2017), and could further cause landslides to occur over
the adjacent area (Jason et al., 2019). Heavy and untimely precipitation (Kritika et al., 2018; Yan
et al., 2018) from warm ARs also causes preexisting snowpack to melt in high latitudes and poles allowing freshwater inflow into oceans and contribute to the sea level rise (Neff William, 2018; Mattingly et al., 2018), leading to coastal flooding (Khouakhi and Villarini 2016). Snowmelt and intense flooding due to ARs could change the geomorphic processes, biodiversity and mass mortality of wildlife (Joan et al., 2015; Brian et al., 2016). Conversely, ARs could also change the ice sheet surface mass balance over poles by 74-80% through heavy snow accumulation (Irina et al., 2014). Thus, ARs are key to understanding extratropical and polar hydro-climate features through polar warming, sea ice melt, and precipitation (Deanna et al., 2018; Kensuke et al., 2018). Consequently, these mesoscale filamentary features play a key role in the global water cycle and represent an important phenomenon linking weather and climate.

There are numerous studies over midlatitudes documenting the AR characteristics, landfall, and their relationship with the extreme hydrometeorological events (De Luca et al., 2017). Many studies have focused on ARs over the Pacific; particularly on the south-west coast of the United States (Ralph et al., 2006, 2019, 2005, 2006; Neiman et al., 2008; Chapman et al., 2019, and the references therein) and South America (Viale and Nunez, 2011). There are a few studies aimed at the global characteristics of ARs (Waliser and Guan 2017; Guan and Waliser 2017, 2015). Recently there is an increasing focus on the precipitation over Europe and ARs over the North Atlantic (Pasquier et al., 2019; Yang et al., 2016; Lavers et al., 2016; Champion et al., 2015). Recent studies in Asia (Kritika et al., 2018; Yang et al., 2018; Youichi et al., 2017) and Africa (Blamey et al., 2018; Alexandre et al., 2018) have focused on the relationship between ARs and extremes in precipitation. However, the study of ARs over the North Atlantic and Europe needs more attention due to potentially increasing extremes in hydrometeorological events such as snowfall, precipitation and flooding (Millán 2014; Kundzewicz et al., 2006; Van den Besselaar et
Most of the extreme wind events catalogued between 1997 and 2013 over Europe with 93 billion US dollar losses were associated with ARs (Waliser and Guan 2017). Hence, it is essential to study both oceanic and atmospheric processes affecting these anomalies and extremes. AR is one such feature guided by both oceanic and atmospheric dynamics and causes extremes in precipitation and influences the hydrology over Europe. Lavers et al., (2013) studied the relationship between ARs and extreme precipitation across Europe and found that the North Atlantic Oscillation (NAO) has a significant impact on precipitation caused by ARs. The same study highlighted anomalies in central European precipitation patterns caused by ARs over the North Atlantic. According to a multi-model ensemble of the Coupled Model Intercomparison Project (CMIP5); AR frequency is projected to increase 127%-275% by the end of this century, at peak AR frequency regions (45°-55°N) over Europe, under the representative concentration pathway 8.5 (RCP8.5) scenario. This enhanced frequency is associated with the wind changes in the midlatitude jet (Yang et al., 2016). ARs cause 20-30% of all precipitation in parts of Europe and the United States, however with strong seasonality. Also, ARs penetrate further inland over Europe than over the United States (Lavers and Villarini 2015). On the other hand, ARs are in sync with the largest floods over Western Europe and the United Kingdom (Lavers et al., 2011, 2012; De Luca et al., 2017).

Several procedures are in practice to detect, track and forecast ARs in advance using observational, reanalysis, and numerical models (Ralph et al., 2019; Fish et al., 2019; Lavers et al., 2018). Integrated Water Vapor (IWV) (Ralph et al., 2004; Neiman et al., 2008b; Guan et al., 2010) and integrated vapour transport (IVT) (Zhu and Newell, 1998; Roberge et al., 2009; Jiang and Deng, 2011) are the two most common techniques used to define, detect and track ARs. Time integrated IVT, Meteograms and cross-sections are some other methods to study ARs. Both IWV
and IVT consider vertically integrated (between 1000 hPa to 300 hPa or less) horizontal water vapour transport (significant poleward moisture transport) as a proxy to AR occurrence when the standardized IVT was greater than a threshold (Roberge et al. 2009). Accurate atmospheric parameters such as winds, specific humidity, and the temperature at different pressure levels obtained from satellites and reanalysis products are essential to study ARs (Neiman et al., 2009; Dettinger, 2011). Though necessary parameters are available from different platforms, atmospheric reanalysis is the best estimate of the historical state of the Earth’s atmosphere. These datasets are produced by assimilating meteorological/oceanic observations into numerical weather prediction model output. In this work, we aim to study the characteristics of ARs over the North Atlantic such as spatiotemporal variability, bias in mapping ARs using different reanalysis products, frequency and decadal variability of ARs. The objective of this study is also to look at the variability and trend of ARs in the North Atlantic in relation to the different atmospheric parameters and IVT in different layers of the atmosphere. The spatial/horizontal resolution dependence of ARs over the North Atlantic from different reanalysis products was computed with reference to IVT300 mapped using ERA5. The paper is organized as follows. Section 2 describes the data and methods, followed by results and discussions in section 3 and conclusions in section 4.

Data and Methods

We have used six-hourly winds, temperature, and specific humidity data at different pressure levels from six reanalysis products available during 1979-2018. These six reanalysis datasets include 20th Century Reanalysis version 2 (20CR-V2) from the NOAA Earth System Research Laboratories (ESRL), ERA-Interim, ERA5 from the European Centre for Medium-Range Weather Forecasts (ECMWF), Modern-Era Retrospective analysis for Research and
Applications (MERRA-2) from National Aeronautics and Space Administration (NASA), Climate Forecast System Reanalysis version 2 (CFSR-v2), NCEP-NCAR Reanalysis version 2 from the National Center for Environmental Prediction (NCEP). Apart from MERRA-2, which has been available since 1980; all datasets are available from 1979 and have different spatial resolutions. Details of reanalysis datasets are given in table 1. In addition to traditionally mapping ARs using both IVT (Equation 2) and IWV (Equation 4), we also included the temperature of corresponding layers in these algorithms to normalize the computed IVT (nIVT, Equation 3) and IWV (nIWV, Equation 5) and study the difference from the normal approach using different reanalysis products and compared these two methods in the Atlantic. Temperature normalization is done to understand the change in the thermodynamic component of IVT and IWV using the Clausius-Clapeyron equation (1), which states that the water-vapour content of saturated air, $q^*$, increases nearly exponentially with temperature $T$ (Payne et al., 2020).

$$\frac{dq^*}{dT} = \alpha(T)q^*$$  \hspace{1cm} (1)

$\alpha(T)$ is the Clausius-Clapeyron scaling factor, defined as

$$\alpha(T) = \frac{L}{R_vT^2}$$

where $L$ is the latent heat of vaporization and $R_v$ is the gas constant of water vapour. Within the saturated environment at the core of an AR where $q\approx q^*$, a small change in the surface warming would cause specific humidity to further increase. Thus, specific humidity in the upper layers of the atmosphere strongly depends on the increase in layer’s temperature due to increasing surface temperature and the Clausius-Clapeyron scaling factor, $\alpha(T)$ and is approximately 6.6% K$^{-1}$ for surface temperatures causing ARs that are land-falling over California in the present climate ($T = 13^\circ$C) (Dettinger 2011; Gonzales et al., 2019).

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Integrated Vapor Transport (IVT):

\[ IVT = g^{-1} \sqrt{\left( \int_{1000}^{p} QU \, dp \right)^2 + \left( \int_{1000}^{p} QV \, dp \right)^2} \]  

(2)

Normalized IVT:

\[ nIVT = g^{-1} \sqrt{\left( \int_{1000}^{p} \frac{QU}{T} \, dp \right)^2 + \left( \int_{1000}^{p} \frac{QV}{T} \, dp \right)^2} \]  

(3)

Integrated Water Vapor (IWV):

\[ IWV = g^{-1} \int_{1000}^{p} Q \, dp \]  

(4)

Normalized IWV:

\[ nIWV = g^{-1} \int_{1000}^{p} \frac{Q}{T} \, dp \]  

(5)

Where Q is specific humidity in kg kg\(^{-1}\), U and V are zonal and meridional components of winds at different pressure levels measured in ms\(^{-1}\), P is the desired pressure (hPa) up to which the atmospheric parameters are integrated; g is the acceleration due to gravity, which is a constant and is given as 9.8 ms\(^{-2}\) (Neiman et al., 2008a; Lavers and Villarini, 2013a, 2013b). Normalization with temperature is done by dividing Q, U and V with the temperature at a corresponding pressure level. Supplementary table S1 shows the details of the variables and their units. Thus, time-integrated (00, 06, 12, and 18) daily ARs data has been generated from six-hourly reanalysis datasets using IVT (kg m\(^{-1}\) s\(^{-1}\)), normalized IVT (kg m\(^{-1}\) s\(^{-1}\) K\(^{-1}\)); IWV (mm) and normalized IWV (mm K\(^{-1}\)) from the surface to 750 hPa, 500 hPa and 300 hPa. Quantification of ARs over the North Atlantic was done using the climatology and standard deviation for different methods. This is
essential to see the spatial variability including the magnitude of water vapour transport over the North Atlantic and into Western Europe. We have used the AR detection method from Lavers and Villarini (2013b).

The time-integrated (00, 06, 12, and 18) daily AR data have been used in further analysis to study temporal and spatial variability of ARs over the North Atlantic, both in climatic and decadal timescales. In addition to studying biases in different atmospheric parameters, the annual and seasonal climatology and strength of ARs at different layers were studied. For the intercomparing of reanalysis datasets, we consider ERA5 as the reference dataset because of its enhancements in parameterization and resolution. Furthermore, the study focuses on the variability of ARs intensities in different products, including major categories of ARs and their frequencies in the North Atlantic. Linear regression analysis with student t-test done in the study helps to understand the spatial trend in the IVT300 (ARs); followed by the study to understand the state of the atmosphere using atmospheric parameters to describe the characteristics of land-falling of ARs.

A general approach used to map ARs is using IVT300 by considering pressure levels from surface (1000 hPa) to 300 hPa (Neiman et al., 2008b; Guan et al., 2010, Lavers et al., 2011). A few studies also considered 900 hPa as the surface reference level (Gorodetskaya et al., 2014); 500 hPa (Yang et al., 2016) and 200 hPa (Sellars et al., 2017; Mattingly et al., 2018) as upper limits. Shields et al., (2018) compiled all available methods including thresholds to map ARs globally as a part of describing the Atmospheric River Tracking Method Intercomparison Project (ARTMIP). Hence, there is persisting ambiguity in using reference pressure levels to map ARs. Therefore, here we quantify the magnitude of annual and semi-annual IVT in different layers. For this purpose, the total column (1000-300 hPa) has been divided into sub-layers consisting 500-300 hPa (IVT_Upper), 750-300 hPa (IVT_Middle), 750-500 (IVT_Lower) which lie above 750 hPa
pressure level in addition to computing IVT500 (1000-500 hPa) and IVT750 (1000-750 hPa). This exercise helps to map the strength of IVT and spatial variability in these layers, which is a function of exponentially decreasing water vapour pressure with height.

Results and Discussions

Climatology and standard deviation of ARs over the North Atlantic:

The annual climatology from ERA5 daily data using IVT and nIVT methods shows the highest AR intensity in the region enclosed between 30°N-60°N (Figure 1). Though the intensity of AR varies from event to event, on average IVT300 (IVT between 1000-300 hPa) (Figure 1a) over the North Atlantic is around 300 kgm⁻¹s⁻¹ and is in line and directed along with the westerly wind over this region. The maximum nIVT300 (nIVT between 1000-300 hPa) over the North Atlantic is in coherence with the maximum IVT300 and along the same path with maximum values (>1 kgm⁻¹s⁻¹K⁻¹) concentrated over the central North Atlantic (Figure 1b). The nIVT300 is accounted for available IVT300 per unit temperature, is a proxy to fractional changes in available specific humidity in the atmospheric column per degree of atmospheric warming. In upper layers of the atmosphere, α varies with the varying temperature. Hence, α increases with the decreasing temperature with height and amplifies changes in the specific humidity aloft and is larger in the upper troposphere. On the other hand, increasing specific humidity in upper layers tends to release more latent heat flux with ascending air, and decrease the lapse rate with warming and thus increasing the temperature with height. If the vertical column of the atmosphere is saturated and has moist-neutral conditions, the combination of these factors implies a rate increase in IVT that is substantially larger than that of near-surface water vapour (Payne et al., 2020). Hence, a fractional change in IVT is a reasonable approximation to the thermodynamic contribution to IVT change. Thus, on top of concentrated warm coastal surface waters due to Gulf stream over western
North Atlantic causing higher evaporation; specific humidity advecting from tropics could be saturating the upper troposphere over the central North Atlantic and showing higher IVT300 and nIVT300.

Though AR mapping and characteristics study initially was started using IWV, the importance of tracking the AR made IVT as a widely used method. However, using IWV would give estimation of concentration of total column condensable water vapour at a given instance (Ralph and Dettinger 2011; Gimeno et al., 2014). Climatology of IWV300 (IWV between 1000-300 hPa) (Figure 1c) and nIWV300 (nIWV between 1000-300 hPa) (Figure 1d) shows the gradient of water vapour varying from a maximum at the equator and fading towards the pole. Using nIWV here shows the amount of total column condensable water vapour per degree Kelvin. The amount of evaporation caused by solar heating and strength of the near surface winds determines the extent and the scale of the water vapour. However, the occurrence of AR over a region and its magnitude guided by the amount of precipitable water vapour are not only bound to the availability of specific humidity in the atmosphere but also on the magnitude and direction of winds carrying the water vapour. Hence, the higher intensity of ARs over the North Atlantic and Western Europe are in the direct vicinity of the region of occurrence of extratropical cyclones and associated strong surface wind speeds (Pinto et al., 2013; De Luca et al. 2017) and along the path of the subtropical westerly winds. Although all the methods used in mapping ARs show higher values over the western North Atlantic, the origin of ARs and the region of moisture flux into ARs in this part of the ocean are still debatable. These elongated features are also affected by the synoptic weather conditions, and their magnitude depends on the midway convergence of water vapour flux from adjacent areas. Despite IVT300 climatology showing a maximum of 300 kgm⁻¹s⁻¹, each AR could be different in magnitude and its strength varies as per the state of the atmosphere at a given instance.
One example of an AR from 6th March 2002 mapped using four different methods in Figure 2 has IVT300 higher than 500 kgm$^{-1}$s$^{-1}$ (Figure 2a). This event was one of the intense ARs that occurred over northern Europe and caused excess rainfall over Britain and southern Scandinavia. While the IVT300 is narrow and short, nIVT300 (Figure 2b) shows the adjacent regions saturated with water vapour. The advected moisture from these surrounding regions could enhance the intensity and lifetime of the AR over a given location. Thus, nIVT300 is a useful method in mapping the true characteristics and saturated water vapour content in AR. Similarly, IWV300 and nIWV300 (Figures 2c, 2d) for this event show the origin of AR and the source of the advection, which is, in this case, occurred from the warm tropical region (20$^\circ$N) enriched with high specific humidity.

**AR intensity and bias in reanalysis data**

In the North Atlantic, different reanalysis products used to map the ARs show variability in magnitudes (Figure 3). The climatology (shaded) and standard deviation (contours) of ERA5 (Figure 3a) has lower IVT300 intensity than any other reanalysis products used; while ERA-Interim has higher climatology (intensity) and standard deviation. The highest variability (standard deviation of 200 kgm$^{-1}$s$^{-1}$) is around 60% of the maximum values of climatology (>300 kgm$^{-1}$s$^{-1}$) in all reanalysis datasets. Both climatology and standard deviation of IVT300 are higher in JJA and lower in MAM and has strong variability (Supplementary Figure 1a, 1b), and all reanalysis products show similar patterns (Figure 3a-3f). Although these values vary with seasons, both SON and DJF have longer stretch/extent of higher climatology and standard deviation values in the North Atlantic. Similarly, both these values have large spread and reaching Western Europe in winter half-year (WHY or ONDJFM), showing a high frequency of ARs during this time (Lavers...
et al., 2011, 2012). Low frequency in summer half-year (SHY or AMJJAS) mainly concentrated over the central Atlantic.

All reanalysis data sets are developed using numerical and statistical approaches integrated with observations with possible bias corrections. Thus, all these reanalysis datasets show a similar spatial pattern over the North Atlantic, but the difference in magnitudes is explained by the variability in the magnitudes of Q, U and V, which could be further due to bias in observations, discrepancies in product development. To illustrate it further, we compared the atmospheric parameters used (Q, U, and V) to map AR in 20CR (coarse resolution) with the ERA5 (high resolution data) (Figure S3). A simple interpolation is used to match grids points of parameters in ERA5 with 20CR data as these data sets have a different spatial resolution. The climatology of these individual parameters during 1979-2014 shows that 20CR overestimating (ERA5-20CR) the magnitude (Figures S3d-S3f) compared to the ERA5 (Figures S3a-S3c). Hence, the 20CR data has a bias of 0.5 gKg$^{-1}$ in Q, 1 ms$^{-1}$ in U and V components of wind in the North Atlantic (Figures S3g-S3i). However, this would not be obvious for different seasons and different ARs in the Atlantic due to strong seasonal variability of IVT300 intensities and atmospheric state which makes each AR unique event.

In the case of AR mapped on 6th March 2002, different reanalysis products show significant bias in IVT300 compared to ERA5 (Figure 4). Both MERRA and ERA-Interim show positive bias at the head of the AR (AR path is marked as the grey arrow in Figure 4) and negative bias in the tail (Figure 4a, 4b). On the other hand, NCEP (NCAR, CFS) and 20CR have a strong negative bias on aggregate (Figures 4c-4e). Both these positive and negative biases are around 50 kgm$^{-1}$s$^{-1}$ in magnitude and are of 10% of the total magnitude of AR (~500 kgm$^{-1}$s$^{-1}$) (Figure 2). The variability in the magnitude of IVT300 in different products might lead to bias in the intensity, estimation of
precipitation and winds during landfall. Hence, we use ERA5 data as a standard dataset in our further analysis in the following sections.

**Spatio-temporal variability of IVT300**

The annual, WHY and SHY mean IVT computed using ERA5 data in IVT_Upper (Figures S4a, S4d, S4g), IVT_Middle (Figure S4b, S4e, S4h) and IVT_Lower (Figure S4c, S4f, S4i) are shown in Supplementary Figure S4. Due to low saturated water vapour in the higher altitudes, IVT_Upper in the North Atlantic has lower magnitude ($20 \text{ kg m}^{-1}\text{s}^{-1}$) as compared to IVT_Middle ($>80 \text{ kg m}^{-1}\text{s}^{-1}$) and IVT_Lower ($70 \text{ kg m}^{-1}\text{s}^{-1}$) during all seasons. Though the magnitude is less, winds in the IVT_Upper plays a key role in guiding these narrow filaments of ARs. As the IVT_Middle (750-300 hPa) includes IVT_Upper (500-300 hPa), the total IVT in the 750-500 hPa layer is $60 \text{ kg m}^{-1}\text{s}^{-1}$. When separating these pressure levels, IVT shows a dipole pattern with a low bellow 20°N over the northwestern African coast and a high in the central North Atlantic extending from 30°N to 60°N. The green rectangular box in Figure S4b shows the region with maximum IVT (30°N-60°N, 80°W-0). The magnitude of high in the dipole is further increased during SHY than other periods in all layers (Figure S4g-S4i). Similarly, the low has become further less during WHY (Figure S4d-S4f). Thus, IVT has maximum strength during SHY which could be due to increased evaporation over the warm waters in the North Atlantic. Figure 5 shows the strength of annual, SHY and WHY mean IVT in the central North Atlantic (30°N-60°N, 80°W-0) computed from ERA5 data using different reference pressure levels at the top (300 hPa, 500 hPa and 750 hPa) with respect to 1000 hPa. No significant difference was seen between IVT500 and IVT300 ($\sim12 \text{ kg m}^{-1}\text{s}^{-1}$) during the study period. But, IVT750 contributing $\frac{3}{4}$ of the total strength of IVT300 and IVT500. Thus, the strength of the IVT300 and IVT500 depends on the near-surface processes below 750 hPa. While there were no large changes in the strength of the IVT in the individual
layers with seasons; IVT in SHY has 3% higher magnitude, whereas IVT in WHY shows 3% lower magnitude compared to the annual mean (Figure 5). Hence, mean IVT is high in the below 500 hPa of the atmosphere irrespective of the season. Thus, improved parameterization, in addition to accurate and high-resolution atmospheric data, at least up to 500 hPa would be handy in better estimating the strength of the IVT in the North Atlantic.

Furthermore, we show a Hovmöller diagram (Figure 6) of the monthly IVT300 in the central North Atlantic averaged between 30°N-60°N along 80°W-0 during 2014-2018 to study the seasonal variability of peak IVT300 in more detail. IVT300 peaks in the western Atlantic (along the east coast of North America) during summer months. Due to large temperature and pressure gradients from south to north coupled with extratropical cyclone season; high IVT300 has been shifting towards the eastern Atlantic in winter as marked with transparent arrows in Figure 6 and thus causing frequent ARs landfall over Western Europe during WHY. Yet, the extent, location and movement of the peak IVT300 were not constant and have large interannual variability with relatively low values during the spring season and hence the low AR activity during this time. This interannual and intraseasonal variability in IVT300 could be caused by the altering winds over this region. To study this further, we explored the decadal variability and trend in IVT300 and related atmospheric components in the following section.

IVT300 decadal variability and trend

It is evident that recent climate change cause global warming and alter the global water cycle. On this note, it important to look for changes in the IVT variability and trend during the past decades due to warming surface and enhanced evaporation as the changing Clausius-Clapeyron scaling factor α(T) which could increase the total water vapour content in the individual
atmospheric layers. In Figure 7, we show the decadal trend and variability of IVT300 (daily anomaly). Similarly, IVT of different layers of the atmosphere and its dependency on the variable atmospheric parameters using ERA5 data is shown in Figure 8. For this purpose, we used the same region in the central North Atlantic (30°N-60°N, 80°W-0). Figure 8a shows an increasing annual trend of IVT300 anomaly (black line) in each decade over this region. IVT300 anomaly show significant seasonal and interannual variability. Though the overall trend shows an increasing IVT300 anomaly over this region with 1063 kgm⁻¹yr⁻¹ in the study period, the decadal trend has seesaw oscillations. An annual IVT300 anomaly trend (199.1 kgm⁻¹yr⁻¹) in the first decade, i.e. 1979-1988 (red) is dominated by the annual trend in the second decade (green) with an increase of 1591 kgm⁻¹yr⁻¹ (1989-1998). Similarly, a large increase in the annual IVT300 anomaly in the recent decade (purple) with 4122 kgm⁻¹yr⁻¹ (2009-2018) dominates the previous decade (blue) with a moderate annual increase of 42.52 kgm⁻¹yr⁻¹ (1999-2008).

This increasing annual IVT300 trend in each decade is in coherent with the increasing IVT below 750 hPa and IVT_Lower (750-500 hPa) (Figure 8a). Particularly IVT750 has contributed more to the large increase in second (1989-1998) and fourth decades (2009-2018). As the IVT is proportional to Q, U and V; changes in these parameters would impact these trends. Thus, the large trend of IVT300 in the second decade is dominated by the availability of Q in the near-surface layer (1000-750 hPa) which has an annual increasing trend of 2.5 gkg⁻¹ and is largest in all decades (Figure 8b). However, the negative trend in the zonal and meridional components of wind (Figures 8c, 8d) in all layers during the same time, guides the total trend in the second decade. Though Q has a positive trend in the first and third decade, the negative trend in wind components during the same time in different layers caused the IVT annual trend to be moderate in these decades. On the other hand, the annual trend of both Q and wind components (U and V) were positive in the fourth
decade (Figure 8b-8d) and thus led to a strong increase in the annual IVT300 in all layers (Figure 8c). Though the mean IVT flow is zonal, in the last two decades the meridional wind shows a positive trend (Figure 8c), which could be increasing the flow towards north and driving ARs poleward.

The spatial trend analysis significant at 95% during annual, WHY and SHY using daily IVT300 data from ERA5 is shown in Figure 9. While the annual trend shows a rapid increase (3000 kgm\(^{-1}\)yr\(^{-1}\)) of IVT300 at 20°N in the Atlantic and along the western Atlantic extended into central Atlantic with mean annual IVT300 increase of 2000 kgm\(^{-1}\)yr\(^{-1}\), there was no significant increase in IVT300 over southwestern Europe during the study period (Figure 9c). There are seasonal differences, where both WHY and SHY show opposite spatial trends. IVT300 was increased in the central Atlantic and the southwestern United Kingdom during SHY, which could be triggered by the large IVT300 available over the western Atlantic during this time (Figure 9b, Figure 6). Though the WHY show opposite patterns with a negative trend in IVT300, the low is over the northern United Kingdom. There was a moderate increase in the IVT300 trend along southwestern Europe and the region below 20°N has a large positive trend during WHY (Figure 9a). IVT300 has been increasing poleward in recent times with a strong positive trend along 45°W during all seasons, which could lead to intense AR moving towards the north.

**Categories and frequency of ARs over the North Atlantic**

The spatial variability of frequency of ARs over North Atlantic using different categories of IVT300 is shown in table 2. We distinguish the daily IVT300 based on Ralph et al., (2019) with some minor changes to the thresholds, but only using the magnitude of the intensity at each grid point or location in the selected region without considering the duration of the event. The spatial frequency has been computed using the percentage of the ratio of the number of days of IVT300
of the specific category to the total number of days in the study period (14610 days). Thus, cat 1 events are more frequent in the North Atlantic which occurs at 50% of the time (Figure 10a) along the southwest coast of Europe and in the central Atlantic. Other category (cat 2, cat 3 and cat 4) events are less frequent (<15%) over the Euro-Atlantic region (Figure 10b-10d). The humidity source of this intense IVT300 is along the western Atlantic and a few events are reaching the west coast of Europe. Thus, the frequency of intense ARs over Europe is less with cat 2, IVT300 being at 8%, cat 3 and cat 4 are at below 1% of the time. Nonetheless, the rare intense events which occur at less than 1% of the time potentially cause large damage over coastal areas. To investigate the same along Western Europe, we draw the frequency histogram (Figure 11a) and compute the probability density function (Figure 11b) along 11°W as a gateway to Western Europe. This is different compared to Lavers et al., (2013) who considered 10°W as the reference longitude which intersects with some parts of the land over the United Kingdom. Assuming 11°W and 35°N-70°N would eliminate the IVT300 interaction with land.

While most of the IVT300 values along the west coast during the study period are below 300 kg m⁻¹ s⁻¹ (Figure 11a), the values reaching 800 kg m⁻¹ s⁻¹ in a few instances could lead to extreme ARs. Similarly, the probability density function computed along the same boundary (Figure 11b) shows the IVT300 could reach up to 1400 kg m⁻¹ s⁻¹ and cat 1 IVT300 has the higher probability of occurrence (>0.02) over Western Europe than other categories. The decadal analysis along the same longitude (Figure 12) shows an increasing extreme IVT300 values and their poleward shift in recent decades. All categories show peak frequency between 40°N-60°N and there is no explicit decadal variability of cat 1 IVT300 along 11°W (Figure 12a). But cat 2 (Figure 12b,12c) shows low frequency during the first (black) and third decades between 40°N-60°N (green). On the other hand, the frequency of cat 3 and cat 4 IVT300 has been increasing with time.
(Figure 12d) and there is poleward movement of ~5° towards north and crossing 60°N. The changes in the atmospheric state and the synoptic condition in recent decades could be causing the poleward movement of the intense IVT300. Hence, in the following section, we study the state of the atmosphere during the occurrence of IVT300 over Western Europe.

**Atmospheric state and synoptic conditions**

The Scandinavian blocking including phases of the NAO dominates weather patterns over Europe and Scandinavia through the impact on precipitation and temperature (Madonna et al., 2017). While these patterns are persistent in the North Atlantic-European sector irrespective of the seasons, mostly these patterns control the wintertime weather regimes (Dawson et al., 2012; Hannachi et al., 2017). On the other hand, Western Europe receives more frequent intense IVT300 (ARs) in wintertime than in any other season. To study the atmospheric and synoptic conditions while IVT300 occurrence and landfall along Western Europe, we study the composite of 500 hPa geopotential (GP) (Figure 13) and surface latent heat flux (SLHF) (Figure 14) anomalies following Lavers et al., (2013) along 11°W using 5° latitude bins spanning 35°N-70°N (Figure S5). Contrary to Lavers et al. (2013), who used only 10 intense ARs, we computed GP and SLHF anomaly composites with all instances (days) where IVT300 greater than 200 kgm⁻¹s⁻¹ in these latitude bins. Initially, we computed these daily GP and SLHF anomalies with respect to the same time (day) period during 1979-2018. Then, these anomalies were picked with respect to time and location of the occurrence of IVT300 greater than 200 kgm⁻¹s⁻¹ within selected bins and the composite mean anomaly was calculated for each latitude bin.

GP shows a tripole pattern with positive anomalies over south of the Iberian Peninsula (Figure 13a and 13b); Iceland and Greenland, and negative anomalies extend from Britain to the Iberian Peninsula. This is also termed as an Atlantic ridge regime with blocking mainly offshore
of the Iberian Peninsula due to Iberian wave breaking (Davini et al., 2014) or southwest European
blocking (Woollings et al., 2010) leading to the southward occurring ARs (35°N-45°N). The
Greenland anticyclone regime occurs mainly over Greenland resembling the negative phase of the
NAO. This negative NAO arrangement would block the flow over northern Europe and the North
Atlantic storm track and related heavy precipitation and thus impacts southern Europe (Pinto and
Raible, 2012). The zonal regime with very little blocking resembling the positive phase of the
NAO. In a positive NAO phase, negative GP anomalies (Figure 1c and 1d) in the 45°N-55°N
latitude band favours occurrence of frequent IVT300 within the extratropical cyclones causing
rainfall over northern France, through the western British Isles to Norway. A Scandinavian
blocking regime is associated with blocking mainly over the European continent and Scandinavia.
The occurrence of IVT300 (ARs) and their associated precipitation in the north between 55°N-
70°N is related to Scandinavian blocking with the dipole of positive anomalies near the British
Isles and negative anomalies over Greenland and Iceland (Figure 1e-1g). Although both NAO
and Scandinavian patterns have strong relation with IVT300 occurrence in Europe, it is not obvious
that each IVT300 (AR) landfall would follow the same synoptic weather patterns as the spatial
pattern of the atmospheric state would vary significantly with time over a region.

Southernmost IVT300 events are drawing water vapour from both western and eastern
North Atlantic as SLHF anomalies show a dipole pattern with positive anomalies on either side
(Figure 14a, 14b). Thus, these regions act as a major source to moisture entraining into ARs and
impact the intensity of IVT300. The north-central Atlantic is the source of moisture for the IVT300
in the 45°N-55°N latitude band (Figure 14c, 14d). Further, a dipole pattern with positive SLHF
anomaly in the west and negative in the east fueling IVT300 in the far north. Though the positive
anomalies over the North Atlantic could lead to intensifying IVT300 in the north, the cold sea
surface and associated negative SLHF anomalies over the Scandinavia could control the total moisture flux into the IVT300 and hence the intensity of ARs over this region.

Conclusions

We have studied the spatiotemporal variability of water vapour transport (IVT and IWV) as a proxy to ARs in the Euro-Atlantic basin using six-hourly ERA5 data and evaluated five other reanalysis data sets available from NOAA, NASA, ECMWF and NCEP during 1979-2018 with ERA5. We use IVT and IWV methods to map the water vapour transport in different atmospheric layers, the North Atlantic and normalized with temperature to study the water vapour variability with temperature. Both IVT and nIVT proved to accurate enough to map ARs. Though the IVT shows seasonal and semi-annual variability; the mean annual intensity of IVT300 is 300 kgm⁻¹s⁻¹, and the standard deviation is at 60% of the intensity in the North Atlantic. On the other hand, both these values vary in different reanalysis products, with recently released ERA5 showing lower climatology and standard deviation whereas ERA-Interim has higher values compared to other reanalysis datasets. However, the average bias in other datasets is around 60 kgm⁻¹s⁻¹ as compared to ERA5 which amounts to 22% of the total observed IVT flux. The bias in the magnitude of IVT in different layers is directly proportional to the bias in the Q, U and V of the respective layers.

Both the accuracy and magnitude of atmospheric variables at different pressure levels (Q, U and V) in mapping ARs are highly dependent on the resolution of the data obtained. Many of the existing algorithms and mapping techniques are using atmospheric data from satellites and numerical models. Numerical models have limitation in integrating the discretized version of the Navier-Stokes equations. Due to uncertainty in initial conditions, numerical approximation, and model deficiencies, the error increases non-linearly and thus have decreasing forecast skill in simulating the state of the atmosphere with a good lead time (e.g. Lorenz, 1963). As the filament
structures move with time, and the Eulerian method used to map filaments make it hard to use observations. On the other hand, most of the ARs, originate from the large open oceans through both local evaporation and remote moisture flux convergence. Land-based stations could be handy in measuring the atmospheric parameters while the AR approach in land and landfall. Data obtained from both satellites and statistical methods have limitations in forecasting the landfall and intensity of ARs well in advance. In recent times machine learning techniques (Chapman et al., 2019; Kashinath et al., 2020) have evolved as other alternatives. However, the mean error in estimating the intensity of ARs through IVT is around 40-60 kg m\(^{-1}\) s\(^{-1}\) using different sources of data including data from reanalysis and amounts to 22% of mean observed flux (Chapman et al., 2019, Lavers et al., 2018).

While most of the water vapour flux located below 500 hPa due to rapidly decreasing saturated moisture flux with height, the upper layer winds are key to transport the flux poleward. Hence, the accurate and high-resolution atmospheric parameters at least up to 500 hPa could improve the detection and tracking of ARs in the North Atlantic. On the other hand, the variability and trend of Q, U and V below 750 hPa guide the strength of the total column IVT. Thus, Q, U, and V below 750 hPa control the magnitude of IVT in the North Atlantic which shows an increasing decadal trend with seesaw decadal variability. The IVT in the North Atlantic shows interannual variability with the zonal movement of peak values from the western Atlantic in summer to the eastern Atlantic in winter. However, the strength of the IVT in the Atlantic is 3% higher in summer as compared to annual mean due to strong evaporation from the warm ocean than 3% low in winter. While the semi-annual spatial trend of IVT300 shows an opposite pattern, the annual trend of IVT300 shows an increasing water vapour flux over the western Atlantic with a poleward movement of this flux during the last decades. Thus, the higher latitudes encountering
intense ARs in recent times. Though the category 1 IVT300 are more frequent (50%) in the North Atlantic, particularly over 40°N-60°N, the rarely occurring (15%) higher category events could cause extreme precipitation, flooding and winds over Western Europe. The atmospheric state and synoptic weather guided by Scandinavian blocking and both phases of NAO set the landfall location of ARs along Western Europe.

There are several questions still open for the future work. We aimed at studying the impact of climate indices such as El Niño-Southern Oscillation (ENSO), Atlantic multidecadal oscillation, Atlantic zonal mode variability and changing synoptic circulation patterns on the intensity and frequency of ARs in the North Atlantic and the Western Europe. Furthermore, we will study the influence of oceanic parameters and the subtropical convection over the source regions in open oceans to understand the rapid enhancement of IVT300. Looking at the surface and sub-surface oceanic parameters in the North Atlantic Ocean would be handy to understand the strength of the IVT300 over the region. Also, changes in the surface temperatures over the Gulf current and subtropical gyre in the North Atlantic Ocean might give some insights on ARs variability. Changes in the coastal sea level and subsurface processes are also key to understand while studying impacts of land falling ARs. Similarly, we will investigate the implications of recent poleward shift in the location of landfall ARs in the North Atlantic on changes in Greenland and Arctic mass balance which is out of scope of present work.

Data availability

All data used in the study are freely available online from the corresponding data sources cited in the article. However, data that support the findings of this study are available on request from the corresponding author.
Code availability

All codes used in this study are available on request from the corresponding author.

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**Tables:**

Table 1: Details of reanalysis datasets used in the present study

| Dataset                              | Source                      | Availability | Temporal Resolution | Spatial Resolution | Reference                |
|--------------------------------------|-----------------------------|--------------|---------------------|--------------------|--------------------------|
| 20th Century Reanalysis (20CR-V2)    | ESRL: PSD/NOAA             | 1851-2014    | 6 hourly            | 2 x 2 x 24, 1000 - 10 hPa | Compo et al., (2011)     |
| ERA-Interim                          | ECMWF                       | 1979-2018    | 6 hourly            | 0.75 x 0.75 x 60, 1000 - 0.1 hPa | Dee et al., (2011)       |
| ERA5                                 | ECMWF                       | 1979-present | 6 hourly            | 0.25 x 0.25 x 37 1000 - 1 hPa | Hersbach, H et al., (2017) |
| Modern-Era Retrospective analysis for Research and Assimilation Office | Global Modelling and Assimilation Office | 1980-present | 6 hourly            | 0.5 x 0.625 x 42 1000 - 1 hPa | Gelaro et al., (2017)     |
Table 2: Categories of IVT300 based on intensity

| S. No | Category | Threshold (kg m\(^{-1}\) s\(^{-1}\)) |
|-------|----------|----------------------------------|
| 1     | Cat 1    | 200 ≤ IVT300 < 500               |
| 2     | Cat 2    | 500 ≤ IVT300 < 750               |
| 3     | Cat 3    | 750 ≤ IVT300 < 1000              |
| 4     | Cat 4    | IVT300 ≥ 1000                    |
Figures:

Figure 1: Climatology of ARs computed from daily ERA5 data using four different methods in the North Atlantic.
Figure 2: AR event on 2002 March 06 mapped using four different methods in the North Atlantic using ERA5 data.

Figure 3: Climatology (coloured areas) and standard deviation (magenta lines) of IVT300 in the North Atlantic from all reanalysis data used in the study.
Figure 4: Bias in reanalysis products compared to ERA5 data in mapping AR on 2002 March 06 using IVT300 algorithm. AR path indicated by a black transparent arrow.

Figure 5: Strength of annual, SHY and WHY mean IVT (kg m⁻¹ s⁻¹) in different layers.
Figure 6: IVT300 monthly variability along 80°W-0 during 2014-2018
Figure 7: Decadal trend of IVT300 daily anomaly averaged in the central North Atlantic (30°N-60°N, 80°W-0).
Figure 8: Decadal trend (significant at 95%) and variability of (a) IVT (b) specific humidity (c) zonal wind (d) meridional wind of different layers in the central North Atlantic (30°N-60°N, 80°W-0).
Figure 9: Spatial trend analysis during (a) WHY, (b) SHY and (c) Annual using daily IVT300 (at 95% significance). White areas indicate no significant trend.
Figure 10: Spatial frequency analysis of different categories of daily IVT300.
Histogram magnitude

IVT300 Range

Latitude: 35 N - 70 N
Longitude: 11 W

(a)
Figure 11: (a) Histogram and (b) probability density function of daily IVT300 (kg m$^{-1}$ s$^{-1}$) along 11°W at latitudes 35-70 N.
Figure 12: Frequency of daily IVT300 (kgm$^{-1}$s$^{-1}$) along 11°W.
Figure 13: Composite of geopotential anomaly along 11°W using different bins
Figure 14: Composite of surface latent heat flux anomaly along 11°W using different bins