Euro-Atlantic regimes and their teleconnections

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1. Introduction

Bridging the traditional gap between the spatio-temporal scales of weather and climate is a significant challenge facing the atmospheric community. In particular, progress in both medium-range and seasonal-to-interannual climate prediction relies on our understanding of recurrent weather patterns, hereafter weather regimes, and the identification of specific causes responsible for their favoured occurrence, persistence or transition. The weather regime paradigm has been extensively used to investigate the midlatitude atmospheric variability. The day-to-day meteorological fluctuations can be described in terms of temporal transition between regimes. The year-to-year (or longer timescale) climate fluctuations can be interpreted as changes in their frequency of occurrence provided the hypothesis of long-term quasi-stationary climate. This climate-oriented interpretation for weather regimes is shared with the so-called continuum paradigm (Franzke and Feldstein 2005) and appears to be very useful to better understand the low-frequency fluctuations of the Northern Hemisphere atmospheric teleconnection patterns. A brief description of the weather regime paradigm is given in the following section. Euro-Atlantic regimes are then presented as well as their teleconnections at intraseasonal and interannual timescales. A quick application for operational seasonal forecast within the EUROSIP framework is illustrated at the end of this paper. Our study is limited here to the winter season.

2. Weather regimes: concept and definition

Travelling synoptic pressure systems or storms contribute to a significant fraction of the daily to interannual variability of the extratropical climate. Those are linked to the unstable nature of the upper-level westerly jet stream and interact with circulation patterns of larger scale, the weather regimes, in which they are embedded. Weather regimes could be understood as envelopes for daily atmospheric variability; they have a typical 6–10 day nominal persistence, are spatially well defined (typically the width of an oceanic basin) and limited in number. They can be viewed, in midlatitudes, as the preferred and/ or recurrent quasi-stationary atmospheric circulation patterns produced by the interaction between planetary-scale and synoptic-scale atmospheric waves (e.g. Ghil and Roberston 2002). Their spatial (basin-wide) characteristics and their temporal (persistence) characteristics are such that, citing Molteni et al (2006), "weather regimes or flow regimes should be regarded as statistical-dynamical equilibria in phase space, which are defined by averaging the dynamical tendencies on a timescale longer than the typical period of baroclinic transients". Decomposition in weather regimes could thus be interpreted as an efficient spatio-temporal filter of the turbulent North Atlantic circulation.

Weather regimes are traditionally obtained using cluster analysis or classification methods (Wicks 1995). Those organize pressure maps into nested sequences of clusters forming a growing tree association (hierarchical method, e.g. the Ward classification, Cheng and Wallace 1993 among others), or iteratively perform the classification from predefined initial states randomly selected from the total sample, according to a given number of cluster *k* (partition method, e.g. the *k-means* approach, Michelangeli et al. 1995 among others). More complex approaches, e.g. the Self-Organizing Map method arising from the field of artificial neural network (Johnson et al. 2008, among others) have been also recently proposed. Statistically or technically speaking, weather regimes are thus classes of atmospheric circulation patterns gathered together from a similarity criterion. Those classes are defined by their mean conditions, or centroids, by their variance and by their frequency of occurrence.

As reported in many studies, there is always a part of subjectivity associated with the spatial domain retained for computation of weather representative patterns as well as their number. Standard reproducibility and classificability tests are usually applied to objectively define *k* and to assess the robustness and the consistency of the partition (e.g. Michelangeli et al 1995, Straus et al 2007). Despite those checks, the significance of the weather regimes and their existence itself based on the presence of multi-modality of the midlatitude atmospheric circulation are still controversial (Stephenson et al 2003, Christiansen 2007, see introduction in Hannachi 2010 for a more complete discussion). It is beyond the scope of this paper to participate to this debate and the weather regimes should be interpreted here as a reading grid of the extratropical atmospheric variability. It is now widely recognized that changes in their occurrence and intrinsic properties may be an important issue for medium-range (weekly to monthly) to climate change forecasts (decadal to trend). Their spatial-temporal characteristics are such that they appear to be promising candidates to optimally extract potential external forcings (e.g. tropical ocean, stratosphere, greenhouse gazes etc.) on the extratropical atmosphere, thus allowing for higher potential climate predictability at midlatitudes.

3. North Atlantic-Europe weather regimes

Wintertime (December-March) North Atlantic-Europe weather regimes (NAE-WR) are first documented based on the classification of 500hPa geopotential (Z500) anomaly maps from ERA40 reanalysis over 1958-2002. Consistently with previous literature (e.g. Vautard 1990), four NAE-WR are obtained (Fig. 1a–d). The corresponding temporal evolution of the total number of days attributed to a given regime over the complete winter season is given in Fig. 1e-h. The two first Z500 regimes can be viewed as the negative and positive phase of the North Atlantic Oscillation (NAO- and NAO+, respectively). Their respective occurrence is indeed strongly correlated (-0.89 and 0.73) to the traditional wintertime NAO index estimated from EOF (Fig. 1e, f). Note that spatial asymmetries between the two NAO phases (longitudinal shift of the anomalous centres of action and associated storm track) are clearly evidenced here by clustering techniques that do not make any assumption for linearity. The year-to-year occurrence of the NAO regimes captures the strong interannual-to-interdecadal variability of the oscillation that has been extensively documented in literature (see e.g. Hurrell et al. 2003 for a review). The third regime is named Atlantic ridge (AR) and is reminiscent of the so-called East Atlantic pattern (negative phase, Barnston and Livezey 1987) viewed as a Euro-Atlantic wave train. The forth regime is often referred to as Scandinavian blocking (S-BL) and is characterized by a strong anomalous height anomaly over Northern Europe (Tyrlis and Hoskins 2008) while a mild deeper trough extends south-eastward from the Labrador Sea to the Iberian Peninsula. These two regimes all together are excited about 50% of time on average in winter and exhibit considerable interannual variability (Fig. 1g, h, less loading in the decadal frequency band compared to NAO regimes).

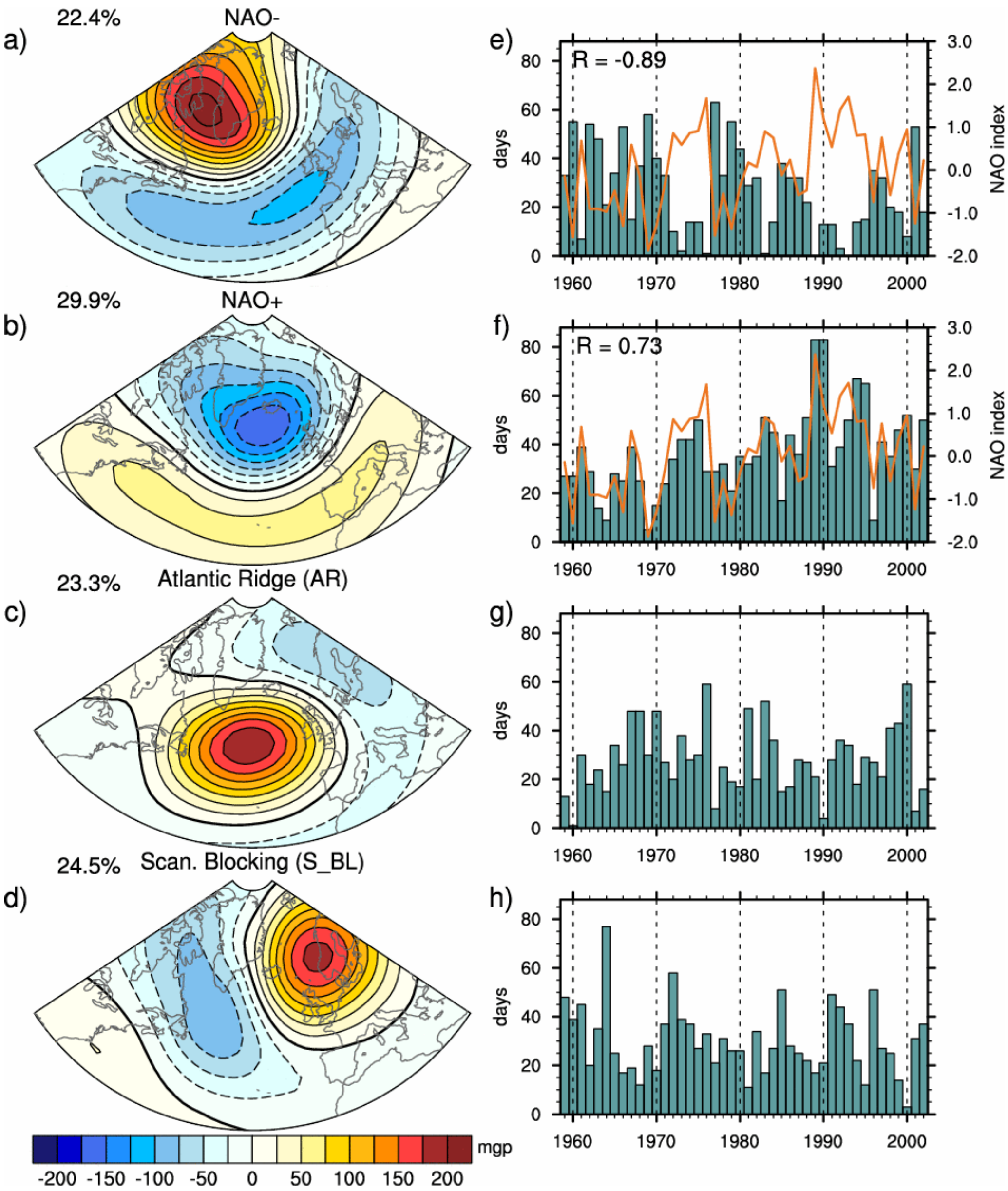

*Figure 1: **a–d** Centroids of the four wintertime NAE Z500 weather regimes (m). Each percentage represents the mean frequency occurrence of the regime computed over 1958–2002 from 1 December to 31 March. Contour intervals are 25 m.*

***e–h** Number of days of occurrence of each regime per winter from 1959 to 2002. The NAO index (orange curve) defined here by the normalized principal component of the leading EOF of averaged DJFM Z500 is superimposed on the upper two panels corresponding to the NAO regimes. Correlation (R) between the NAO index and the frequency of occurrence of the NAO regimes is provided. From Cassou et al (2010).*

Links between flow regimes and anomalous climate conditions over continental Europe have been extensively documented in literature from daily (e.g. Slonosky and Yiou 2001) to decadal timescales (Hurrell 1995). For instance, high pressure over Greenland (reinforced Icelandic Low) during NAO- (NAO+) leads to slackened and southward-shifted (reinforced and northward displaced) westerly winds affecting downstream temperature and precipitation over the entire Europe. A strong relationship between extremes and regimes occurrence also exists. There are various definitions for indices of extreme events; here we adopt the method of exceeding threshold obtained from percentiles. Daily anomalies are first calculated for each individual meteorological station from the ECA dataset over Europe (Klein et al 2002). The 95th percentiles (or 5% chance of occurrence) referred hereafter to as "climatological thresholds" are then calculated for each station and from 5-day windows centred on each calendar day over the full period. To assess the changes in the probability of extreme cold and wet day occurrence as a function of the four regimes (Fig.2), four new distributions are built by selecting only the days where a given regime is excited. New percentages of days that exceed the climatological thresholds are then computed for each individual station and for the four distributions. These are finally compared to the original climatological 5% probability of occurrence (Plaut and Simmonet 1994). Within such a framework, if the new percentage is 10% for instance for a given regime, it should be viewed as multiplying the likelihood for extreme cold or wet days to happen by 2.

In terms of temperature, NAO+ precludes any cold extremes over the entire Europe while NAO- regimes clearly favour cold outbreaks over a large northern domain. No significant changes in cold extremes can be tracked for AR, except over the Iberian Peninsula where their probability of occurrence is significantly increased. During S-BL, cold events are favoured in central Europe extending westward towards France. In terms of precipitation, NAO- increases chances of extreme rainfall events to occur over a large Western Europe with values as strong as 3 to 4 over most of the Iberian Peninsula. By contrast, precipitation extreme events are favoured for NAO+ regimes in north-western Europe while they are less likely in the Mediterranean basin. S-BL also precludes extreme rainfall to occur over a large portion of Europe. In addition to large-scale signals, Fig.2 shows that the regime decomposition also provides useful information about regional features especially for precipitation extremes (Spanish Mediterranean coast during S-BL, Pyrenees signature at the French-Spanish border etc.). Based on the observed link between weather regimes and temperature/precipitation extremes, we suggest that any potential predictive skill score for NAE-WR could indirectly provide a reliable probabilistic view for chances of extreme events to occur over the entire European continent as shown later on in this paper for winter 2010.

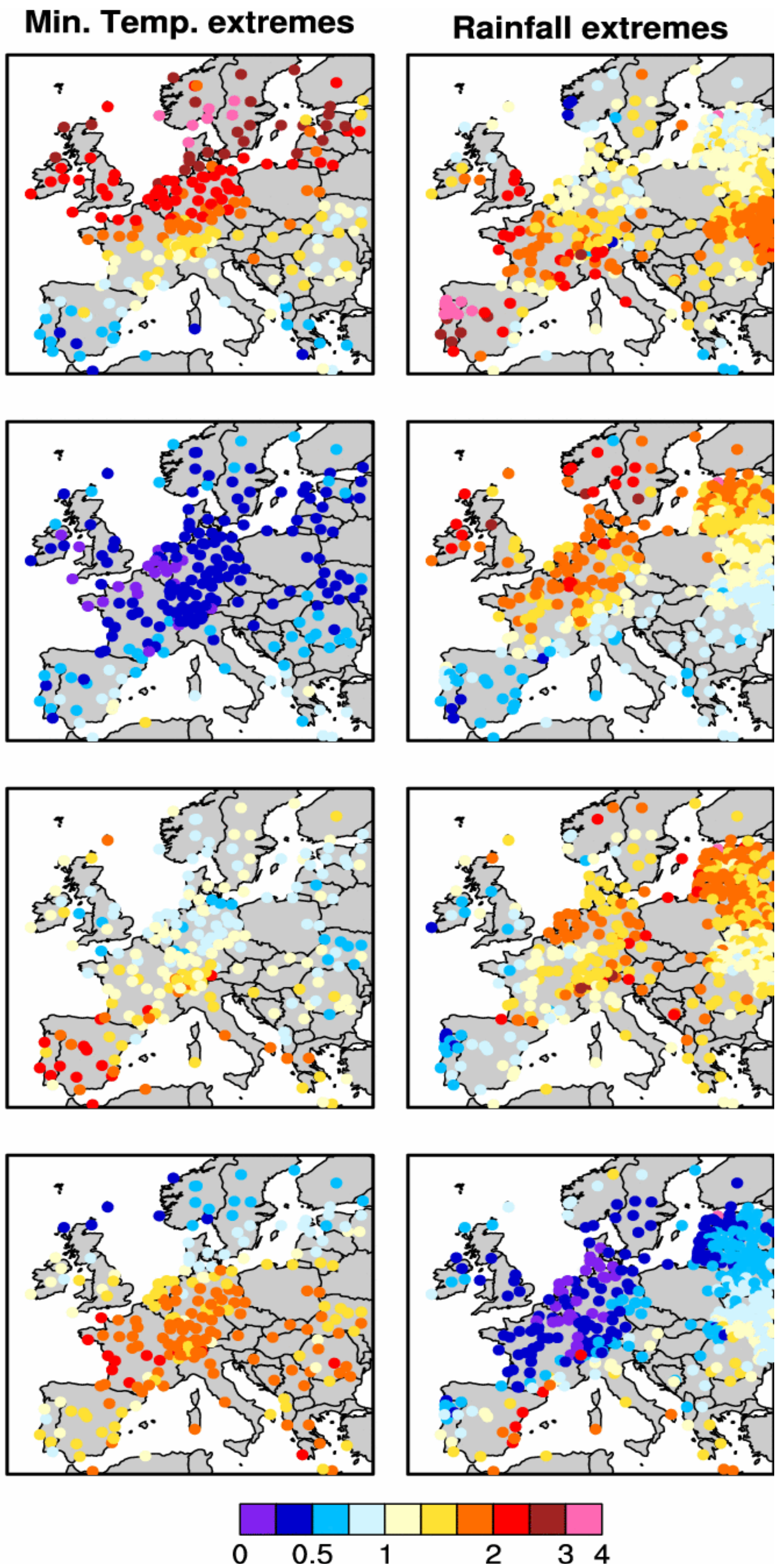

Figure 2: Relative changes for each individual regime in the frequency of extreme cold days (left maps) and wet days (right maps) defined by the 95th percentile for each station data from the ECA dataset. The value 1 means that the regime is not discriminative for extremes, while 0 shows that there is no chance of extreme to occur when the regime is excited, and 2 or 4 indicates that there is respectively double or quadruple chance of extreme to occur in association with the regime.

At low frequency, viewed as an efficient spatio-temporal filter of the mostly chaotic atmospheric flow at midlatitudes, the weather regime entity also appears to be compatible with the time-integrator properties of the ocean and is relevant to explain part of its long-term changes (Cassou et al 2010). In the latter reference, clustering has been additionally performed over the tropical Atlantic region to obtain surface wind classes (T-WC). Those are combined to the NAE-WRs above described to evaluate the ability of the daily classifications to capture the observed low-frequency variability of the surface ocean variables over the entire Atlantic. A least square multiple linear regression model is built: the December–March frequencies of occurrence of the circulation patterns (Fig. 1e-f) are used as predictors in the model while 10-meter wind (UV10) and 2-meter temperature (T2) are the predictants. Figure 3 compares the observed trends to the reconstructed ones based on this very simple regression model.

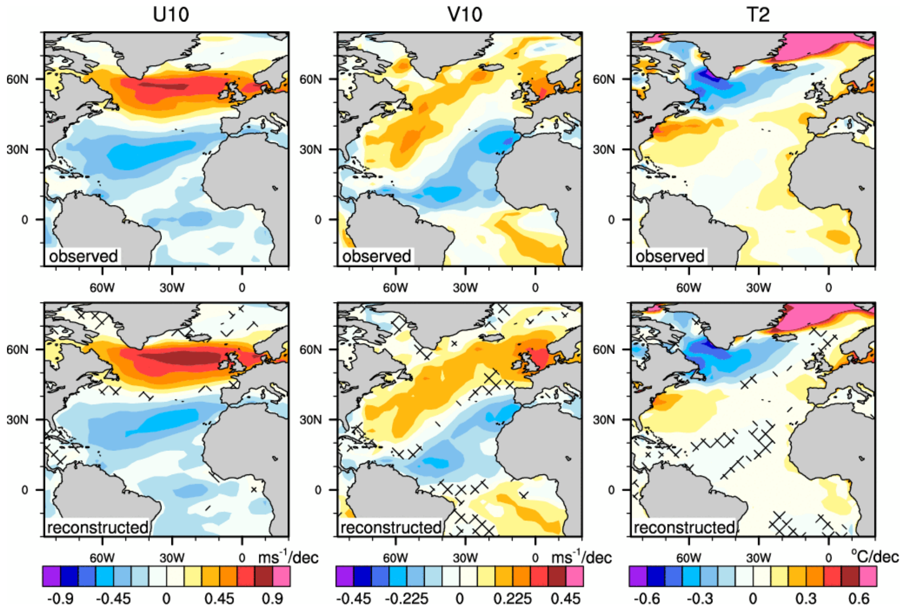

Figure 3: Linear trends computed over 1959–2002 for wintertime (upper panels) ERA40 U10, V10 and T2 and for (lower panels) reconstructed fields from multiple linear regression using NAE-WR+T-WC occurrences as predictors. Hashing stands for sign disagreement between observed and reconstructed trends. Contour intervals are 0.15 m s^{-1}/decade, 0.075 m s^{-1}/decade for UV10 and V10, respectively, and 0.1°C/decade for T2

Large-scale atmospheric circulation trends are characterized over 1959-2002 by an intensification of the midlatitude westerlies and a concomitant strengthening of the NE trades (Fig. 3 upper left) and easterlies along the equator. Northward displaced storm track is associated with prevalent southerlies north of 40°N whereas equatorward convergence associated with a southward retreat the Inter-Tropical-Convergence-Zone (ITCZ) is enhanced (Fig. 3 upper centre) due to reinforced trade winds in both hemispheres. Consistently, the largest trends for T2 occur in the western part of the North Atlantic and strongly project on the so-called North Atlantic tripole (Cassou et al 2004, Fig. 3 upper right). The pattern of trends reconstructed from multiple regression compares very well with observations (Fig. 3, lower panels). The pattern correlations are 0.98, 0.93, and 0.98 for U10, V10 and T2, respectively. Maximum discrepancies are found for T2 where the observed warming along the eastern side of the Atlantic is barely captured. This underestimated warming likely indicates that the T2 tendency there is additionally determined by influences other than the sole atmospheric dynamics; direct radiative contribution from increased greenhouse gazes concentration appears to be a reasonable candidate, as well as the Atlantic Multidecadal Oscillation (Knight et al 2005).

Based on this result, we conclude that observed trends in surface ocean fields can be viewed to a large extent as the temporal integration of the anomalous frequencies of occurrence of both NAE-WR and T-WC over the last 50 years or so. Similar results are found from intraseasonal to interannual timescales in agreement with the consensual idea that the atmosphere mostly imprints its variability at the surface ocean through anomalous surface turbulent fluxes and advection of air masses (Deser and Timlin 1997). We have shown so far that NAE-WRs are a significant driver of the variability over both the Atlantic Ocean and its adjacent continents. In the following, we will investigate how NAE-WR occurrences are affected through teleconnection mechanisms.

4. North Atlantic-Europe weather regimes and teleconnection

4.1. Intraseasonal timescale

A similar approach in weather-type classes has recently been applied in the tropics to describe and monitor in real time the dominant mode of intraseasonal climate variability, the Madden-Julian Oscillation (MJO, Madden and Julian 1994). The MJO is a natural component of the tropical coupled ocean–atmosphere system, and is characterized by a planetary-scale alternation of wet and dry periods associated with several changes in both tropical and subtropical atmospheric dynamics. The MJO packet propagates eastwards around the globe with a typical 30–70-day cycle. By combining real-time satellite outgoing long-wave radiation used as a proxy for convection and atmospheric dynamical fields from operational reanalysis, it is possible to partition the daily MJO activity into eight intrinsic phases of 7–8-day nominal persistence (Wheeler and Hendon 2004). These phases or classes can be interpreted as the tropical analogues of the extratropical weather regimes, except that regimes have episodic behaviour due to dominant chaos at midlatitudes whereas the time evolution of the MJO phases is mostly oscillatory.

Extratropical responses to MJO kicks have been extensively described in the literature (see Zhang 2005 for a review). In Cassou (2008), a novel approach more relevant to forecasting issues relies on the combination of both tropical and extratropical NAE clusters to investigate how the MJO influences the known and independent North Atlantic modes on medium-range timescales. Within this framework, evidence is presented that that the MJO controls part of the distribution and sequences of the four daily NAE-WR in winter. For the eight MJO phases and for lags of up to +15 days (MJO in advance), the number of occurrences of each NAE-WR is counted and compared to its mean excitation computed over the total sampling period (Fig. 4). Phase 3 of the MJO (enhanced convection over the Indian Ocean) is not discriminative for the NAO regimes at lag 0, whereas ~10 days later the probability of a NAO- event occurring is reduced by ~40% and is mostly compensated for by an increase of ~60% in the probability of NAO+. The opposite is found for phase 6, which shows an increase of NAO- occurrence probability building up to ~70% for lags greater than 10 days with NAO+ less probable. The AR regime seems to be less affected by the progress of the MJO, its occurrence being simply reduced by construction when NAO+ or NAO- regimes are dominant (phases 3–4 and 7, respectively). S-BL occurrence is also weakly altered, except at short lag time in phase 6 (enhanced excitation) before NAO- maturation. The changes in the regime distribution occur progressively in accordance with the nominal 7–8-day persistence of the eight MJO phases.

Fig. 4 constitutes a contingency table providing evidence that phases 3 and 6 of the MJO can be interpreted as precursors of NAO+ and NAO- regimes, respectively. The days following phase 3 are dominated by a mid-latitude low-frequency anomalous wave train that originates in the eastern Pacific, stretches across the North American continent and propagates eastwards following the Northern Hemispheric waveguide. Its penetration along the North Atlantic mean storm track (40^{o}–60^{o}N) is associated with dominant anticyclonic synoptic-scale wave breakings (AWBs) known as precursors for NAO+ (Benedict et al 2004, Franzke et al 2004) from a lag of ~6 days onwards. The opposite picture emerges for the days following phase 6 of the MJO. The proportion of AWBs is clearly reduced from a lag of +6 days onwards in that case. This reduction is almost entirely controlled by very high-frequency transients, which is consistent with the preferred in situ development of NAO- events associated with more frequent cyclonic wave breakings, and contrasts with NAO+ events, in which intermediate- frequency eddy activity has a role (Feldstein 2003) . For NAO-, there is no signal coming from the Pacific; height anomalies originating from Europe and propagating westwards are found instead from a lag of 0 to +6 days, before the development of a quasi-standing pattern projecting on NAO-.

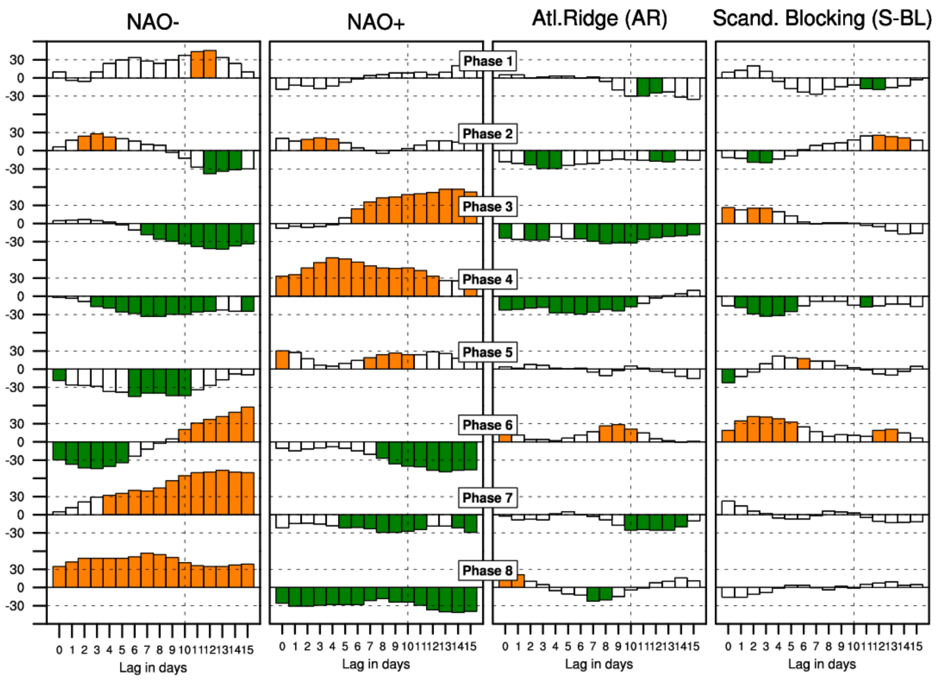

Figure 4: Lagged relationships between the eight phases of the MJO and the four North Atlantic weather regimes. Table of contingency between the MJO phases (rows) and the North Atlantic weather regimes (columns). For each MJO phase, I plot the anomalous percentage occurrence of a given regime as a function of lag in days (with regimes lagging MJO phases). The 0% value means that the MJO phase is not discriminative for the regime whose occurrence is climatological. A 100% value would mean that this regime occurs twice as frequently as its climatological mean; 2100% means no occurrence of this regime. The presence of a slope as a function of lag is suggestive of the MJO forcing. For white bars, either the change in the distribution between the four regimes is not significant on the basis of x2 statistics at the 99% significance level, or the individual anomalous frequency of occurrence is lower than the minimum significant threshold tested at 95% using a Gaussian distribution (approximation for binomial distribution because of the sufficiently large sampling). For orange and green bars, the regimes occur significantly more or, respectively, less frequently than their climatological occurrences.

Two non-exclusive mechanisms are proposed to explain the MJO/NAO- relationship. S-BL regimes are present at short lag time in phase 6 (Fig. 4) and may subsequently trigger the onset of NAO- events (e.g. Croci-Maspoli et al 2007). In this case, enhanced NAO- in late phase 6 and phase 7 would not be directly forced by the MJO, but would correspond to the timescale resonance between the eastward propagation of the MJO and the preferred sequence of the NAE-WR (NAO+ → S-BL → NAO-, Vautard 1990). The second mechanism proposed for teleconnection between MJO and NAO- relies on direct tropical forcings originating from the eastern Pacific. This picture is consistent with there being a Rossby wave source around 20^{o}N, 110^{o}W that initiates a downstream wave train propagating northeastwards towards Europe following a preferred curving path in line with several studies (e.g. Matthew et al 2004).

In summary, positive NAO events mostly respond to a mid-latitude low-frequency wave train initiated by the MJO in the western–central tropical Pacific and propagating eastwards following forced Rossby wave theories. Precursors for negative NAO events are found by contrast in the eastern tropical Pacific–western Atlantic, leading to changes along the North Atlantic storm track. Wave-breaking diagnostics tend to support the MJO preconditioning and the role of transient eddies in setting the phase of the NAO allowing for its medium-range predictability far exceeding the limit of around one week that is usually quoted.

4.2. Interannual to decadal timescale

While intrinsic or internal atmospheric variability exhibits temporal incoherence, the ocean tends to respond to it with marked persistence of heat content anomalies that potentially feedback to the local atmosphere. The level of retroaction of the anomalous extratropical sea surface temperature (SST) upon NAE-WR is however still under debate (Kushnir et al 2002 for a review). In any case, it appears to be weak except for specific domains (Nordic Seas) and through specific mechanisms such as the ocean re-emergence (Cassou et al 2007). Adding to the complexity of local ocean-atmosphere interaction is the existence of remote forcings of the NAE-WR from tropical oceans; the latter influence appears to be more robust from interannual to decadal timescale.

Several studies have concluded that the NAO variability is closely tied to SST variations in the tropical Atlantic basin. We have mentioned that the NAO weather regimes are linked to the North Atlantic SST tripole. Despite the latter is consistent with the atmosphere acting as a forcing for the surface ocean (e.g. Deser and Timlin 1997), a series of model experiments have shown that the tropical part of the ocean tripole does feed back to the NAO (e.g. Sutton et al 2001). Anomalous interhemispheric SST gradient alters the strength and location of the tropical convection along the ITCZ and thus ultimately modulates the NAE circulation via the excitation of Rossby waves propagating northeastward from the anomalous diabatic heating source (Terray and Cassou 2002). In winter, the latter is located over the northern part of South America. In Cassou et al (2004), results from the ARPEGE atmospheric global circulation model (AGCM) confirms that the excitation of NAO+ (NAO-) regimes are favoured when the North tropical Atlantic basin is cold (warm) with some asymmetry in the teleconnection (Figure 5).

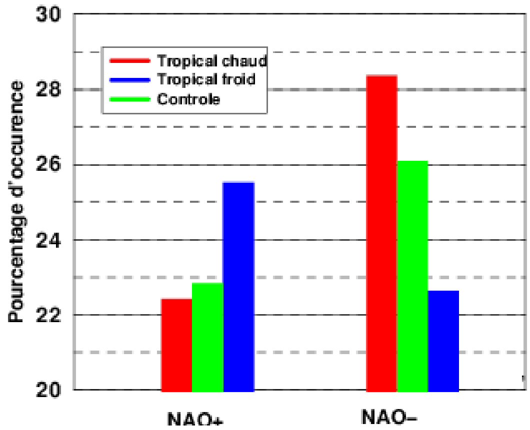

Figure 5: Mean percentage of occurrence of wintertime NAO weather regimes in three 30-member ensembles of the ARPEGE AGCM where +1std SST anomalies (red) and -1std SST anomalies (blue) are imposed in the tropical Atlantic (0-25°N) and compared to a climatological run (green)

At interannual timescale, the impact of the El Nino Southern Oscillation (ENSO) on NAE-WR appears to be present but remains open to debate. The relationships might be asymmetrical with respect to the ENSO phases. Cold La Nina events seem to favour both NAO+ and AR regimes and to trigger more systematic connections than warm El Nino events (e.g. Hannashi 2001). AR seems to be linked during La Nina to the tropospheric Pacific–North America (PNA) arching pattern extending toward the western Atlantic (e.g. Straus and Shukla 2002). Recently some papers have reported that the links between ENSO and NAO weather regimes might be indirect via the ENSO remote impact on tropical North Atlantic SST or via the ENSO influence on the stratospheric polar vortex leading downward in the troposphere to changes in the preferred phase of the NAO (e.g. Bell et al 2009). Most of these mechanisms that have been suggested from the observations or their estimation via the reanalysis are confirmed through model experiments and through a series of more or less sophisticated statistical techniques to optimally extract teleconnections (e.g. Venske et al 1999 using fingerprint methods).

5. Euro-Atlantic regimes and seasonal forecast: the winter 2009/2010 case study

The winter of 2009/2010 was characterized by a record negative value of the NAO index since 1824, which caused several severe cold spells over Northern and Western Europe (see Fig.2). This somehow unusual winter with respect to the most recent ones arose concurrently with public debate on climate change, during and after the Copenhagen climate negotiations. It was a big buzz in the European media giving some echoes to climate sceptics' voices that, more or less deliberately, pass over the concept of natural variability. Winter 2010 extreme NAO index is associated with both a record frequency of NAO− occurrences over winters 1958–2010 (63 days out of 90 days from Dec. the 1st to March the 31st, 8 more days than the previous record in winter 1966) and a very low frequency of NAO+ occurrences (5/90 days, the 3rd lowest after 1969, 3/90, and 1963, 4/90, Cattiaux et al 2010). Winter 2010 was characterized by an exceptional Northern Hemisphere mean atmospheric circulation (Wang et al. 2010); the Z500 anomaly exhibits a strong zonal hemispheric pattern, with anomalously high (low) pressures over the pole (mid-latitudes) as high as 3 standard deviations. Those anomalies project onto the canonical negative phase of the Arctic Oscillation (AO) (Thompson and Wallace 1998) that was measured at a record level (L'Heureux et al. 2010). The legitimate question to ask is the following one: has this atypical winter and associated cold spells over Europe been correctly predicted by seasonal forecast systems?

The tercile summary of T2 anomalies forecast initialized on December 1st is given at leadtime +2 months (averaged January-February-March –JFM conditions) for ECMWF (Fig.6, left). In association with the ongoing warm El Nino event at the end of 2009, above upper-tercile T2 conditions were expected with very high confidence within the tropical band. Warmer conditions were also forecast over a large northernmost part of the North American continent extending over the ocean towards the Labrador and Irminger Seas. Lower but still significant probability for warmer conditions over Europe was also expected. Forecasts were very consistent between the three models (ECMWF, UKMO and Météo-France) of the EUROSIP multi-model seasonal forecast project.

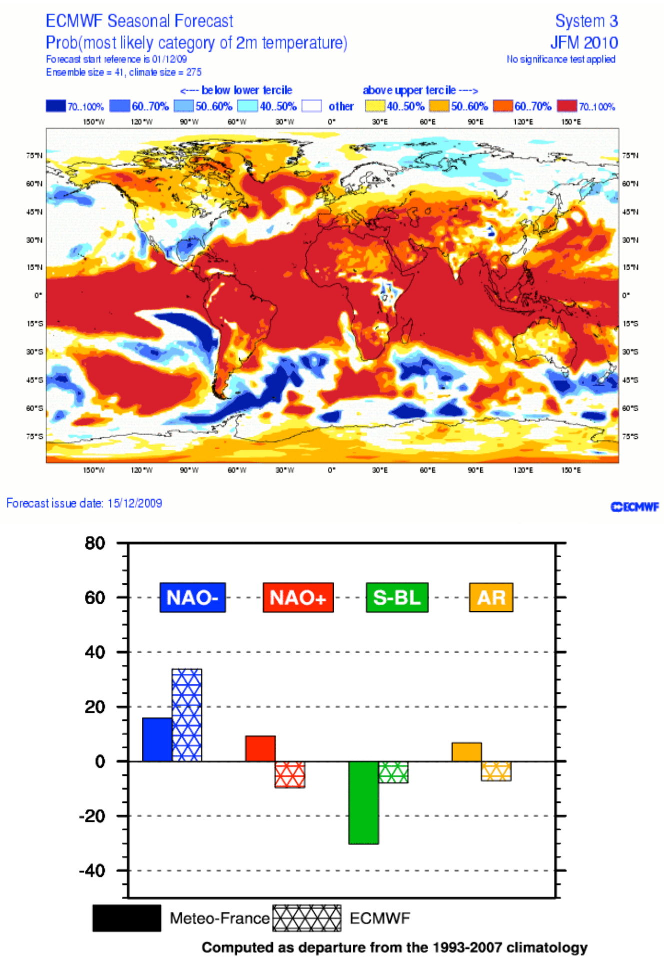

Figure 6: (left) Probability of the most likely category for 2-meter temperature (T2) anomalies predicted for JFM from ECMWF forecast initialized in December 1st. (right) Anomalous percentage of occurrence of the four NAE-WR predicted for JFM from Météo-France and ECMWF forecast systems. 0% for a given regime means that climatological occurrence is predicted while 100% (-100%) indicates that there is double (no) chance for that regime to occur.

Classification of the daily anomalous Z500 forecasts into the four NAE-WR (Fig. 1) shows that NAO- regimes are predicted to occur more frequently than its climatological conditions in JFM 2010 in both Météo-France and ECMWF systems (Fig.6 right). This excess is compensated by a deficit of S-BL occurrence in Météo-France and a deficit of the three other regimes in ECMWF. The latter forecast is very successful when confronted to the observations (Cattiaux et al 2010). The NAO- predicted surplus occurrence (around +35%) corresponds to ~1.5 standard deviation anomalies calculated over 1979-2007 (reference period for models prediction). Without being extreme albeit strong, the predicted dominance of NAO- regime for winter 2010 is explained in the operational models by the combined forcing role of ENSO and warm tropical Atlantic.

Skill is thus found in the prediction of the large-scale atmospheric circulation anomalies for winter 2010 (Fig.6, right) whereas all the models fail to predict the observed cold conditions at the surface over most of Europe (Fig.6, left). This case study suggests that using raw forecasts for impact variables (temperature, precipitation, etc.) for which skills are rather limited, might be “dangerous” and might produce erroneous information for users. Skills appear to be higher in forecasting large-scale free atmospheric circulation patterns from which surface conditions could be assessed indirectly through a transfer function built from the observations only. In winter 2010, if NAE-WR occurrences had been considered for the forecast of the temperature anomalies over Europe, emphasis would have been laid on the enhanced probability for cold waves to occur over northwestern Europe due to predicted strong favoured excitation of the NAO- regime. Instead, the “mild winter conditions” scenario that was incidentally suggested by the raw surface temperature forecast was preferred for operational use. This example highlights the added value of the weather regime paradigm to bridge the gap between weather and climate variability not only for the understanding of the low-frequency atmospheric changes but also for forecasting issues.

References :

Barnston A.G. and Livezey R.E (1987): Classification, seasonality and persistence of low-frequency atmospheric circulation patterns. *Mon. Wea. Rev.*, **115**, 1083–1126.

Bell C.J., Gray L.J., Charlton-Perez A.J., Joshi M.M. and Scaife A.A. (2009): Stratospheric communication of ENSO teleconnections to European Winter. *J. Clim*., **22**, 4083-4096.

Benedict J.J. ,Lee S. and Feldstein S.B. (2004): Synoptic view of the North Atlantic Oscillation. *J. Atmos. Sci.,* **61**, 121–144.

Cassou C., L. Terray, J.W. Hurrell and C. Deser (2004): North Atlantic winter climate regimes: spatial asymmetry, stationarity with time and oceanic forcing. *J. Clim*., **17**, 1055-1068.

Cassou C., C. Deser and M. A. Alexander (2007): Investigating the impact of reemerging sea surface temperature anomalies on the winter atmospheric circulation over the North Atlantic. *J. Clim*., **20**, 3510-3526.

Cassou C. (2008): Intraseasonal interaction between the Madden and Julian Oscillation and the North Atlantic Oscillation. *Nature*, **455**, doi:10.1038/nature07286

Cassou C., M. Minvielle, L. Terray and C. Périgaud (2010): A statistical-dynamical scheme for reconstructing ocean forcing in the Atlantic. Part I: weather regimes as predictors for ocean surface variables. *Clim. Dyn.,* doi:10.1007/s00382-010-0781-7.

Cattiaux J., R. Vautard, C. Cassou, P. Yiou, V. Masson-Delmotte, and F. Codron (2010): Winter 2010 in Europe: A cold extreme in a warming climate, *Geophys. Res. Lett.,* **37**, L20704, DOI:10.1029/2010GLO44613.

Cheng X.H. and Wallace J.M. (1993): Cluster analysis of the northern hemisphere winter 500-hPa height field: spatial patterns. *J. Atmos. Sci.,* **50**, 2674–2696.

Christiansen B. (2007): Atmospheric circulation regimes: can cluster analysis provide the number? *J. Clim.,* **20**, 2229–2250.

Croci-Maspoli M., Schwierz C. and Davies H. (2007): Atmospheric blocking: space-timelinks to the NAO and PNA. *Clim. Dyn*., **29**, 713–725.

Deser C. and M.S. Timlin (1997): Atmosphere-ocean interaction on weekly timescales in the North Atlantic and Pacific. *J. Clim*., **22**, 396-413.

Feldstein S.B. (2003): The dynamics of NAO teleconnection pattern growth and decay. *Q. J. R. Meteorol. Soc.,* **129**, 901–924.

Franzke C., Lee, S. and Feldstein, S. B. (2004): Is the North Atlantic Oscillation a breaking wave? *J. Atmos. Sci.,* **61**, 145–160.

Franzke C. and Feldstein S.B. (2005): The continuum and dynamics of Northern hemisphere teleconnection patterns. *J. Atmos. Sci.,* **62**, 3250–3267.

Ghil M. and Roberston A.W. (2002): 'Waves' vs 'Particles' in the atmospheric phase space: a pathway to long-range forecasting? *Proc Natl Acad Sci USA*, **99**, 2493–2500.

Hannachi A. (2001): Toward a Nonlinear Identification of the Atmospheric Response to ENSO. *J. Clim.,* **14**, 2138-2149.

Hannachi A. (2010): On the Origin of Planetary-Scale Extratropical Winter Circulation Regimes. *J. Atmos. Sci.,* DOI: 10.1175/2009JAS3296.1.

Hurrell JW (1995): Decadal trends in the North Atlantic Oscillation: regional temperature and precipitation. *Science,* **26**, 676–679.

Hurrell W.J., Kushnir Y., Otterson G. and M. Visbeck (2003): An overview of the North Atlantic Oscillation. *AGU Geophys. Mono*, **134**, doi:10.1029/134GM01.

Johnson N.C., Feldstein S.B., Tremblay B. (2008): The continuum of Northern Hemisphere teleconnection patterns and a description of the NAO shift with the use of self-organized maps. *J. Clim.,* **21**, 6354–6371.

Klein-Tank, A. M. G. and Coauthors (2002). Daily dataset of the 20th century surface air temperature and precipitation series for the European Climate Assessment. *Int. J. Climatol.*, **22**, 1441-1453.

Knight J.R., Allan R.J., Folland C.K., Vellinga M., Mann M.E. (2005): A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophys. Res. Lett.,* **32**, L20708. doi:10.1029/ 2005GL024233

Kushnir Y., W. A. Robinson, I. Bladé, N. M. J. Hall, S. Peng, and R. T. Sutton (2002): Atmospheric GCM response to extratropical SST anomalies: Synthesis and evaluation. *J. Clim.,* **15**, 2233–2256.

L'Heureux M., A. Butler, B. Jha, A. Kumar, and W. Wang (2010): Unusual extremes in the negative phase of the Arctic Oscillation during 2009, *Geophys. Res. Lett.,* **37**, L10704, doi:10.1029/2010GL043338.

Madden, R. A. and Julian, P. R. (1994): Observations of the 40–50 day tropical oscillation. *Mon. Wea. Rev.,* **112**, 1109–1123.

Matthews A. J., Hoskins B. J. and Masutani M. (2004): The global response to tropical heating in the Madden-Julian Oscillation during northern winter. *Q. J. R. Meteorol. Soc.,* **130**, 1991–2011.

Michelangeli P., Vautard R. and Legras B. (1995): Weather regimes: recurrence and quasi-stationarity. *J. Atmos. Sci.,* **52**, 1237–1256.

Molteni F., Kuscharski F. and Corti S. (2006) in *Predictability of Weather and Climate* (eds Palmer, T. & Hagedorn, R.) 365–389 (Cambridge Univ. Press).

Plaut G. and Simonnet E. (2001): Large scale circulation classification, weather regimes and local climate over France, the Alps and western Europe. *Climate Res.* **17**, 303- 324.

Slonosky V.C. and Yiou P. (2001): The North Atlantic Oscillation and its relationship with near surface temperature. *Geophys. Res. Lett.,* **28**, 807–810.

Stephenson, D. B., Hannachi, A., & O'Neill, A. (2004). On the existence of multiple climate regimes. *Quart. J. Roy. Meteor. Soc.,* **130**, 583-606 .

Straus D., Corti, S. and Molteni, F. (2007): Circulation regimes: chaotic variability versus SST-forced predictability. *J. Clim.,* **20**, 2251–2272.

Sutton R. T., W. A. Norton and S. P. Jewson (2001): The North Atlantic Oscillation—What role for the ocean? *Atmos. Sci. Lett.,* **1**, 89–100.

Terray L. and C. Cassou (2002): Tropical Atlantic sea surface temperature forcing of the quasi-decadal climate variability over the North Atlantic–Europe region. *J. Clim.,* **15**, 3170–3187.

Thompson D., and J. Wallace (1998): The Arctic Oscillation signature in the wintertime geopotential height and temperature fields, *Geophys. Res. Lett.,* **25**(9), 1297–1300.

Tyrlis E. and Hoskins B.J. (2008): Aspects of a Northern Hemisphere atmospheric blocking climatology. *J. Atmos. Sci.,* **65**, 1638–1652.

Straus D. and Shukla J. (2002): Does ENSO for the PNA? *J. Clim.,* **15**, 2340-2358.

Vautard R. (1990): Multiple weather regimes over the North Atlantic: Analysis of precursors and successors. *Mon. Wea. Rev.,* **118**, 2056–2081.

Venzke S., M. R. Allen, R. T. Sutton, and D. P. Rowell (1999): The atmospheric response over the North Atlantic to decadal changes in sea surface temperature. *J. Clim.,* **12**, 2562–2584.

Wang C., H. Liu and S. Lee (2010): The record-breaking cold temperatures during the winter of 2009/2010 in the Northern Hemisphere. *Atmos. Sci. Lett.,* doi:10.1002/asl.278.

Wheeler M. C. and Hendon H. H. (2004): An all-season real-time multivariate MJO index: development of an index for monitoring and prediction. *Mon. Wea. Rev.,* **132**, 1917–1932.

Wilks D. S. (1995): *Statistical methods in the atmospheric sciences*. Academic Press, pp 476.

Zhang C. (2005): Madden-Julian oscillation. *Rev. Geophys*., **43**, doi:10.1029/2004RG000158