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ASL Atmospheric Science Letters

Atmospheric Science Letters 5 (2004) 43–55

www.elsevier.com/locate/issn/1530261X

Precipitation and Northern Hemisphere regimes

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Received 10 July 2003; revised 6 November 2003; accepted 2 December 2003

Abstract

Rainfall anomalies in a longterm integration of general circulation model highlight the non-stationarity of the ocean–atmosphere coupling in the North Atlantic which becomes manifest in two regimes. Anti-correlations between the precipitation in the tropical and subtropical western Atlantic illustrate the changes of the Hadley cell with El Niño/Southern Oscillation (ENSO).The precipitation anomaly pattern in the north eastern Atlantic resembles variations of the North Atlantic storm track and the North Atlantic Oscillation (NAO). In the hemispheric regime, where 40% of the NAO variability can be explained by ENSO, both precipitation pattern are connected, whereas in the regional regime the ENSO-link with the North Atlantic storm track and the subtropical 500 hPa geopotential height disappears.

Keywords: Circulation regimes; Decadal variability; El Niño/Southern Oscillation–North Atlantic Oscillation connection; Precipitation

1. Introduction

The North Atlantic Oscillation (NAO) is the most prominent mode in the North Atlantic and affects the European climate and its variability (Hurrell, 1995, and references therein). During the last century its temporal behaviour covers a wide spectral band with a distinct low-frequency contribution. The analysis of proxy data of the NAO shows phases of enhanced (active) and reduced (passive) decadal variability (Appenzeller et al., 1998). Different hypotheses have been put forward to explain these variations ranging from external parameters such as volcanoes (Graf, 1994) or solar activity, via internal

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1530-261X/$ - see front matter
doi:10.1016/j.atmoscilet.2003.12.001

wave-wave interaction in the atmosphere (James and James, 1989; James et al., 1994; Franzke et al., 2000) to stratosphere–troposphere coupling (Perlwitz and Graf, 2001). In addition, ocean–atmosphere coupling and its impact are discussed in several observational and model studies, which yield often contradicting and confusing results (Latif, 1998; Wanner et al., 2001). The North Atlantic Oscillation has a hemispheric counterpart named Northern Hemispheric annular mode (NAM) or Arctic Oscillation (AO). The NAO paradigm is inherently sectoral (or regional) whereas the AO–NAM paradigm is inherently annular (or hemispheric; Wallace and Thompson, 2002b).

The sectoral NAO paradigm, here named *regional* mode, is supported by Bjerknes (1964). In his concept the atmosphere leads the ocean on interannual time scales, whereas for longer time scales the ocean dynamics increase in importance. More recent investigations conclude that local atmospheric forcing of winter sea surface temperature (SST) anomalies can also be important for longer periods (Deser and Blackmon, 1993; Halliwell, 1997). Using general circulation models (GCMs) a coupled ocean–atmosphere mode is identified in the North Atlantic where the subpolar gyre reacts on wind field variations and the ocean is able to feedback heat to the atmosphere (Grötzner et al., 1998). Moreover, analysing the NCEP-NCAR reanalysis, Czaja and Frankignoul (2002) identified a positive feedback between Atlantic SST and NAO. The response in GCMs to extratropical SST anomalies depends strongly on the characteristics of the model's intrinsic variability (Peng et al., 2002). Near the SST anomaly the nature of the response is predominantly determined by anomalous transient-eddy vorticity fluxes which have their origin from interactions with the storm track (Peng et al., 2002; Walter et al., 2001).

The annular AO–NAM paradigm, here named *hemispheric* mode, is encouraged by other observational studies which emphasise the importance of tropical variations for the European climate (Rowntree, 1972; Fraedrich, 1994). These authors propose an ENSO–Europe link which has been substantiated by model sensitivity experiments using an idealized SST anomaly pattern (Merkel and Latif, 2002). High-pressure phases over northern Europe are connected to El Niño conditions (Fraedrich and Müller, 1992). Increasing upper-tropospheric relative humidity in the tropical Pacific flanked by decreasing humidity further north reflects the enhanced strength of Hadley cell over the eastern Pacific during El Niño (Klein et al., 1999). Comparable considerations show a weakening of the Hadley cell over the Atlantic.

The intensity of decadal NAO variability (Appenzeller et al., 1998) and the connection between NAO and the SST (Raible et al., 2001; Walter and Graf, 2002) is not the same for all decades. Moreover, the atmospheric centres of action are time dependent: the teleconnection patterns for the period 1949–1961 derived from observations are shifted westwards in comparison to the next 15 years (Wallace and Gutzler, 1981). Using NCEP-NCAR reanalysis data from 1958 to 1999 Wallace and Thompson (2000b) identified a PNA-like mode as the second Empirical Orthogonal Function in the 500 hPa geopotential height which extends into the Atlantic. Note that this PNA–NAO connection can also be present in an active phase, but in this case it is of second order as classified by the decadal NAO variability. In this letter precipitation anomalies illustrate a sectoral behaviour in the regional regime and an annular behaviour in the hemispheric regime.

The outline is as follows: after introducing the coupled GCM simulation and the analysis techniques (Section 2) the two decadal regimes and the underlying processes within each regime are described separately (Section 3). Finally, Section 4 merges the results to two concepts of distinct signatures for both regimes and proposes possible processes which can explain the regime shift.

2. Coupled GCM simulation and analysis techniques

To obtain the longterm coupled GCM simulation an atmosphere and an ocean model are used. The atmospheric component is the fourth version of the *E*uropean *C*entre model of *Ham*burg (ECHAM-4) using triangular truncation at wavenumber 30 (T30, corresponding to a longitude–latitude grid of $3.75° \times 3.75°$) and 19 hybrid sigma-pressure levels up to 10 hPa. The ocean is the *H*amburg *o*cean model in *p*rimitive *e*quations (HOPE) simplified by the Boussinesq approximation and formulated on a Gaussian T42 Arakawa-E grid and 20 irregularly distributed levels. Both models are connected via Ocean Atmosphere Sea Ice Soil coupler (Terray et al., 1998). To obtain a stable mean climate an annual mean flux correction scheme for heat and fresh water fluxes is added. A 600-yr simulation for present day climate conditions is carried out. A detailed description of the experimental setup can be found in Legutke and Voss (1999).

The focus of the model analysis is the winter season December to February (DJF). The simulated NAO index is defined as the normalized 500 hPa geopotential difference averaged over four grid points near the Azores and Iceland. A simulated ENSO index is calculated from the SST in the Niño3 region. The spectrum for each 30-yr window of the 600-yr NAO time series is deduced using the maximum entropy method. The centred variance (deviation from the long term mean) on the 5–30-yr spectral band characterizes phases with enhanced (active) or reduced (passive) decadal variability. An active (passive) phase is a 30-yr segment with decadal variability above (below) one standard deviation. An active (passive) phase with a sectoral (annular) spatial scale in the teleconnection pattern is named a regional (hemispheric) decadal regime. Note that not all model 30-yr segments can be assigned as active or passive. The teleconnection is defined as the strongest negative correlation of the 500 hPa geopotential at one base point will all other grid points (Wallace and Gutzler, 1981). Only strong negative correlations which are clustered together in large areas are considered as 'centres of action'.

Standard composite (threshold one standard deviation) and correlation analyses are applied to (active or passive) classified segments using the associated teleconnection indices NAO and ENSO. A standard student *t*-test is applied to the composite analysis, but it is not displayed in Figs. 1–3, because the significant areas of the composite analysis are very similar to the region of significant correlations coefficients (significance level 95%, above $+0.3$ and below -0.3).

3. Precipitation and ocean–atmosphere coupling

Differences in precipitation anomalies are discussed in the context of the decadal regimes (Section 2). The hemispheric regime occurs more frequently than the regional regime in the coupled GCM simulation (150 versus 90 years of 600 years). The question

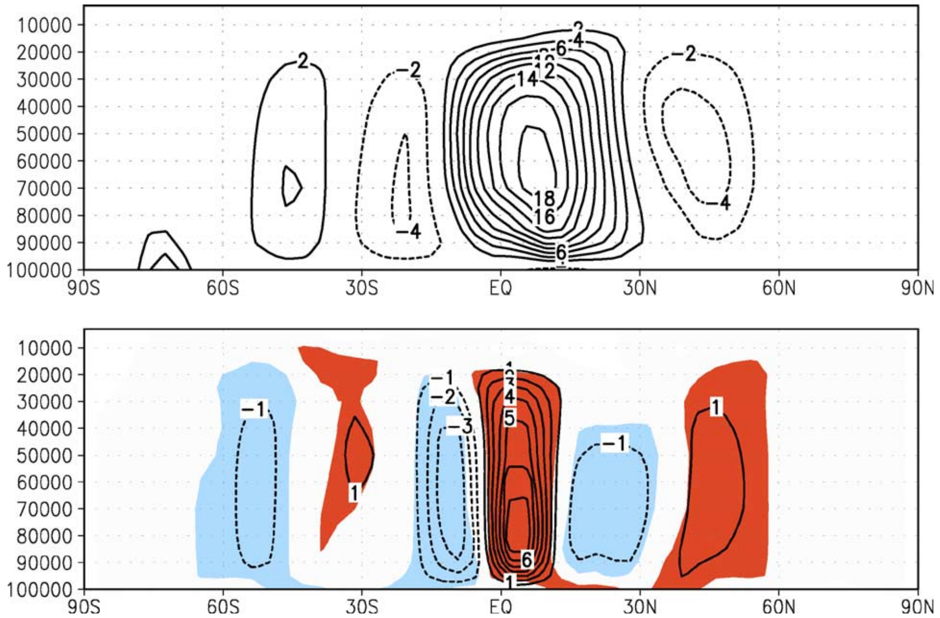

Fig. 1. Meridional mass streamfunction in the hemispheric regime: top panel shows the mean over the passive phase (model year 135–164, unit: 10^{10} kg/s). Bottom panel shows as contours the results of the composite analysis (threshold: one standard deviation): displayed is the difference between the mean meridional mass streamfunction for all positive and the mean for all negative ENSO phases, respectively. Moreover, the shading in the bottom panel shows correlation coefficients ≤ -0.3 (blue) and ≥ 0.3 (red).

arises which mechanisms determine these characteristic spatial and temporal scales of the atmospheric circulation in the Northern Hemisphere. In the following the results of two 30-yr phases of the coupled GCM simulation (active phase: model year 135–164; passive phase: model year 182–211) are presented to demonstrate the characteristics of the two decadal regimes.

3.1. The hemispheric regime

The hemispheric regime is characterized by reduced low-frequency NAO variability and a teleconnection pattern which shows centres of action in the Atlantic and the Pacific (Raible et al., 2001). The Aleutian centre of the Pacific-North American (PNA) pattern is strongly connected with the whole tropics; the Florida centre extends into the Atlantic and merges with the southern centre of the NAO.

The meridional mass streamfunction[1] shows that the Hadley circulation is enhanced and the south for positive versus negative ENSO situations (Fig. 1). The subtropical jet (not displayed) is intensified and shifted equatorward resembling observations (Trenberth et al., 1998). Note that this region is a source of Rossby waves propagating in a wave train

[1] The meridional mass streamfunction is calculated as a zonal average from 0 to 360° longitude.

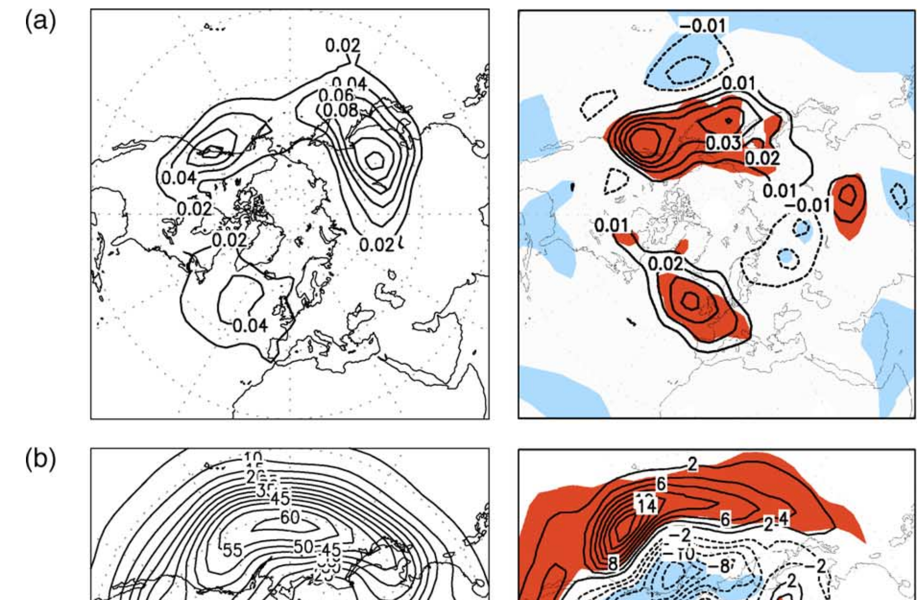

Fig. 2. Wave activity in the hemispheric regime: as Fig. 1, but for (a) the vertical component of the wave activity flux vector $\mathbf{F}_s$ in 500 hPa (Plumb, 1985; unit: m^2/s^2) (b) the 2.5–6 days bandpass filtered standard deviation of the 500 hPa geopotential height field (unit: gpm), (c) precipitation (unit: mm/day) using the ENSO index. Left panels show the mean, right panel the composites and correlation coefficients.

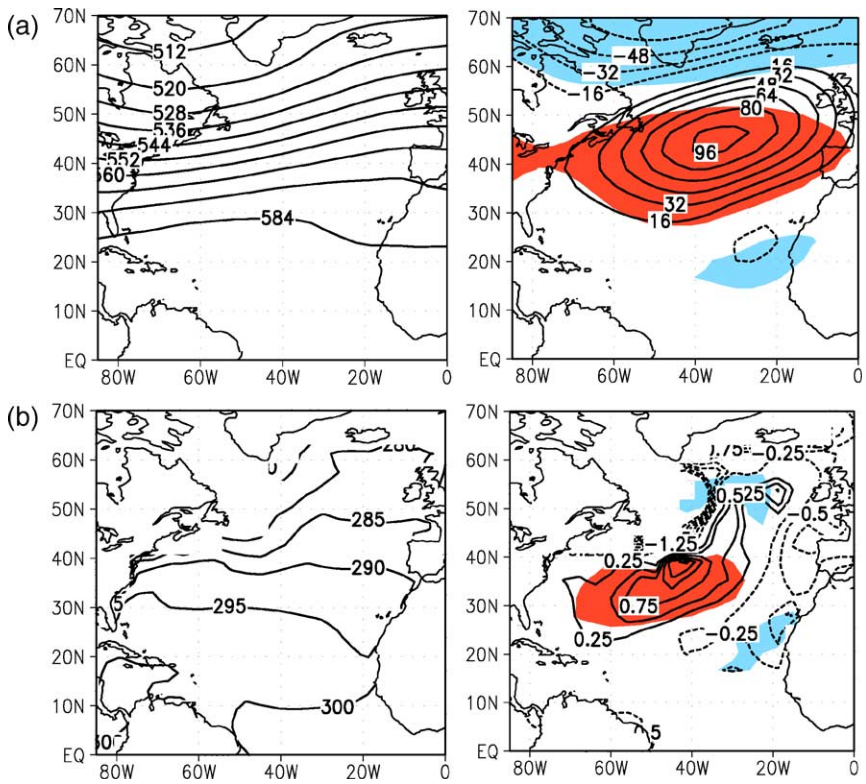

Fig. 3. Atmosphere–ocean interaction in the regional regime: as Fig. 2, but for (a) 500 hPa geopotential height (unit: gpm), (b) SST (unit: K) in the active phase (model year 182–211) using the NAO index.

to the northeast (Horel and Wallace, 1981; Hoskins and Karoly, 1981). These stationary waves are characterized by two measures: the 250 hPa streamfunction with its zonal mean removed and the vertical component of the wave activity flux vector (Plumb, 1985). The 250 hPa streamfunction shows a positive PNA-like pattern in the composite analysis which connects the Pacific with the Atlantic (not shown). The vertical component of the wave activity flux vector $\mathbf{F}_s$ in 500 hPa illustrates sources and sinks of stationary wave activity (Fig. 2a). The source near the Aleutians and in the North Atlantic is strongly enhanced, indicating that a more meridional wind appears during El Niño than during La Niña phases. The synoptic variability is also influenced by El Niño phases; more cyclones intensify the Aleutian Low and weaken the Azores High in the coupled simulation (Merkel and Latif, 2002; Raible and Blender, 2004). The classical measure of the storm track, the bandpass filtered standard deviation of the 500 hPa geopotential height (Fig. 2b), is shifted southwards in both the Pacific and the Atlantic. This southward shift at the exit of the Atlantic storm track is associated with the negative phase of the NAO. Both the stationary and transient wave behaviour resembles the observation where high pressure over northern Europe is associated with enhanced stationary eddy activity and a stronger jet maximum over the east coast of North America; the strongest activity of the transient eddies over the North Pacific extends about 10–20° further south-eastward (Fraedrich et al., 1993). High pressure phases over northern Europe itself appear to be connected to El Niño conditions (Fraedrich and Müller, 1992). Moreover, the local correlations between El Niño and precipitation (Fig. 2c) shows increasing (decreasing) rainfall in the Atlantic near 30°N (equator). A precipitation index based on this dipole pattern in the tropical Atlantic (defined as standardised difference of precipitation in the area 25–35°N, 80–30°W and the area 0–10°N, 60–40°W) is highly correlated with ENSO ($r = 0.7$). This is consistent to findings of Klein et al. (1999) showing that the strength of the Hadley cell indicated by upper tropospheric humidity and cloud cover over the eastern Pacific (Atlantic) is enhanced (reduced) during El Niño. The Azores High is anomalously weak during El Niño; 40% of the subtropical 500 hPa geopotential height in the Atlantic can be explained by ENSO (see Fig. 7d in Raible et al., 2001). In the hemispheric regime also the teleconnection indices NAO and ENSO are significantly correlated ($r = -0.61$ for 30 winters).

For ENSO cold events (La Niña, not shown) the storm tracks and the jet stream shifted poleward in the simulation which agrees with observational findings of Trenberth et al. (1998) and the NAO index is positive. Thus, ENSO triggers the atmospheric variability and the hemispheric mode appears as a slave of the ocean–atmosphere coupling in the tropical Pacific.

3.2. The regional regime

The regional regime is identified in phases of enhanced low-frequency NAO variability, with two sectorally independent teleconnection patterns. Teleconnection maps show centres of action in the North Atlantic and the North Pacific: a baroclinic PNA and an equivalent-barotropic NAO pattern. NAO is not correlated with ENSO or PNA. Half of the 500 hPa geopotential height variance can be explained by the SST anomalies near 30°N in the Atlantic. In the following the maintenance of a NAO sign is illustrated by two steps: (i) the oceanic response to atmospheric circulation anomalies and (ii) the atmospheric response to the wind induced SST anomalies.

As the atmosphere leads the ocean on interannual time scales (Bjerknes, 1964), we can start with a positive NAO index (the pressure difference between Iceland Low and Azores High is enhanced) and discuss the oceanic response in a first step: anomalously strong advection of cold air (due to the geostrophic wind component of the Iceland Low) cools the ocean upstream of the Iceland Low by latent and sensible heat fluxes. This cooling is increased by anomalous Ekman transport from the north (Luksch, 1996). Similarly, the enhanced Azores High forces by advection a warm SST anomaly upstream (westward to its centre) and a cold one downstream (eastward to its centre). This leads to a SST tripole as described by Wu and Gordon (2002) which is demonstrated within the model simulation by composite/correlation analysis using the NAO index (Fig. 3a and b). The 500 hPa geopotential height shows the NAO pattern which is strongly connected with the SST tripole; the correlation coefficient is highly significant ($r = 0.77$) when applying a canonical correlation analysis (CCA) between both fields. This finding resembles observations (Bresch, 1998), which are dominated by the active phase (1963–1992).

Secondly, the wind-induced cold SST anomaly influences the atmospheric circulation and the North Atlantic storm track. For a positive NAO index the storm track shifts northward (not shown). Sensitivity experiments carried out with an atmospheric model of intermediate complexity show that a cold thermal anomaly reinforces the Iceland Low (Walter et al., 2001) and thus, closing this positive feedback loop. This equivalent-barotropic response due to synoptic activity is strongly non-linear and depends on the sign and position of the thermal forcing relative to the storm track in the North Atlantic (Peng and Whitaker, 1999; Walter et al., 2001). Note that the linear theory leads to a negative feedback, that is, the surface pressure is enhanced (reduced) near the cold (warm) SST anomaly (Hoskins and Karoly, 1981). Applying a CCA to 500 hPa geopotential height and SST in the North Atlantic sector shows that the SST anomalies associated with the leading CCA mode during late winter are uncorrelated with the following summer, but are significantly correlated (up to 0.62) with anomalies during the following winter. Again there is a correspondence to observational findings (Bhatt et al., 1998). The year-to-year correlation of winter SST can be explained by seasonal change of the mixed layer depth, and the reincorporation of water from below into the deepening fall mixed layer (Alexander and Deser, 1995). Thus, the memory of the ocean and the strength and the position of the storm track in the Atlantic are important for the enhancement of the low-frequency variability.

Hence, the positive feedback loop explains the maintenance of a distinct atmospheric circulation anomaly for several years. The regional regime is governed by a coupled atmosphere–ocean mode being composed of the SST tripole and the NAO pattern.

3.3. Precipitation in the decadal regimes

In the hemispheric regime the correlation pattern of precipitation with the NAO (Fig. 4a) and ENSO index (Fig. 4b) are very similar over the North Atlantic. A warm tropical eastern Pacific (ENSO index positive) and weak westerlies (NAO index negative) are connected with a weak Hadley cell in the Atlantic and a southward shift of the storm track. Moreover, the Atlantic teleconnection pattern (Fig. 6 in Raible et al., 2001) in the hemispheric regime is on a southwesterly position with its centres near 50 and 20°N (instead of 65 and 40°N in the regional regime) resembling findings from Alexander and Scott (2002). Due to the different positions of the teleconnection patterns in the hemispheric regime, atmospheric circulation anomalies are less effective in creating the SST-tripole compared to the regional ones.

In the regional regime the Hadley cell is still triggered by ENSO, the storm track is connected with NAO and the ENSO–NAO connection disappears. This is illustrated by the disagreement of the correlation pattern between precipitation and the teleconnection indices (Fig. 4, lower panel). The ENSO-rainfall connection in the tropical–subtropical western Atlantic is independent of the decadal regime whereas the North Atlantic storm track and therefore the precipitation decouples from ENSO in the regional regime.

A discussion of the decadal regime using observations is difficult due to the shortage of data. The precipitation of the CRU data set (Climate Research Unit) derived from land stations excludes the important North Atlantic ocean and the global NCEP-NCAR

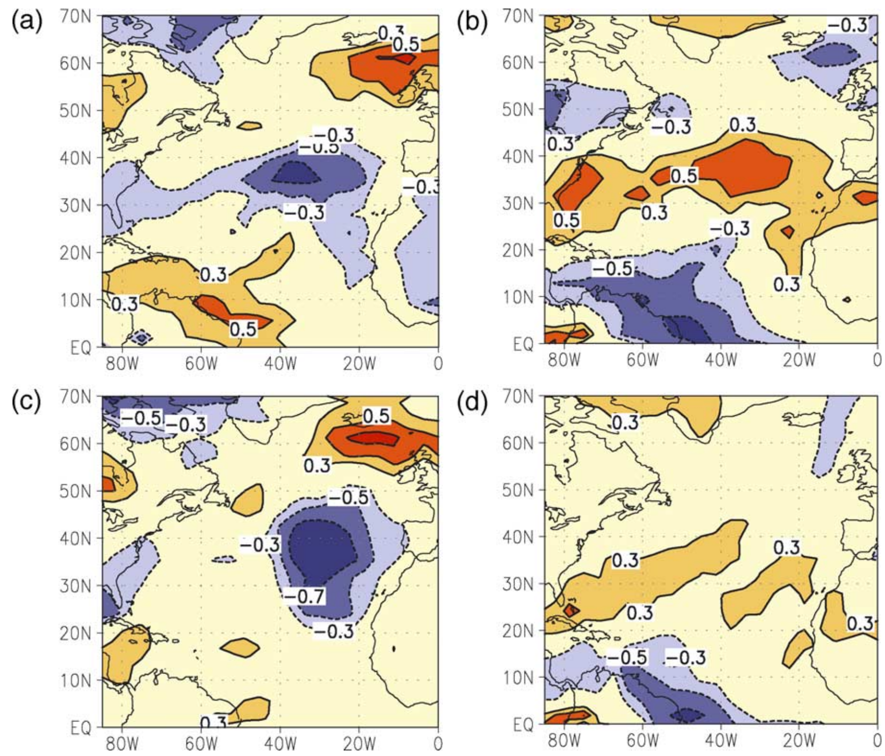

Fig. 4. Decadal regimes and changes of the local correlation between North Atlantic precipitation and the teleconnection indices for NAO (a,c) and ENSO (b,d). Upper panel: hemispheric regime, lower panel regional regime. Contour increment is 0.2. Correlations of more than 0.3 and less than − 0.3 are significant at a level of 95%.

reanalysis data set does not cover the passive phase. In the active phase (1963–1992) the high correlation coefficients of the NCEP precipitation (Fig. 5a) with NAO in the eastern North Atlantic illustrate the shift of the storm track to the north (south) during positive (negative) values of NAO (Sickmöeller et al., 2000). Also the weakening of the Hadley cell during an El Niño event (Klein et al., 1999) in the regional regime is supported by the NCEP reanalysis precipitation (Fig. 5b). There is no significant ENSO–Europe connection in this phase.

Due to the lack of reanalysis data for the passive phase we analyse CRU precipitation data derived from land station data for the period 1933–1962 (not shown). The shift of the storm track connected with NAO is indicated by the negative (positive) correlations of precipitation in Spain (England and Iceland). Near Florida the correlations with NAO are negative. During an El Niño phase a correlation dipole over Northern and Southern Europe is found with lower than normal precipitation over Denmark and Sweden (significant) and higher than normal values over Spain (not significant) which indicates a southward shift of the North Atlantic storm track. These observational findings resemble the passive phase of the simulation (Fig. 4, lower panel). Moreover, strong positive correlations over Florida

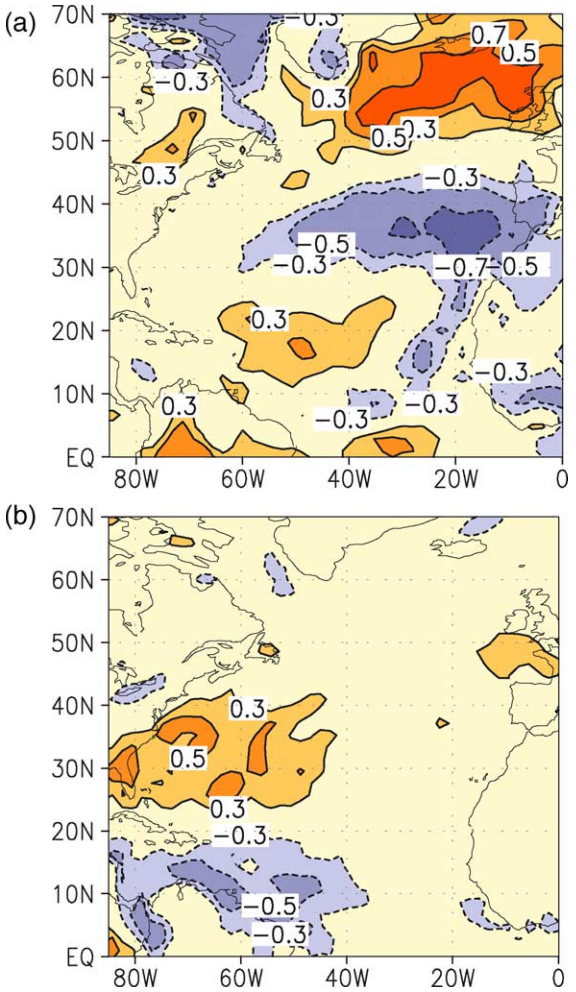

Fig. 5. As Fig. 4a and b, but for NCEP Reanalysis data from 1963 to 1992 which is governed by the regional regime.

and strong negative correlations over Venezuela show the connection to the weakening of the Hadley cell in the Atlantic during an El Niño phase.

4. Summary and conclusions

North Atlantic precipitation anomalies are considered to describe the differences of the decadal regimes in a longterm integration of a coupled ocean–atmosphere GCM. The dominant mechanisms of these regimes can be summarized schematically.

The hemispheric regime is triggered by ENSO. An anomalously warm tropical Pacific is connected with a stronger (weaker) Hadley cell in the Pacific (Atlantic) and a southward shift of the respective storm tracks. Therefore, the Azores High is anomalously weak and the precipitation increases (reduces) in the subtropical Atlantic (tropical and northern Atlantic). The NAO index is negative. A cold or La Niña event leads to sign reversal of these hemispheric circulation patterns resulting in a positive NAO. In this sense, the NAO is a slave of the ocean–atmosphere interaction in the tropical Pacific and about 40% of the NAO variability can be explained by ENSO.

The regional regime is characterized by enhanced decadal NAO variability which is generated by a positive feedback between NAO and SST anomalies. Starting with a positive NAO index, cold continental air and the associated Ekman transport cools the ocean near 50°N. Likewise, the enhanced Azores High generates a warm upstream (cold downstream) SST anomaly, leading to the well-known tripole SST pattern. The storm track shifts northward and the associated precipitation anomalies are negative near the Azores and positive further north. A cold (warm) SST anomaly strengthens the Iceland Low (Azores High) and closes the positive ocean–atmosphere feedback loop. The warm anomaly is advected along the path of the North Atlantic Currents which is an extension of the Gulf Stream with a mean velocity of about 2 cm/s (Sutton and Allen, 1997). Thus, with a delay of 5–10 yr the equivalent-barotropic response of this advected warm SST anomaly nearby Greenland weakens the Iceland Low possibly inducing a sign reversal of NAO.

The correlation of precipitation with the NAO and the ENSO index point to the non-stationarity of the northern hemisphere teleconnections: during active phases, enhanced decadal variability of the NAO index, the local ocean–atmosphere coupling in the Atlantic and the sectoral NAO paradigm may be appropriate. For passive phases, reduced decadal variability of the NAO index, the annular AO/NAM paradigm (wave train from the Pacific to the Atlantic) triggered by ENSO may be reasonable. Thus, proxies for precipitation in this key region could provide an estimate of the preceding non-stationarity of the atmosphere–ocean interaction.

The processes causing the shift from one to the other regime are still vague. Basically, atmospheric eigenmodes of NAO- and PNA-like structure exist already in atmospheric GCMs with fixed (climatological lower) boundary conditions (Barnett, 1987; Raible et al., 2001) and, clearly, without these purely atmospheric eigenmodes the decadal regimes with their distinct ocean–atmosphere interaction could not develop. The atmospheric dynamics itself may be a candidate for changing from one decadal regime to another (Raible et al., 2001). In a coupled ocean–atmosphere simulation an initially north-eastward shift of the teleconnection may be the start point for a positive feedback loop as described by the regional regime above.

Another possibility for the regime switch is the impact of the stratospheric variability on tropospheric modes (Perlwitz and Graf, 2001) which is not considered here. Strengthening or weakening of the stratospheric polar night jet may also influence the circulation patterns near the surface (Wallace and Thomson, 2002a). In the conceptual view presented here these regimes are signatures of the distinct ocean–atmosphere coupling: a positive feedback loop in the North Atlantic versus a hemispheric atmospheric response to tropical ENSO-like variations. The implications for seasonal midlatitude predictability of this non-stationary ocean–atmosphere coupling and the modification by the coupling to the stratosphere require further investigations.

Acknowledgements

The authors thank U. Merkel for discussions, U. Cubasch, S. Legutke, and R. Voss for providing the data of the coupled GCM simulation (Model Support Department at the Deutsches Klimarechenzentrum). This work is supported by the Deutsche Forschungsgesellschaft (Sonderforschungsbereich 512) and the National Centre for Competence in Research (NCCR) Climate funded by the Swiss National Science Foundation.

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