Low-frequency variability in the coupled ocean-atmosphere system at midlatitudes

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Low-frequency variability in the coupled ocean-atmosphere system at midlatitudes

Laagfrequente variabiliteit in het gekoppelde oceaan-atmosfeer systeem op gematigde breedte

(met een samenvatting in het Nederlands)

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Chapter 1

Introduction

1.1 Climate variability over the North Atlantic

In recent years the variability of the global climate system has been intensively studied. The motivation for many scientists is the need to be able to distinguish between natural and anthropological causes of climate change. Much attention is therefore devoted to low-frequency climate variability ranging from interannual to interdecadal time scales. Focusing on the Northern Hemisphere (NH), it has been considered that the possible impacts of global warming might differ from one area to another. The atmospheric NH climate system can be divided into a small number of sub-regions, mainly determined by standing oscillations in the planetary waves, with time scales on the order of one month or longer. Wallace and Gutzler (1981) identified and documented five spatial patterns which dominate the NH variability, namely the eastern Atlantic, the Eurasian, the western Pacific, the Pacific/North American and the western Atlantic pattern.

An area that has gained interest from many scientists over the last 5 to 10 years is the Atlantic sector. The winter-mean atmospheric circulation in this region becomes clear from the winter-mean 1000 hPa height field, shown in Fig. 1.1a. Between the high and low pressure centers near the Azores and Iceland a westerly wind prevails, causing cold and dry winter climate in Northeast America and moderate conditions in Western Europe. Deviations from the mean pattern are highest in winter and the main mode of NH interannual variability is much more evident in winter than in summer. In the Atlantic region, Wallace and Gutzler (1981) found a very strong teleconnection on the 1000 hPa level with strongest negative correlation between the grid points (65°N, 20°W) and (30°N, 20°W). This teleconnection pattern was first termed the North Atlantic Oscillation (NAO) by Walker and Bliss (1932). If this teleconnection pattern is considered in combination with the average situation (Fig. 1.1a) the NAO expresses as a weakening and strengthening of the meridional pressure dipole.

A more general way to find dominant patterns of variability is the so-called Empirical Orthogonal Function (EOF) analysis (details on EOF-analysis are found in Von Storch and Navarra (1993)). The most dominant pattern of atmospheric variability in the Atlantic region is given by EOF 1 (Fig. 1.1b) and clearly resembles the teleconnection of Wallace and Gutzler (1981). In order to quantify the amplitude of the NAO pattern, Hurrell (1995) defined an
index, the so-called NAO-index, as the winter-mean difference of normalized sea-level pressure (SLP) between Lisbon (Portugal) and Stykkisholmur (Iceland). A positive NAO-index indicates an increased meridional pressure difference and vice versa. The timeseries of the NAO-index is shown in Fig. 1.2. From this figure it becomes clear that the NAO shows variability on time scales, ranging from a few years to decades. A spectrum of this timeseries (Fig. 1.3) shows that, apart from a small peak at 2.5 years, the variability can not be distinguished from that due to white noise processes (Hurrell and Van Loon, 1997). However, the spectral peak between 6-9 years is important, as timeseries of variables related to the NAO do show peaks in the decadal range of the spectrum.

An interesting feature of the NAO is that the pattern also shows up at synoptical time scale, from a few days to several weeks. On this short time scale, a positive NAO phase is accompanied by a heavy zonal storm track over the North Atlantic, transporting cyclones towards western Europe. During a negative NAO phase persistent blocking occurs over
Figure 1.2: Winter (December through March) index of the NAO based on the difference of normalized sea-level pressures between Lisbon and Stykkisholmur, from 1864 through 2000. The seasonal SLP anomalies at each station were normalized by the long-term (1864-1983) standard deviation. Source: http://www.cgd.ucar.edu/~jhurrell/nao.html.

Figure 1.3: Power spectrum of the winter (December-March) NAO index for 1865-1994. From Hurrell and Van Loon (1997).

the North Atlantic reducing the influence of the Atlantic on Europe’s weather. Considering the Atlantic region (20-80°N, 90°W-20°E), the relative amount of variance (calculated from EOF-analysis) explained by the NAO (in the winter months December-March) is about 15% if 1-31 day high pass filtered daily pressure data are analyzed, about 35% for monthly mean values and about 50% for winter-mean pressure values. Hence, the relative importance of the NAO increases on longer time scales; throughout this thesis, focus will be on these longer time scales.

The impacts of variations in the NAO on the American and European winter climate are dramatic. The increased meridional pressure gradient during a positive phase of the NAO
causes an increased eastward jet over the Atlantic (Rogers, 1990; Cayan, 1992; Hurrell, 1995; Hurrell and Van Loon, 1997). For the North-American region, this means an enhanced inflow of dry and cold polar air from Northern Canada. On the Atlantic, an increased number and severeness of storms arise on the stronger jetstream, causing wetter and warmer than normal conditions in Northern Europe (Van Loon and Rogers, 1978; Hurrell, 1996; Rogers, 1997). Conditions in Southern Europe are more dry and cold, because simultaneously the storm track is directed less zonally and more north-eastward (Rogers, 1990; Cayan, 1992). All these impacts are reversed when the NAO moves into the negative phase.

Part of the research on the NAO has been focussed on the upward trend in the NAO-index since the early sixties (Fig. 1.2). It has been suggested that global warming is a possible cause for this trend. Therefore, several studies have focused on NAO behavior in future climate scenario’s. The results of these studies are not in agreement with each other, but in a majority of cases it is found that the NAO becomes more positive and storm track activity over Europe increases (Ulbrich and Christoph, 1999). This is tentatively explained by the strong increase of the meridional temperature gradient in the upper air over the North Atlantic.

In the above, focus has been on atmospheric low-frequency variability. However, in the last decades it has been shown in a number of studies that related variability is found in the Atlantic Ocean. Various atmospheric quantities appear to correlate very well with ocean variables on time scales ranging from years to decades. Therefore, the majority of studies on the low-frequency fluctuations of the NAO in recent years consider this variability as an ocean-atmosphere, coupled phenomenon. An oceanic quantity often used in studies on this subject is sea-surface temperature (SST). Fig. 1.4 shows the pattern of winter-mean SST and the first EOF, explaining 21 % of the variance. Characteristic is the sharp mean meridional SST gradient east of the American coast (Fig. 1.4a). The EOF pattern shown in Fig. 1.4b expresses as a weakening/strengthening of this meridional SST gradient. This pattern of variability and its temporal behavior correlates very well with the NAO pressure dipole shown in Fig. 1.1b, however, with the opposite sign.

Deser and Blackmon (1993) observe low-frequency variability by performing a spectral analysis on the Principal Components (timeseries of the EOF’s) of the dominant modes of interannual variability of observational oceanic and atmospheric data in the Atlantic region. Analyzing both SST and surface-air temperature (SAT), the dominant mode of variability displays a preferential period of about one decade (about 9 years before 1945 and about 12 years after the war). Kushnir (1994) observes timeseries from the beginning of this century of SAT, SST, surface winds and SLP. Two different modes of variability are examined. Kushnir (1994) finds anomalous high SST to be associated with reduced westerlies on interannual time scales, suggesting that the atmosphere is driving the ocean. On an interdecadal time scale, the anomalous SST fields show the inverse association between the wind and pressure pattern, such that the atmosphere seems to damp SST-variability. Kushnir (1994) suggests that on these long time scales the ocean might have a significant influence on the atmosphere.
Climate variability over the North Atlantic

Figure 1.4: a) Winter-mean SST (Dec-Feb, 1945-1994). b) EOF 1, explaining 21% of the total variance. Dark (light) shading where meridional gradient of the winter-mean SST exceeds 2°C (1°C). From Watanabe et al. (1999).
1.2 Mechanisms of low-frequency NAO variability

Many hypotheses have been put forward on the physical processes controlling the interannual and interdecadal NAO variability in ocean and atmosphere. An overview of the literature on the origin of low-frequency variability points in three directions:

(i) The variability in the atmospheric system might be generated within the atmosphere itself. Nonlinear interactions on various time scales could generate an internal atmospheric mode, showing low-frequency behavior. Recent studies dealing with this hypothesis are described in subsection 1.2.1. The related oceanic NAO variability is a response to the atmospheric fluctuations.

(ii) Variability in the oceanic system shows up at interannual and interdecadal time scales. Therefore, it is often suggested that the NAO-like low-frequency variations in the atmosphere are a reflection of variations in SST. Mechanisms responsible for internal ocean variability are extensively dealt with throughout this thesis. The atmospheric response to ocean variability, through SST, has been subject to a large number of studies, described in subsection 1.2.2.

(iii) Neither the atmosphere nor the ocean alone might be able to generate this kind of variability and an active coupling between the two systems plays a crucial role. For such an active coupling it is not only necessary to know how the atmosphere responds to the ocean (subsection 1.2.2), but also how the atmosphere drives the ocean. This is described in subsection 1.2.3 and then in subsection 1.2.4 mechanisms are discussed, in which coupled ocean-atmosphere feedbacks may cause low-frequency variability.

Although much progress has resulted from the recent efforts, yet there is no generally accepted theory to explain the low-frequency variability of the NAO.

1.2.1 Internal atmospheric variability

Experiments with simplified models have shown that interannual and interdecadal (and even longer time scale) variations of the extratropical atmospheric circulation may occur without any variation of the external forcing. James and James (1989) performed atmospheric model runs, in which they excluded any external forcing. They concluded that internal, nonlinear atmospheric processes are able to generate low-frequency variability. The atmospheric circulation in the extratropics showed the well-known baroclinic development (time scale 1-10 days), regime changes (10-100 days) and the annual cycle. It might be possible that by interaction of these different time scales interannual and interdecadal variability is generated. Liu and Opsteegh (1995) examined blocking activity on the Northern Hemisphere in a quasi-geostrophic model and found that external forcing does not significantly change the natural interannual to interdecadal variations.

Multi-year atmospheric variability may arise from interactions between the stratosphere and the troposphere. Perlwitz and Graf (1995) have shown that there is a strong statistical relation between the intensity of the stratospheric cyclonic vortex in winter and the tropospheric circulation over the North Atlantic. A strong stratospheric vortex is associated with a
positive phase of NAO and vice versa. Whether or not this is caused by changes of the vertically propagating planetary waves is not certain. However, it is well known that the structure of the stratospheric circulation is strongly determined by chemical and radiative processes, i.e. the stratospheric composition (ozone, volcanic particles) and variation in UV-fluxes (solar activity), respectively. If this external forcing would excite the NAO via interaction with the stratospheric circulation, this mechanism can not be interpreted as an internal mode. The strong connection between the polar vortex and the NAO has initiated a large discussion on the degree of locality of the NAO?" Thompson and Wallace (1998) show that, if the whole Northern Hemisphere is considered, the NAO dipole pressure pattern stretches in a zonal direction and extends all over the midlatitudes. They call this zonally stretched dipole structure the Arctic Oscillation (AO). Although it does seem reasonable that local midlatitude pressure variations (as the NAO) will affect other midlatitude regions via planetary Rossby waves, only very weak correlations have been found between the North Atlantic and Pacific atmospheric variability (Ambaum et al., 2001).

1.2.2 The atmospheric response to SST-anomalies

In an Atmospheric General Circulation Model (AGCM) with prescribed boundary conditions, a simulation of the low-frequency variability of the NAO over the last 50 years resembled observations quite well (Rodwell et al., 1999). What they found to be crucial for this good simulation was that the lower boundary of the model was forced with realistic, observed SST. This result suggests that the atmospheric response to SST-anomalies is important for the low-frequency variability of the NAO.

However, many discrepancies in the various studies on the atmospheric response to SST-forcing have shown that it is a rather difficult subject. The only consensus reached so far concerns the weakness of the atmospheric response to extratropical SST-anomalies it is hard to get definite conclusions on details of this response. The research on the El Niño-Southern Oscillation (ENSO) has shown that in the tropics diabatic heating (cooling) of the atmosphere by warm (cold) underlying SST-anomalies explains a relatively large part of the atmospheric variance. However, in the extratropics most of the variance is generated by internal processes, which thereby conceal any oceanic origin. This problem of a low signal-to-noise ratio is a recurrent obstacle in the various studies on this subject.

Many papers which appeared in the past decades essentially followed the approach of Palmer and Sun (1985), in order to shed some light on the extratropical atmospheric response to oceanic forcing. By examining the response of an AGCM to prescribed SST as the lower boundary condition, they preceded a row of comparable AGCM-studies. Palmer and Sun (1985) performed four pairs of 50-day runs (starting in November), a pair consisting of one run with warm and another with cold SST-anomalies. The anomalous SST pattern they prescribe is a cell centered eastward of Newfoundland. The motivation for this choice is that in the Newfoundland region the SST-anomalies in winter have relatively large amplitudes (up to 2-3 °C) and the overlying atmosphere is both dynamically and thermodynamically very active. The geopotential height response to warm SST-anomalies shows a large positive cell centered over the central Atlantic with an equivalent barotropic vertical structure. This means that a warm SST-anomaly will be associated with enhanced westerlies to the north-east and reduced westerlies to the south-east of the Newfoundland region. Thus, the climatological
jet stream shifts northward in this case. The high-pass eddy temperature and geopotential height variance (quantities associated with baroclinic wave activity) show a similar behavior: weakening (strengthening) to the south- (north-) east of the Newfoundland region. Palmer and Sun (1985) suggest that the SST-anomaly and the associated anomalous meridional temperature gradient could generate anomalous baroclinic development resulting in the response described above. With a further analysis on AGCM-runs with different amplitudes of the SST-anomaly in the New Foundland region, Palmer and Sun (1985) show a linear atmospheric response: geopotential height increases with about 20 meters per °C of SST forcing near sea level in the center of the anomalous geopotential height field. Lau and Nath (1990) confirm the results of Palmer and Sun (1985) and additionally they find an important role of transient disturbances to translate SST-anomalies into the time-mean atmospheric response. Anomalies in the meridional SST-gradient cause a shift of the storm track, which in turn causes anomalous patterns in eddy transports and sinks and sources of latent and sensible heat.

The papers of Pitcher et al. (1988) and Kushnir and Lau (1992) show discrepancies with the results described above. Both studies examine the AGCM-response to prescribed SST-anomalies in the Pacific. Pitcher et al. (1988) also find a negative (equivalent barotropic) geopotential height response just downstream of a prescribed cold SST-anomaly. However, in the opposite case of a warm anomaly they also find a (weaker) negative downstream pressure response, thus implying a nonlinear atmospheric response to SST forcing. Kushnir and Lau (1992) confirm the nonlinearity in the atmospheric response to Pacific SST-anomalies. They argue that the response does not only depend on air-sea interaction but also on the relative importance of the climatological atmospheric circulation. Strong horizontal advection of heat and momentum might enhance or decrease the response to the SST pattern and therefore cause nonlinear behavior.

The latter suggestion is examined in a study by Peng et al. (1995). They notice the importance of the different initial climatological situation in the AGCM-runs from Palmer and Sun (1985), starting in November, and the runs performed by Pitcher et al. (1988) and Kushnir and Lau (1992), which are both January integrations. Peng et al. (1995) perform exactly the same experiment as Palmer and Sun (1985), however, both with November and January initial conditions. The atmospheric response appears to be quite different in the early winter situation than in the midwinter one. In addition, Peng et al. (1995) examine vertical cross sections of atmospheric variables over the Atlantic and find, in November, that the heat released from the ocean into the atmosphere penetrates up to the upper troposphere and thereby causes a relatively strong response in geopotential height, downstream of the Newfoundland region. In January, only the lower troposphere is warmed by the anomalous SST. The strong horizontal cold advection from the continent dominates the vertical distribution of heat in the atmosphere overlying the Newfoundland region and thus the surface heating from the prescribed SST-anomalies is overruled by this cold advection.

Kushnir and Held (1996) point at another difficulty in determining the atmospheric response to SST-anomalies. In both a realistic and an idealized AGCM, they find the spatial patterns to be in reasonable agreement with the ones obtained by Palmer and Sun (1985). However, the geopotential height response shows a baroclinic vertical structure. Kushnir and Held (1996) state that their findings are probably due to the coarse gridded AGCM they use, which is unable to generate sufficient eddy-momentum fluxes to get the equivalent barotropic
atmospheric anomalies. Also other authors (e.g. Ferranti et al. (1994)) have suggested that a (coarse gridded) AGCM does not satisfactorily simulate the eddy momentum effects of transient midlatitude disturbances. An AGCM might therefore underestimate or change the nature of the atmospheric response to extratropical SST forcing.

In the studies summarized above, a recurrent conclusion is that large SST-anomalies up to some degrees Celsius (which do occur in the Atlantic ocean) might be able to cause a shift in the North Atlantic storm track which in turn influences the mean atmospheric circulation over the Atlantic region. However, the amplitude of this atmospheric response, which depends on the mean atmospheric circulation, is very small relative to that resulting from internal atmospheric variability.

1.2.3 The atmosphere driving the ocean

A basic theoretical view on this subject was given by Hasselmann (1976) and Frankignoul and Hasselmann (1977). Hasselmann (1976) shows that the ocean "integrates" atmospheric variability. The atmospheric variables that drive the ocean (e.g. heat flux, windstress) show a white frequency spectrum with energy uniformly distributed over all time scales. If these variables force oceanic variables to change in time, the latter will show a red frequency spectrum, with a larger energy contribution at the longer time scales. Thus, Hasselmann (1976) states that rapidly varying weather systems are able to excite a slow oceanic response. Wallace et al. (1990) analyzed the observed December-February mean 500 hPa height (the atmospheric variable which is often assumed to be representative for the atmosphere system, at least in studies dealing with the time scales we are dealing with) and SST, both in the North Atlantic and Pacific region. They expand the timeseries of the SST pattern in EOF's, where the first EOF explains 24% of the temporal variance of the SST field. Then they correlate the principal component of this EOF with the timeseries of the 500 hPa height and find (in the North Atlantic region) the spatial pattern of the correlation coefficient to resemble the NAO pattern. In order to address the question whether the anomalous SST pattern is an imprint of the correlated atmospheric pattern the same analysis is performed on 500 hPa height and midwinter SST-tendency, defined as mean February through April SST minus mean October through December SST. Again, Wallace et al. (1990) find the NAO-like 500 hPa height pattern to correlate well with an SST-tendency field, which resembles EOF 1 of SST. Hence, they conclude that the relatively strong coupling between EOF 1 of SST-tendency and 500 hPa height is a reflection of the atmospheric forcing of the ocean.

In a paper by Cayan (1992) the results from Wallace et al. (1990) are confirmed and explained in further detail. Cayan (1992) also takes the latent and sensible heat fluxes at the ocean-atmosphere interface into account. The results show that the heat-flux pattern (either latent or sensible heat flux) matches very well on the SST-tendency field and is also well correlated to the NAO pattern. In the positive NAO phase the stronger westerlies cause higher upward heat fluxes (and thus decreasing SST) between Newfoundland and Ireland.

A paper by Deser and Timlin (1997) deals with the temporal character of the ocean-atmosphere interaction. The 500 hPa height, SST-tendency and latent and sensible heat fluxes are analyzed, using Singular Value Decomposition (SVD, Wallace et al. (1992); Bretherton et al. (1992); Cherry (1997)). Deser and Timlin (1997) perform the SVD-analysis on weekly averaged fields and introduce time lags between the different fields varying from -4 weeks
Introduction

(ocean variables leading) to +4 weeks (atmosphere leading). The ocean-atmosphere correlation appears to be largest when 500 hPa height leads SST by 2-3 weeks. The spatial correlation patterns associated with this lag, resemble the spatial patterns found by Wallace et al. (1990) and Cayan (1992). These are a NAO-like dipole pressure pattern in the atmosphere and a tripole pattern in SST, with negative SST-anomalies between Newfoundland and Ireland and east of Africa and positive SST-anomalies west of the North-American coast. Thus, the results suggest that a positive NAO cools the ocean surface in areas of anomalously strong winds and warms the ocean surface where winds are weaker than normal.

1.2.4 Coupled mechanisms

From the previous subsections it already appeared that low-frequency oceanic and atmospheric variability are highly related. Motivated by the success in the understanding of the ENSO phenomenon, a large part of the research on the NAO has focused on identifying ocean-atmosphere feedbacks which might induce or enhance low-frequency variability.

In a very simple model, consisting of a one-layer atmosphere coupled to an 1.5-layer, quasi-geostrophic ocean, Liu (1993) identifies three possible interannual positive feedbacks (Fig. 1.5). The "upwelling mode" becomes unstable if the atmospheric response to a warm SST-anomaly is a high-pressure center (Fig. 1.5a). The associated anti-cyclonic wind stress

![Figure 1.5: This schematic figure shows meridional sections of a) the unstable upwelling mode, b) damped upwelling mode, c) unstable SST-Sverdrup mode, d) the damped SST-Sverdrup mode, e) the unstable SST-evaporation mode, f) the damped SST-evaporation mode. In all cases, the initial disturbances are a warm SST anomaly, denoted by a small W; the perturbation temperature due to coupling is denoted by a midsized W. The arrows represent the ocean current forced by the anomalous wind stress. The $\mathbf{W}$ and $\mathbf{C}$ in c) and d) represent the mean temperature; $\mathbf{C}$ is the mean wind. See the text for explanation. From Liu (1993).]
forces downwelling and therefore warms the SST further. This mode seems to favor low-latitude regions in which the mixed layer is shallow. If the atmospheric response over the warm SST-anomaly is a low-pressure anomaly, the upwelling mode is damped (Fig. 1.5b). However, in this situation the "SST-Sverdrup mode" is unstable. The low-pressure center with cyclonic wind stress forces a northward anomalous Sverdrup ocean current, resulting in an increased positive SST-anomaly (Fig. 1.5c). This situation is most likely to occur in regions with a high meridional SST-gradient (such as the Gulf Stream region). The SST-Sverdrup mode is damped if the atmospheric response is a high-pressure anomaly (Fig. 1.5d). The "SST-evaporation mode" in Liu's model is unstable (stable) if an easterly (westerly) anomalous wind overlies a warm SST-anomaly. This anomalous wind reduces (enhances) the mean wind pattern and thereby the evaporative cooling, intensifying (weakening) the initial SST-anomaly (Figs. 1.5e,f). This mode favors high latitudes.

Using an analytical stochastic model, Saravanan and McWilliams (1998) show that only few ingredients are needed to excite a decadal coupled mode. The ocean-atmosphere coupling is simply represented by a one-dimensional sinusoidal north-south dipole pattern in the atmosphere, stochastically forcing a slab ocean with a constant northward advection velocity. Provided that the horizontal advection dominates thermal damping (when the advection velocity and the depth of the slab ocean are large enough), the oceanic response shows a significant spectral peak. An associated preferential time scale for the atmosphere is hard to detect because of the white noise which might conceal a significant atmospheric peak.

In Weng and Neelin (1998) and Neelin and Weng (1999), the variability of the atmosphere is described by a stochastic part and a part which is associated with a specific SST pattern. This atmosphere is coupled to a quasi-geostrophic upper-ocean layer within a rectangular basin. The power spectrum of the coupled system shows a preference for interdecadal time scales. This is caused by ocean Rossby-wave dynamics of which the zonal length scale is imposed by the scale of the atmospheric wind stress forcing. Hence, a spatial and temporal "matching" of the oceanic and atmospheric perturbations occurs, raising a spectral peak at that specific time scale.

Sutton and Allen (1997), Da Costa and Colin de Verdière (2002), Latif and Barnett (1994), Latif et al. (1996), Grötzner et al. (1998), Timmermann et al. (1998) and Selten et al. (1999) describe different mechanisms leading to low-frequency variability in the coupled system. However, the basic feedbacks involved show strong similarities: a positive, direct (synchronous) ocean-atmosphere feedback enhances the amplitude of an initial perturbation and a negative, lagged feedback eventually pushes the coupled system into the reversed situation. A schematic picture of these feedback mechanisms is plotted in Fig. 1.6. A positive NAO phase (Fig. 1.6a) results in an enhanced stormtrack at high latitudes and reduced westerlies at lower latitudes (Fig. 1.6b). This anomalous atmospheric forcing cools the ocean east of Newfoundland and warms the ocean east of Florida (Fig. 1.6c). If the atmosphere responds to this anomalous SST pattern according to Palmer and Sun (1985), the resulting pressure response is a positive cell in the central Atlantic and a negative cell in the northern Atlantic (Fig. 1.6d), enhancing the positive NAO phase. Apart from this positive feedback, the positive NAO phase (Fig. 1.6e) also causes an anomalous forcing of the ocean circulation. The response of the ocean to this forcing differs in the various studies, but the common feature is that the response lags the forcing by a few years to decades. The response is either a changed wind-driven circulation, Rossby-wave adjustment, or a changed thermohaline circu-
lation (Fig. 1.6f). This lagged response eventually (in different ways, details described below) causes a reversed SST pattern, with warm SST near New Foundland and cold anomalies near Florida (Fig. 1.6g), which forces a negative NAO (Fig. 1.6h), giving a negative feedback. Thus, the positive feedback mechanism determines the amplitude, the negative feedback the timing of the oscillation. Details of these feedbacks in the various studies are given below.

Sutton and Allen (1997) propose a mechanism of decadal variability in the Atlantic region. Based on an observational study, they find SST-anomalies in the Gulf Stream to vary on decadal time scales (12-14 year) and to propagate along the Gulf Stream path, however, with an average velocity considerably slower than the near-surface Gulf Stream current. Warm SST-anomalies near Florida increase the winter-mean, continent-ocean temperature gradient, which would enhance cyclone activity. As in this region many cyclones form, they consider this area to be sensitive to such a change in temperature gradient. A larger number of cyclones (on synoptic time scale) might cause the NAO to favor the positive phase (in case of warm SST an increased temperature gradient causes an intensified storm track). This changed atmospheric circulation will change the subtropical wind-driven ocean gyre and furthermore excite large-scale baroclinic Rossby waves. This combination of advection and Rossby waves will eventually cause cold anomalies in the "storm-formation region" near Florida. Hence, the decadal time scale is set by a combination of advection and Rossby wave dynamics. The propagation of observed SST-anomalies along the path of the Gulf Stream was also found by Watanabe et al. (1999) and Da Costa and Colin de Verdière (2002). In these studies, the propagation is also assumed to be a combination of advection by the mean circulation and Rossby wave propagation. However, in contradiction to Sutton and Allen (1997), Da Costa and Colin de Verdière (2002) find the NAO to switch sign if SST-anomalies reach the Newfoundland region.

Another decadal oscillation mechanism is proposed by Latif and Barnett (1994). This was later examined in Coupled General Circulation Models (CGCM’s) by Latif et al. (1996), mainly focusing on the Pacific and by Grötzner et al. (1998) for the North Atlantic. The mechanism is based on the interaction between low-frequency variability in the large-scale atmospheric circulation and the wind-driven oceanic subtropical gyre circulation. The preferred period is found to be about 20 years in the Pacific (Latif and Barnett, 1994; Latif et al., 1996) and about 17 years in the North Atlantic (Grötzner et al., 1998). Starting with a positive phase of the NAO, the positive feedback mechanism described in Fig. 1.6 increases the initial anomalies. However, in the meantime, during this positive NAO-phase, there is an enhanced wind-stress curl, which spins up the double gyre ocean circulation. An increased subtropical gyre transports warm water northwards, thereby decreasing the meridional SST-gradient, which in turn weakens the meridional pressure gradient. Thus, the latter mechanism causes a negative ocean-atmosphere feedback. Latif and Barnett (1994) argue that the ocean circulation takes a while to adjust to the atmospheric anomalous forcing and that this time lag sets the oscillation period. Recently, this idea of direct air-sea interaction and a lagged interaction through the ocean adjustment time is thoroughly examined by Cessi (2000), Primeau and Cessi (2001) and Marshall et al. (2001).

In a multi-century run of a global CGCM, Timmermann et al. (1998) simulate inter-decadal variability with a dominant period of 35 years. However, here, the variability in the thermohaline circulation (THC) is involved in the negative feedback mechanism. If the THC is strong, warm water is advected into the North Atlantic basin. This anomalous SST pattern
Figure 1.6: Schematic picture of the feedback mechanisms described in the text. a)- d) show direct forcing mechanisms, resulting in the positive feedback, e)- h) show subsequent forcing mechanisms, resulting in the negative feedback. H and L in a), d), e) and h) denote high and low pressure anomaly, respectively.
1. Introduction

excites a strong positive geopotential height cell over the central Atlantic, indicating a positive NAO phase. The intensified storm track will generate negative sea-surface salinity (SSS) anomalies in the sinking region near Greenland, due to an increased freshwater flux. This low SSS in turn causes the deep convection to weaken, thereby weakening the THC.

Selten et al. (1999) use a three level quasi-geostrophic, coupled, global intermediate model, essentially constructed for research on midlatitude climate variability. The last 1000 years of a 5000-year run simulate the observed spatial patterns of variability quite well and display a preferential period of 16-18 years. The essential feedbacks include ocean-atmosphere interaction, geostrophic ocean circulation and convection. Starting with a positive NAO phase the ocean is forced into the dipole pattern with a cold anomaly near Newfoundland. The negative SST-anomaly cools the lower troposphere, isolating the SST-anomaly and thereby increasing its lifetime. Furthermore, the response in the atmosphere causes an increase in the probability that the NAO persists in its positive phase. The internal oceanic response to the SST dipole is a geostrophic circulation which advects relatively less salty water northward, where it will decrease convection, which in turn warms the northern SST, finally resulting in the opposite phase of the SST dipole as well as the NAO pattern. Selten et al. (1999) perform additional experiments which emphasize that the atmospheric response to the SST forcing is not essential to the decadal coupled mode, but that it does enhance the amplitude of the oscillation and possibly changes its period.

Bladé (1997) and Saravanan (1998) compare several simulations of both coupled and uncoupled GCM’s. The major conclusion in these studies confirms the findings of Selten et al. (1999) in that ocean-atmosphere coupling does not modify the low-frequency spatial patterns of variability, but does alter both amplitude and temporal character of the variability.

1.3 Scope of this thesis

The work in this thesis provides an alternative approach for the understanding of the physical mechanisms of low-frequency variability in the midlatitude coupled ocean-atmosphere climate system.

Although observations clearly are the most direct link to what is happening in the climate system and especially on time scales up till a few years, analyzing observed data in order to explore the decadal to interdecadal variability of the NAO has two major drawbacks.

(i) Detailed data records of atmospheric fields exist only for about a hundred years and reliable oceanic data can only be found after the second World War. Because of the relatively small difference between the length of the data records and time scale of the phenomenon we are investigating, it will be difficult to establish results which are statistically significant.

(ii) The low-frequency signal of NAO only explains a relatively small part of the total variance. In other words, the signal-to-noise-ratio is very small and therefore all observed data will show variability on a number of time scales. Within this very ‘noisy’ climate system, it will be difficult to extract the specific patterns and time scales of variability leading to an understanding of the behavior of the NAO.
Hence, an extensive modeling effort is certainly needed, in addition to the analysis of observations, in order to understand the physics of low-frequency NAO variability.

Results with state-of-the-art coupled GCM’s have shown that these models are quite capable of simulating NAO-like low-frequency fluctuations. Simulations of hundreds or even thousands of years have been performed and the analyses have revealed statistically significant decadal and/or interdecadal variability. However, also this approach has a number of drawbacks.

(i) The ‘virtual’ world represented in these GCM’s is a multi-scale system, with many degrees of freedom. In these systems, "the signal” will still be hard to distinguish from "the noise”.

(ii) Many physical processes are acting and analyses of several runs will usually not lead to specific physical mechanisms. Hence, it is hard to identify the processes which dominantly contribute to the low-frequency variability of the NAO.

In order to isolate physical processes which can, in principle, be responsible for the low-frequency variations in the midlatitude coupled climate system, conceptual models are necessary.

In the latter models, which range from simple box-models to intermediate complexity models, only those characteristics from the real climate system are extracted, which are thought to be of crucial importance. More advanced techniques of applied mathematics can be assigned to these simpler models to identify the physical mechanisms controlling a certain type of phenomenon. In this way, results within a hierarchy of models can be obtained which will eventually lead to an interpretation framework for the results of GCM’s and observations.

In confronting modeling results with observations, one has to realize that it is the largest scales that are best and most reliably captured. The low-frequency variability of the large scales arises from two sources: (i) the competition among the finite-amplitude instabilities and (ii) the net effects of the smaller scales of motion. Dynamical systems theory (Guckenheimer and Holmes, 1983) provides a perfect toolkit to analyze the physics of the variability on these large scales, since it is possible to identify the specific spatial-temporal patterns involved in (i).

The new element in this thesis is that a dynamical systems approach is used within midlatitude coupled ocean-atmosphere models of intermediate complexity to identify the origin of low-frequency variability of the NAO and the results are used to interpret decadal variability found in a coupled state-of-the-art GCM.

The system studied in chapter 2 (Van der Avoird et al., 2002b) consists of a two-layer quasi-geostrophic atmospheric model which is coupled to a one- or two-layer quasi-geostrophic ocean model within a zonally unbounded channel. A constant depth surface layer model describes the evolution of sea-surface temperature of the ocean. Since the atmospheric response to SST-anomalies is a delicate issue, an analysis on observational data is performed in order to find a proper parameterization of the ocean-atmosphere coupling. This coupled model is one of the simplest dynamical models which may represent low-frequency variability of the NAO and has been chosen because the internal variability in the uncoupled systems is well known (Pedlosky, 1987).

The analysis of this model consists of monitoring stationary solutions as one specific physical parameter (the control parameter) is varied. This control parameter will either vary
the strength of the ocean circulation or the strength of the coupling between ocean and atmosphere. Simultaneously, the stability of these solutions is determined with respect to infinitesimally small perturbations and the patterns of the most unstable (or least stable) oscillatory modes are calculated. At selected locations in parameter space, the study is complemented with transient flow computations. The central questions in this study are:

(i) Does low-frequency variability arise through oscillatory modes which destabilize stationary solutions?

(ii) Can the physical processes be identified which destabilize these oscillatory modes?

(iii) Do these oscillatory modes indeed contribute to the low-frequency variability in the transient flows in the highly nonlinear regime?

In chapter 2, these questions can be answered in detail and the results motivate the analyses in two other members of the model hierarchy in subsequent chapters. In chapter 3, the only new element introduced is the closure of the eastern and western ocean boundaries of the ocean basin. This leads to a more realistic double-gyre ocean circulation which is coupled to an atmospheric eastward jet. Again, the internal variability of the uncoupled components is well known (Dijkstra, 2000) and a dynamical systems analysis is performed on the coupled model, with the basic questions as above.

The qualitative results, in particular the hypotheses on the physical mechanisms of low-frequency variability coming from these intermediate complexity models, next provide a very specific data-analysis of results from a 300-year simulation of a state-of-the-art coupled ocean-atmosphere GCM in chapter 4 (Van der Avoird et al., 2002a). Patterns of low-frequency variability are extracted using recent statistical techniques. The central questions in chapter 4 are:

(i) Can we identify specific patterns in the ocean and the atmosphere which can be connected to low-frequency variability of the NAO?

(ii) Can we understand this low-frequency variability from the interpretation framework provided by the results of the intermediate coupled models?

Finally in chapter 5, results of the thesis are put in context and the new physical mechanisms involved in low-frequency variability are evaluated along with current theories.
Chapter 2

Low-frequency variability in the jet-jet flow

Low-frequency variability in the coupled midlatitude ocean-atmosphere system is explored. The intermediate-complexity model used for this purpose assumes quasi-geostrophic dynamics in both ocean and atmosphere. The latter are coupled through processes controlling the sea-surface temperature within a constant depth surface layer. Linear stability analysis of an idealized background state, characterized by zonal jets both in the atmosphere and ocean and a simplified sea-surface temperature profile, reveals that a large-scale mode may destabilize once the coupling strength is large enough. The mode corresponds to a near-stationary barotropic Rossby wave in the atmosphere coupled to a large-scale baroclinic oceanic Rossby wave. Computations of transient flows in the highly nonlinear regime show that nonlinear rectification processes of this coupled mode change the mean state in such a way that it becomes susceptible to high-frequency instabilities which, again through rectification processes, stabilize the coupled mode. The (generic) low-frequency variability which arises from these rectification processes is associated with amplification/weakening and North/South shifting of the zonal jets, very much resembling observed patterns of variability at midlatitudes.

2.1 Introduction

In chapter 1 it is shown that the NAO expresses as a weakening and strengthening of the winter mean pressure dipole between Iceland and the Azores with an associated weaker and stronger eastward jet over the northern part of the basin (Rogers, 1990; Cayan, 1992; Hurrell, 1995; Hurrell and Van Loon, 1997). Another significant part of the atmospheric variance in this area is caused by meridional shifting of this jet (Rogers, 1990; Cayan, 1992). Related low-frequency variability in the North Atlantic sector is also present in the ocean. Observational studies show interannual to interdecadal fluctuations in large-scale SST patterns (Deser and Blackmon, 1993; Kushnir, 1994). Several hypotheses have been put forward to explain the physical mechanisms, responsible for the low-frequency variability in both ocean and atmosphere.
In this chapter we will analyze the physical mechanism of low-frequency variability present in a strongly idealized model of the midlatitude climate system. The basic ocean dynamics of the model are captured in quasi-geostrophic vorticity equations for two layers. The lower layer, representing the deep ocean is initially at rest. In the upper layer, which is forced by the atmosphere, a simple eastward jet represents the Gulf Stream/North Atlantic Current. The atmosphere model also has two layers in which quasi-geostrophic vorticity equations describe the flow. In both layers an eastward jet represents the atmospheric jet-stream between the Iceland Low and Azores High (Fig. 1.1). In both atmospheric layers and in the upper ocean layer periodic boundary conditions at the eastern and western boundaries lead to a zonal flow. The zonal velocities are zero at the northern and southern boundary and maximum in the center of the basin.

In our model, the lower atmosphere forces the upper ocean layer through the wind stress, but apart from this dynamical coupling an SST-equation couples ocean and atmosphere. SST depends on both atmosphere layers and the upper ocean layer and feeds back on both atmospheric layers. Because chapter 1 has shown that this coupling is subject to an ongoing debate, an alternative conceptual parameterization of the air-sea interaction is proposed to represent coupled processes and the parameters involved are estimated using observations in the North Atlantic region.

The stability of zonal flows in a two-layer channel has been studied extensively. In a continuously stratified inviscid fluid, Eady (1949) examines the stability of a zonal jet, having only vertical shear. This basic system becomes baroclinically unstable if the wavelength of the perturbations is larger than the internal Rossby deformation radius. Pedlosky (1964), using a zonal flow with both vertical and horizontal shear, shows that stability characteristics change when the flow becomes unstable to both baroclinic and barotropic perturbations. Much later, Pedlosky (1987) reviews the stability characteristics of various stationary jet profiles. Both the growth rate and period of unstable, oscillatory, mixed baroclinic/barotropic modes not only depend on the strength but also on the shape of the zonal jet.

The idealized, coupled system of an atmospheric jet over an ocean jet has recently been studied by Goodman and Marshall (1999). They have also considered a two-layer quasi-geostrophic atmosphere model coupled to a two-layer quasi-geostrophic ocean model with a simplified representation of the air-sea interaction. From the dispersion relations, the mechanisms of amplification of the perturbations are described, which are more or less the same as that of the SST-upwelling and the SST-Sverdrup mode in Liu (1993).

The chapter is arranged as follows. Having presented the coupled ocean-atmosphere model (section 2.2) a linear stability analysis of the idealized background state is performed in section 2.3. Subsequently, the highly nonlinear regime is investigated using computations of transient flows in section 2.4. In section 2.5 the results are summarized and discussed.
2.2 Model formulation

Two dynamically active layers both in the ocean and atmosphere are coupled through a constant depth surface layer embedded in the upper ocean layer (Fig. 2.1). A standard channel set-up, having a width $L$, on a $\beta$-plane with Coriolis parameter $f = f_0 + \beta_0 y$ is chosen.

![Figure 2.1: Schematic representation of the coupled model with a two-layer ocean model coupled to a two-layer atmosphere model through a surface layer.](image)

2.2.1 Governing equations

In the two-layer ocean model, the density $\rho_i$ and mean layer thickness $H_i$ ($H = H_1 + H_2$) of each layer $i$ are constant and the flow in the ocean is forced by a wind stress $\tau = (\tau_x, \tau_y)$. By defining the streamfunction $\psi_{oi}$ in both layers through $u_{oi} = -\partial \psi_{oi}/\partial y, v_{oi} = \partial \psi_{oi}/\partial x$, where $u_{oi}$ and $v_{oi}$ are the zonal and meridional velocity, respectively, the quasi-geostrophic equations describing the flow are given by (Pedlosky, 1987)

\[
\begin{align*}
\frac{D\psi_1}{dt} & = \nabla^2 \psi_1 + \beta_0 y - \frac{1}{\lambda^2} (\psi_1 - \psi_2) = \nabla \cdot \left( \frac{\tau}{\rho_1 H_1} \times e_3 \right) + A_o \nabla^4 \psi_1 \quad (2.1a) \\
\frac{D\psi_2}{dt} & = \nabla^2 \psi_2 + \beta_0 y + \delta^0 \frac{1}{\lambda^2} (\psi_1 - \psi_2) = A_o \nabla^4 \psi_2 - r \nabla^2 \psi_2 \quad (2.1b) \\
\end{align*}
\]

In (2.1), $\lambda = \sqrt{g' H_1 / f_0}$ is the Rossby deformation radius, in which $g' = g(\rho_2 - \rho_1)/\rho_2$ is the reduced gravity, $e_3$ is the unit vector in vertical direction, $A_o$ is the lateral friction coefficient, $\delta^0 = H_1/H_2$ is the ratio of the layer thicknesses and $r$ is the bottom friction coefficient. The parameterization of the wind stress $\tau$ is considered in section 2.2.2 below.
The evolution of the sea-surface temperature $T$ is only represented in a surface layer with constant depth $H_m$ in which horizontal advection of heat occurs with the velocities of the upper ocean layer. The SST equation becomes:

$$\rho_o H_m C_p^o \frac{D_o T}{dt} + u_E \frac{\partial T}{\partial x} + v_E \frac{\partial T}{\partial y} = \rho_o H_m C_p^o K_T \nabla^2 T + Q_o$$

(2.2)

where $\rho_o$ is the density of the surface layer ($\rho_o = \rho_1$), $C_p^o$ is the oceanic heat capacity, $K_T$ is the thermal diffusivity, $u_E, v_E$ are Ekman velocities and $Q_o$ is the diabatic heating of the surface layer through the downward heat flux at the ocean-atmosphere interface ($Q_{oa}$) and through entrainment fluxes at the base of the surface layer. In Goodman and Marshall (1999), it is shown that both vertical as well as horizontal advection processes can give rise to unstable interactions. In this chapter, we have chosen to focus only on the latter transport mechanism for reasons of simplicity and because the essential results are not affected by neglecting the effect of one of these processes. Therefore, in the results shown below we use $u_E = v_E = 0$ and $Q_o = Q_{oa}$.

The atmosphere model consists of two layers (in isobaric coordinates) of equal mean thickness $\Delta p$. The streamfunction $\psi_{ai}$ is defined at $p_1 = 1/2 \Delta p$ and $p_2 = 3/2 \Delta p$ in such a way that $u_{ai} = -\partial \psi_{ai}/\partial y$ and $v_{ai} = \partial \psi_{ai}/\partial x$. The equations describing the atmospheric flow are given by (Mukougawa, 1988; Holton, 1992)

$$\frac{D_1^q}{dt} \left[ \nabla^2 \psi_{a1} + \beta_0 y - \frac{f_0^2}{S \Delta p^2} (\psi_{a1} - \psi_{a2}) \right] = -\alpha_a H_a - k_a \nabla^2 \psi_{a1} + A_a \nabla^4 \psi_{a1}$$

(2.3a)

$$\frac{D_2^q}{dt} \left[ \nabla^2 \psi_{a2} + \beta_0 y + \frac{f_0^2}{S \Delta p^2} (\psi_{a1} - \psi_{a2}) \right] = \alpha_a H_a - k_a \nabla^2 \psi_{a2} + A_a \nabla^4 \psi_{a2}$$

(2.3b)

In these equations, $S$ is the atmospheric static stability parameter, $\alpha_a$ is a coefficient relating internal heating to vorticity tendency (Holton, 1992), $H_a$ is the internal heating of the atmosphere at the mid-pressure level, $k_a$ is a friction coefficient arising from internal friction between the two layers (Mukougawa, 1988) or from the bottom atmospheric Ekman layer and $A_a$ is the (Laplacian) lateral friction parameter. The atmospheric jetstream is known to be unstable to an (almost) unlimited number of perturbations. However, because in this study we are mainly interested in low-frequency variations, the frictional term $A_a$ damps the high-frequency variability.

In both ocean and atmosphere, slip boundary conditions are used at the lateral walls. The heat flux in the surface layer is assumed to vanish at the lateral walls and periodic boundary conditions for all quantities are applied at the eastern and western boundaries.
2.2.2 Air-sea coupling

The parameterization of the wind stress coupling is based on the assumption that it can be taken proportional to the velocity in the lower atmospheric layer, i.e.

\[
\frac{\tau^x}{\rho_1 H_1} = -\gamma_3 \frac{\partial \psi_{a2}}{\partial y} \quad (2.4a)
\]

\[
\frac{\tau^y}{\rho_1 H_1} = \gamma_3 \frac{\partial \psi_{a2}}{\partial x} \quad (2.4b)
\]

where \(\gamma_3\) is the momentum coupling coefficient. As a standard value for \(\gamma_3\) we use \(1/100\) days\(^{-1}\) as in Liu (1993).

For the downward heat flux \(Q_{oa}\) (Wm\(^{-2}\)) at the ocean-atmosphere interface and the internal heating \(Q_a\) (Wm\(^{-3}\)) at the mid-pressure level, we distinguish between external forcing mechanisms independent of air-sea interaction (\(\overline{Q}_{oa}, \overline{Q}_{a}\)) and a remainder part. Introducing an atmospheric temperature just above the sea surface \(T_s^a\), the downward heat flux can be written as (Haney, 1971)

\[
Q_{oa} = \overline{Q}_{oa} + K_S V_a (T_s^a - T) \quad (2.5)
\]

where \(K_S\) is an overall heat transfer coefficient depending on the airspeed \(V_a\) just above the ocean-atmosphere surface. The internal heating of the atmosphere \(Q_a\) is expressed as (Frankignoul, 1985)

\[
H_a = H_{oa} + K_S V_a (T - T_s^a) g_1 \quad (2.6)
\]

The second term on the right hand side of equation (2.6) arises by multiplying the surface heat flux by some fixed heating structure in the atmosphere, represented here by a constant \(g_1\) (m\(^{-1}\)). In this way, the internal heating is not only caused by anomalous SST but also depends on the dynamical atmospheric response (through \(T_s^a\)).

In order to close the set of equations we have to parameterize \(T_s^a\) in terms of the unknown quantities \((\psi_{o1}, \psi_{o2}, \psi_{a1}, \psi_{a2}, T)\). The most straightforward way to do this is to assume a linear dependence of \(T_s^a\) on both sea-surface temperature \((T)\) and the atmospheric temperature at the interface between both atmospheric layers \((T_a)\), giving

\[
T_s^a = \gamma_1 T + \gamma_2 T_a \quad (2.7)
\]

where \(\gamma_1\) and \(\gamma_2\) are thermal coupling coefficients. \(T_a\) is related to the atmospheric stream-function through

\[
T_a = -\frac{p_a}{R} \frac{\partial \Phi}{\partial p} \quad (2.8)
\]

where \(R\) is the specific gas constant for air, \(p_a\) is the (fixed) reference atmospheric pressure at the interface and \(\Phi\) is the geopotential, which is given by \(\Phi = f_0 \psi_a\); the gradient \(\partial \Phi/\partial p\) can be approximated by \(f_0 (\psi_{a1} - \psi_{a2})/\Delta p\).
2.2.3 Parameter estimation

In equation (2.6) it is assumed that the internal heating of the atmosphere is linearly related to the heat flux at the ocean-atmosphere interface. Furthermore, the air temperature just above this interface is assumed to be linearly dependent on SST and on the atmospheric mid-level temperature $T_a$ (cf. 2.7). However, are these relations reasonable representations of the complex processes which actually take place and can the values of the parameters $g_1$, $\gamma_1$ and $\gamma_2$ be estimated? To address both issues, the monthly mean NCEP-reanalysis data set (2.5°×2.5°) and reconstructed SST’s (linearly interpolated to 2.5°×2.5°) over the years 1950-1995 were analyzed. To identify observations and model variables, we take $T_a$ and $T_{as}$ as the temperatures at 500hPa and 1000hPa, respectively and $V_a$ as the wind speed at 1000hPa.

The midlatitude atmospheric response to SST forcing in GCM studies is rather weak, but several authors (e.g. Ferranti et al. (1994); Kushnir and Held (1996)) have suggested that in coarse grid GCM’s there is an underestimation of the vertical heat transfer through anomalous eddy forcing. Other studies (Palmer and Sun, 1985; Lau and Nath, 1990) indicate that for the Atlantic sector, the atmosphere is most sensitive in winter to SST anomalies eastward of New Foundland. In this region, the southwest-northeast directed North Atlantic storm track has maximum amplitude. This gives a relatively high convergence of the monthly mean eddy heat flux $\overline{u T}$ at the $500\text{hPa}$ level. The field $\nabla.\overline{(uT)}$ is plotted in Fig. 2.2a; values in the rest of the North Atlantic region are much smaller and are not shown. These results indicate that there is an internal heating of the winter mean mid-level atmosphere through eddy heat fluxes. At the ocean-atmosphere surface there is an upward heat flux, as can be seen in the mean field of $T - T_{as}$ (Fig. 2.2b). Also here the values are highest near New Foundland. Thus, the heat released from the ocean seems to be partly transported to the mid-level atmosphere through eddy heat fluxes.

When a linear correlation between $-\rho_o C_p^a \nabla.\overline{(uT)}$ and $K_S(V_a)(T - T_{as})$ is calculated, the following results are obtained. For constant $K_S = 40 \text{ Wm}^{-2} \text{K}^{-1}$ (Frankignoul, 1985),
Figure 2.3: Estimate of the value of the coefficient $g_1 \,(m^{-1})$. Shading indicates regions where the relation between $-\rho_c C_p \nabla (\overline{\mathbf{u}^T T})$ and $K_s(V_a)(T - T_o)$ is significantly linear (cross-correlation values exceeding the 95% confidence level). a) For $K_s(V_a)(T - T_o) = 40 \cdot (T - T_o)$. b) For $K_s(V_a)(T - T_o) = [V_a \cdot (T - T_o) + V_a \cdot (T - T_o)]$. Fig. 2.3a shows that in a relatively small area (which is shaded) of the North Atlantic, the linear relation seems appropriate. Using least squares, a value of $g_1$ can be estimated as $g_1 = 3 \times 10^{-4} \, m^{-1}$. In reality, the wind speed $V_a$ is not constant and accordingly, also the value of $K_s$ is not constant. In this slightly more realistic case, a rather similar result (Fig. 2.3b) is found, but with a larger region in which the linear relation is significant. It holds best in the region where the mean warming through convergence of the eddy fluxes is largest. In both cases, $g_1$ has the same order of magnitude and we use $3.0 \times 10^{-4} \, m^{-1}$ as the standard value for $g_1$. Using this value for $g_1$, $g_1 K_s(V_a)(T - T_o)$ accounted for 30% (averaged over the shaded areas in Figs. 2.3a and b) of the variance of $-\rho_c C_p \nabla (\overline{\mathbf{u}^T T})$.

Figure 2.4: Estimated $\gamma_1$ (a) and $\gamma_2$ (b), using October-March anomalies of $T_o$, $T_a$ and $T$. 
thereby demonstrating the significance of the linear relation. Furthermore, it should be noted that although the implementation of the coupling strength between ocean and atmosphere is quite different from Goodman and Marshall (1999), it can be shown (from their equation (25) and our equations (2.3) and (2.6)) that the order of magnitude is the same.

The same data were also used to estimate the coefficients $\gamma_1$ and $\gamma_2$ in (2.7). The linear dependence of $T_a$ on the sum of $\gamma_1 T$ and $\gamma_2 T_a$ is significant over the entire domain and estimated values of $\gamma_1$ and $\gamma_2$ (using least squares) are shown in Fig. 2.4. The high values of $\gamma_1$ in the northwestern part of the domain are probably due to ice coverage, which decreases the wintertime SST anomalies. In the southern and eastern part of the domain, the values of $\gamma_1$ vary between 0.7 and 0.9. Over these areas, $\gamma_2$ varies considerable but it seems reasonable to take $\gamma_1 = 0.8$ and $\gamma_2 = 0.2$ as the standard values in the model.

Note that the estimated parameters are calculated for the anomalous response to an anomalous forcing and do not give proper relations between mean SST and the mean atmospheric circulation. In equations (2.1) - (2.8) standard values for these and all other parameters are used as given in Table 2.1, unless mentioned otherwise.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$f_0$</td>
<td>$1.0 \times 10^{-4}$ [s$^{-1}$]</td>
<td>$\gamma_1$</td>
<td>0.8 [-]</td>
</tr>
<tr>
<td>$\beta_0$</td>
<td>$1.6 \times 10^{-11}$ [ms$^{-1}$]</td>
<td>$\gamma_2$</td>
<td>0.2 [-]</td>
</tr>
<tr>
<td>$L$</td>
<td>$3.0 \times 10^6$ [m]</td>
<td>$\gamma_3$</td>
<td>$1.2 \times 10^{-7}$ [s$^{-1}$]</td>
</tr>
<tr>
<td>$g'$</td>
<td>$2.0 \times 10^{-2}$ [ms$^{-2}$]</td>
<td>$S$</td>
<td>$2.8 \times 10^{-6}$ [m$^2$s$^{-2}$Pa$^{-2}$]</td>
</tr>
<tr>
<td>$H_1$</td>
<td>$5.0 \times 10^2$ [m]</td>
<td>$p_a$</td>
<td>$5.0 \times 10^4$ [Pa]</td>
</tr>
<tr>
<td>$H_2$</td>
<td>$4.0 \times 10^3$ [m]</td>
<td>$\Delta p$</td>
<td>$5.0 \times 10^4$ [Pa]</td>
</tr>
<tr>
<td>$\rho_1$</td>
<td>$1.0 \times 10^3$ [kgm$^{-3}$]</td>
<td>$\rho_a$</td>
<td>1.2 [kgm$^{-3}$]</td>
</tr>
<tr>
<td>$r$</td>
<td>0 [s$^{-1}$]</td>
<td>$k_a$</td>
<td>$1.2 \times 10^{-6}$ [s$^{-1}$]</td>
</tr>
<tr>
<td>$A_o$</td>
<td>$1.76 \times 10^4$ [m$^2$s$^{-1}$]</td>
<td>$A_a$</td>
<td>$1.0 \times 10^8$ [m$^2$s$^{-1}$]</td>
</tr>
<tr>
<td>$H_m$</td>
<td>$1.0 \times 10^2$ [m]</td>
<td>$R$</td>
<td>$2.87 \times 10^2$ [J/(kgK)$^{-1}$]</td>
</tr>
<tr>
<td>$C_o^p$</td>
<td>$4.0 \times 10^3$ [J/(kgK)$^{-1}$]</td>
<td>$g_1$</td>
<td>$3.0 \times 10^{-4}$ [m$^{-1}$]</td>
</tr>
<tr>
<td>$C_p^a$</td>
<td>$1.0 \times 10^3$ [J/(kgK)$^{-1}$]</td>
<td>$K_T$</td>
<td>$5.0 \times 10^2$ [m$^2$s$^{-1}$]</td>
</tr>
<tr>
<td>$\alpha_a$</td>
<td>$3.4 \times 10^{-3}$ [mskg$^{-1}$]</td>
<td>$u_E, v_E$</td>
<td>0 [ms$^{-1}$]</td>
</tr>
</tbody>
</table>

Table 2.1: Standard values of the dimensional parameters.
2.3 Stability of the jet-jet system

2.3.1 Basic steady state

The model as presented in the previous section is (by far) inadequate to model the coupled ocean-atmospheric mean state. Hence, a mean state has to be constructed in such a way that it is a solution of the governing equations. This is done by first prescribing the atmospheric basic state velocities as a zonal jet profile of the form

\[ \bar{u}_{a1}(y) = u_{a1,m} \sin(\frac{\pi y}{L}) \] (2.9a)

\[ \bar{u}_{a2}(y) = u_{a2,m} \sin(\frac{\pi y}{L}) \] (2.9b)

with \( u_{a1,m} = 20 \text{ ms}^{-1} \) and \( u_{a2,m} = 10 \text{ ms}^{-1} \). When these expressions are substituted into the governing equations with \( g_1 = \gamma_2 = 0 \), in which case there is no impact of SST anomalies on the atmospheric flow, an expression for \( H_a \) results. When this expression is subsequently used in the equations, the profiles (2.9) are – by construction – a stationary solution of the uncoupled atmosphere.

The structure of the atmospheric jet is coupled to that in the ocean through the wind stress. Using (2.4) with \( \gamma_3 = 3.3 \times 10^{-10} \text{ s}^{-1} \) and (2.9), this is accomplished by directly solving the stationary oceanic state in (1), using the methods outlined in the appendix. Note that using a more realistic windstress parameter for the steady state would give an unrealistically high ocean jet, because there are no eastern and western boundaries. In this way, the amplitude of the oceanic upper layer jet velocity \( \bar{u}_{o1} \) depends on the lateral friction coefficient \( A_o \). As no forcing on the lower ocean layer is present at steady state, it is at rest, i.e. \( \bar{u}_{o2} = 0 \). Finally, the temperature (in °C) within the surface layer is prescribed as

\[ \bar{T}(y) = 11.25 + 7.5 \cos(\frac{\pi y}{L}) \] (2.10)

and the forcing \( \bar{Q}_{oa} \) is determined, such that (2.10) is a solution of the SST equation using the previously computed oceanic fields. Profiles of the velocity of the lower atmosphere, upper ocean and the SST fields are shown in Fig. 2.5 for \( A_o = 1.76 \times 10^4 \text{ m}^2\text{s}^{-1} \). For this value of \( A_o \), the maximum velocity of the upper ocean jet is 17.2 cm s\(^{-1}\).

2.3.2 Linear stability

In a linear stability analysis, perturbations \( \tilde{\psi}_{a1}, \tilde{\psi}_{a2}, \tilde{\psi}_{o1}, \tilde{\psi}_{o2} \) and \( \tilde{T} \) are superposed on the basic state. Subsequently, the equations are linearized into these quantities and a normal mode approach is pursued, i.e. solutions of the form

\[ \tilde{\phi}(x, y, t) = \tilde{\phi}(x, y)e^{\sigma t} \] (2.11)

are considered, where \( \phi \) is any of the quantities above and \( \sigma = \sigma_r + i\sigma_i \) is the complex growth factor. The discretized problem on a 33 × 33 horizontal equidistant grid in the channel leads to an algebraic generalized eigenvalue problem; methods to solve this problem are discussed in the appendix.
Low-frequency variability in the jet-jet flow

Figure 2.5: Meridional profile of the basic state zonal velocities (solid) of upper atmosphere (scaled with its maximum velocity of 20 ms$^{-1}$), lower atmosphere (scaled with its maximum velocity of 10 ms$^{-1}$) and upper ocean (scaled with its maximum velocity of 17.2 cms$^{-1}$); all these three fields have the same shape. The dashed curve is the basic state sea-surface temperature profile (scaled with its maximum of 18.75$^\circ$C).

Note that through the construction of the basic state, the effect of SST on the atmosphere is only through anomalies and hence the standard values of $g_1$ and $\gamma_2$ are used. Similarly, velocity anomalies in the atmospheric lower layer provide a perturbation wind stress to the upper ocean with the standard value of $\gamma_3$ (Table 2.1). In the first set of results, we fix the value of $g_1$ and vary the value of the oceanic lateral friction $A_o$. Note that this parameter controls both the strength of the oceanic upper jet as well as the (viscous) dissipation of perturbations on the ocean.

For $A_o = 1.0 \times 10^5$ m$^2$s$^{-1}$, the basic state is stable, i.e. the real part $\sigma_r < 0$ for every eigenvalue. When $A_o$ is decreased, the first instability occurs at $A_o = 1.76 \times 10^4$ m$^2$s$^{-1}$. The patterns of the mode which destabilizes the basic state are shown for one phase of the oscillation in Fig. 2.6. The oscillation period ($P = 2\pi/\sigma_i$) is about 50 years at criticality and the mode is westward propagating. The ocean perturbations have a baroclinic character and the largest amplitudes of the anomalies are found in the regions of large vertical shear which occurs near the jet axis (Fig. 2.6c-d). The pattern of the SST perturbation (Fig. 2.6e) is controlled by that of the upper ocean through advection. The phase difference between the perturbation structure in the upper and lower atmosphere layer is about 5/16 times the total period. This particular phase difference results from a combination of a purely baroclinic response to SST forcing and an equivalent barotropic stationary Rossby wave in the atmosphere. Sensitivity test indicate that the stability of the basic state is hardly affected by reasonable variations in $\gamma_2$; non-local heat flux effects seem to be relatively unimportant here.

When $A_o$ is further decreased, more westward propagating modes become unstable. For $A_o = 1.5 \times 10^4$ m$^2$s$^{-1}$, $A_o = 1.45 \times 10^4$ m$^2$s$^{-1}$ and $A_o = 1.36 \times 10^4$ m$^2$s$^{-1}$, modes with $P = 8$ years, $P = 10$ years and $P = 7$ years destabilize, respectively. For smaller values
Stability of the jet-jet system

Figure 2.6: Perturbation structures of the large-scale mode at one phase of the oscillation. a) $\tilde{\psi}_1$, b) $\tilde{\psi}_2$, c) $\tilde{\psi}_3$, d) $\tilde{\psi}_4$ and e) $\tilde{T}$. In these and the following contour plots in Figs. 2.7, 2.10, 2.11, 2.12 and 2.13, all fields are scaled with their maximum and (non-dimensional) contour levels are with respect to this maximum.

of $A_o$, these three modes are all unstable and patterns of the streamfunction perturbations (at one phase of the oscillation) in both ocean layers (at $A_o = 1.36 \times 10^4 \ m^2s^{-1}$) are plotted in Fig. 2.7. Clearly, these modes are substantially different from the one plotted in Fig. 2.6 in that they have much smaller spatial scale. Either a symmetric (Fig. 2.7a-b and Fig. 2.7e-f) or an anti-symmetric (Fig. 2.7c-d) mode occurs and for all three modes the upper and lower ocean streamfunction perturbations are out of phase.
Figure 2.7: Perturbation streamfunctions $\tilde{\psi}_{o1}$ and $\tilde{\psi}_{o2}$ of the second (a,b), third (c,d) and fourth (e,f) most unstable mode. Left panels (a,c,e) show $\tilde{\psi}_{o1}$, right panels (b,d,f) show $\tilde{\psi}_{o2}$.

Hence, at standard value of $g_1$ and $A_0 = 1.36 \times 10^4 \text{ m}^2\text{s}^{-1}$, four modes are unstable. Next, the change in growth factor $\sigma_\rho$ of the two most unstable modes (the mode in Fig. 2.6 and the symmetric mode in Fig. 2.7a-b) is computed as the coupling strength is changed (with fixed $A_o$). The mode with small spatial scale (Fig. 2.7a-b) is stabilized when the coupling with the atmosphere is increased, whereas the large-scale mode (Fig. 2.6) is destabilized (Fig. 2.8). The other two unstable modes (Fig. 2.7c-f) also stabilize with increasing coupling (not shown). The oscillation period of both the small-scale modes and the large-scale mode decreases with increasing $g_1$. Note that the oscillation period of the large-scale mode also decreases with decreasing $A_o$ since it was 50 year at criticality ($A_o = 1.76 \times 10^4 \text{ m}^2\text{s}^{-1}$) and it is about 20 year at $A_o = 1.36 \times 10^4 \text{ m}^2\text{s}^{-1}$. 
Stability of the jet-jet system

Figure 2.8: Growth rate (solid lines) and oscillation period (dashed lines) versus the coupling parameter $g_1$ for $A_o = 1.36 \times 10^4 \, \text{m}^2\text{s}^{-1}$. a) for the large-scale mode and b) for the symmetric small-scale mode. The standard value of $g_1$ is indicated by the straight vertical lines.

In the limit of an uncoupled ocean-atmosphere system ($g_1 = 0$), the large-scale mode is damped (Fig. 2.8) and the modes in Fig. 2.7 are unstable. This clearly indicates that the destabilization of the large-scale mode is due to coupled processes. The instability originates from the coupling of a (near-)stationary atmospheric barotropic Rossby mode with a baroclinic ocean wave along a similar mechanism as described in Goodman and Marshall (1999) and Colin de Verdière and Blanc (2002). Basic theory (Holton, 1992) shows that in a one-layer free atmosphere, a barotropic Rossby wave with a wavelength of about 6000 km becomes stationary if the zonal jet velocity is about 15 ms$^{-1}$, which is similar to that of the basic state constructed here.
2.4 Transient flows in the nonlinear regime

The results of the previous section motivate to investigate the nonlinear transient flow for those values of $\lambda_0$ and $g_1$, where only the large-scale mode is unstable. It appears that for $\lambda_0 = 1.76 \times 10^4 \text{ m}^2 \text{s}^{-1}$ and $g_1 = 9 \times 10^{-3} \text{ m} \text{s}^{-1}$ (three times the standard value), only the large-scale mode is unstable and has a growth rate of 0.07 year$^{-1}$ and an oscillation period of about 45 years. For these values of parameters, the initial condition for the time integration is a superposition of the basic state and the spatial pattern of Fig. 2.6, the latter having a very small amplitude with respect to the basic state.

In the first 2000 years of the transient flow, the perturbation grows very slowly; for example, the maximum value of the perturbation field $\psi_2$ increases from $10^{-6}$ to $10^{-4}$ over this period. From this point, where for simplicity we define $t = 0$, time steps (dt) of approximately 1 year are taken until year 500 and 0.05 year from year 500 to year 1000. Note that these large time step can be taken, because implicit time integration methods are used (see appendix). For the particular gridpoint (16,8), which is located in the center of the channel in $x$-direction and halfway the center and the southern boundary in $y$-direction, the timeseries for the total quantities $\psi_{a1}, \psi_{a2}, \psi_{o1}, \psi_{o2}$ and $T$ are shown in Fig. 2.9. There are three stages of flow development, which are described below.

2.4.1 The growth of the coupled oscillation

Within the first few hundred years of the integration, a fairly regular oscillatory signal is seen in all fields with the oscillation period decreasing in time. The anomalous field (with respect to the time mean state over the years 510-530) is computed at year 520 and shown for the upper ocean layer streamfunction in Fig. 2.10. This pattern strongly resembles that of the large-scale mode (Fig. 2.6c) and indicates that this mode is indeed growing. The period of the oscillation is initially related to that determined from the linear stability analysis but it is strongly modified in time.

Due to the growth of this mode, rectification processes occur which cause changes between the basic steady state and the time mean state. Over the first 560 years, the upper atmosphere jet-maximum decreases from 20.0 to 19.5 ms$^{-1}$ and the lower atmospheric jet-maximum increases from 10.0 to 10.4 ms$^{-1}$. The upper ocean jet-maximum increases from 17.2 to 36.3 cms$^{-1}$ and a lower ocean jet arises with a maximum of 6.2 cms$^{-1}$. Furthermore, the mean state at year 560 indicates that the upper ocean jet has sharpened in meridional direction (Fig. 2.11c) with even small westward currents at the northern and southern boundaries. The lower ocean layer (initially at rest) now shows a zonal jet with a sharp meridional gradient (Fig. 2.11d). Also the atmospheric basic state is rectified by the nonlinear interactions of the perturbations, changing the amplitude of the zonal jets in the upper and lower atmospheric layer (Fig. 2.11a-b). Although a jet-like structure now also develops in the lower layer, the vertical shear in the ocean is increased due to a strong increase of the upper ocean jet; this effect is similar as a decrease of $A_0$ for the basic state would do. From the linear stability analysis, it was found that a reduction (increase) in $A_0$ decreases (increases) the period of oscillation. Hence, the changes in period along the transient flow can be explained by the change in the basic state due to rectification. At year 550 the period is reduced to a few years due to these rectification processes.
Figure 2.9: Timeseries of streamfunctions and SST at point (16,8): a) $\psi_a$, b) $\psi_b$, c) $\psi_c$, d) $\psi_d$ and e) $T$. 
Figure 2.10: Anomalous streamfunction of upper ocean layer at $t=520$, calculated as the anomaly with respect to the mean value [510, 530].

Figure 2.11: Mean zonal velocities averaged over the years 550 to 570, in (a) atmosphere layer 1, maximum is 19.5 ms$^{-1}$, (b) atmosphere layer 2, maximum is 10.4 ms$^{-1}$, (c) ocean layer 1, maximum is 36.3 cms$^{-1}$, (d) ocean layer 2, maximum is 6.2 cms$^{-1}$. 
2.4.2 Secondary instabilities

At a certain point in the time integration (near year 560), dramatic changes occur in the transient flow (Fig. 2.9), as the regular oscillatory time dependence suddenly comes to an end. High-frequency oscillations arise in each field and a transition to a new transient flow regime occurs. In order to see which patterns are associated with the high-frequency oscillations, the anomalous upper ocean streamfunction at year 580 (with respect to the time mean over the years 570-590) is plotted in Fig. 2.12. Although the perturbation structure does not resemble one of the individual small-scale modes as shown in Fig. 2.7, the pattern appears to be a combination of these modes. This indicates that the structure and amplitude of the jets and the SST field have changed by the rectification processes in such way, that the time mean state becomes unstable to the smaller-scale modes. Also the period of the small scale modes is substantially shortened due to the rectification processes which explains the high-frequency oscillations.

![Figure 2.12: Anomalous streamfunction of upper ocean layer at t=580, calculated as the anomaly with respect to the mean value [570,590].](image)

From the mean state at year 605 (averaged over the years 595 to 615), as shown in Fig. 2.13, it can be seen that the effect of these secondary instabilities on the mean state is opposite to that of the large-scale mode. The upper and lower oceanic jet and the lower atmospheric jet have decreased (compared to the mean state at year 560) whereas the upper atmospheric jet has increased in strength. Hence, the nonlinear interactions of these high-frequency modes show a rectification towards the original basic state. Another interesting point is that the jet-axes in ocean and atmosphere have moved little to the south (in Fig. 2.13 only visible for the lower ocean layer).

Hence, whereas the large-scale mode introduces a rectification towards a stronger upper ocean jet, the rectification through secondary instabilities weakens this jet. Then, the mean state is again susceptible for instability of the large-scale mode. Since rectification occurs on a much longer time scale than the growth of the instabilities, the results suggest a (potentially generic) mechanism of low-frequency variability in the coupled system. Indeed, irregular low-frequency behavior is found in years 600 to 1000 which are analyzed next.
2.4.3 Sustained low-frequency variability

Empirical Orthogonal Function (EOF) analysis has been performed to determine the dominant patterns associated with variability seen from year 600 to year 1000. The results reveal that low-frequency changes in the jet system account for almost all the variance in both atmosphere and ocean. In both atmospheric layers, the first EOF is a zonally stretched dipole which can be seen as a strengthening and weakening of the zonal jet (Figs. 2.14a and 2.15a). The first Principal Components (PC’s) of the upper and lower atmospheric layer (Figs. 2.14c and 2.15c) correlate well (lag-crosscorrelation coefficient is 0.99 and is highest for lag 0) and from the opposite signs of both EOF’s (Figs. 2.14a and 2.15a) it is clear that a strong (weak) upper atmospheric jet relates to a weak (strong) lower atmospheric jet.

The same type of pattern accounts for less of the variance in the ocean and is shown as EOF 2 in ocean layer 1 and 2 (Figs. 2.16b and 2.17b). Comparing the first PC’s of the atmospheric layers (Figs. 2.14c and 2.15c) with the second PC’s of the oceanic layers (Figs. 2.16d and 2.17d), it can be seen that the weakening (strengthening) of the lower (upper) atmospheric jet is related to a weakening of both ocean jets, and vice versa. Lag-crosscorrelation shows highest values (again exceeding 0.95) if the atmosphere leads the ocean by about half a year. These results indicate that the changes of the strength of the ocean and atmosphere jets are associated with an irregular oscillation with a period of 10 to 40 years.
The second EOF’s in upper and lower atmosphere (Figs. 2.14b and 2.15b) relate to the first EOF’s in the upper and lower ocean layers (Figs. 2.16a and 2.17a) and show a northward/southward shifting of the zonal jets. This meridional displacement of the jet was already noticed in subsection 2.4.2, but here it appears that this is also a recurrent feature, with an oscillation period of 20 to 50 years. Lag-cocorrelation (again exceeding 0.95) between atmospheric PC’s 2 and ocean PC’s 1 show highest values if the atmosphere leads the ocean by about 1.5 years.

As shown in Fig. 2.18, the variability in SST is dominated by high-frequency oscillations. The low-frequency behavior (weakening/strengthening and northward/southward shifting of the jets) is not reflected in SST. This is due to the purely zonal mean velocities in the upper ocean layer, whereas the mean zonal SST gradient is approximately zero. The EOF’s (also EOF’s 3 to 15, not shown) of SST seem to represent a combination of some smaller-scale...
waves associated with the small-scale modes shown in Fig. 2.7.

The third and fourth EOF of the upper atmosphere (Fig. 2.19a and Fig. 2.19b) show that the pattern associated with the large-scale unstable mode is still present from year 600 to 1000. EOF 3 and 4 taken together form a propagating large-scale wave, resembling the large-scale mode described in section 2.3. Furthermore, PC3 and 4 (Figs. 2.19c and 2.19d) show that this oscillation is now very irregular due to nonlinear interactions. The dominant oscillation period is 3.6 years, which is about the same as just before the rectification of the mean state due to the secondary instabilities (around year 560).
2.5 Summary and discussion

The midlatitude ocean-atmosphere system can be viewed as a highly nonlinear system with an uncertain degree of coupling. Various studies on the observed (coupled) variability (Sutton and Allen, 1997; Tourre et al., 1999; Da Costa and Colin de Verdière, 2002) find large-scale propagating anomalies with temporally correlating SST and SLP and a spatial phase difference between these quantities, suggesting that an equivalent barotropic atmospheric structure (Palmer and Sun, 1985) is prevailing at low frequency.

Within the intermediate-complexity model used here, with a strongly idealized representation of the coupling between ocean and atmosphere, linear stability analysis on a coupled oceanic/atmospheric jet system shows the existence of a large-scale coupled mode. This mode can become unstable when the coupling strength is large enough and introduces a preferential spatio/temporal pattern into the coupled system. The physics of the unstable mode is quite similar to that of unstable interactions (the horizontal advection case) as discussed
Low-frequency variability in the jet-jet flow

Figure 2.17: as figure 2.14, for lower ocean layer streamfunction: a) EOF 1 (92 %), b) EOF 2 (8 %), c) PC 1 and d) PC 2.

in Goodman and Marshall (1999): a near-stationary, large-scale barotropic Rossby wave in the atmosphere gives a positive feedback with a large-scale baroclinic ocean wave. Thus, the existence of this mode appears robust under a more detailed parameterization of the air-sea coupling and a more realistic spatial structure of the jets. The period and growth rate of the coupled mode is quite sensitive to changes in the amplitude of the basic jet in the ocean as well as to the strength of the air-sea coupling. Furthermore, smaller-scale oscillatory modes become unstable at standard coupling strength although these are stabilized when the strength of the coupling is increased.

Having the unstable coupled interdecadal mode, it is tempting to connect its pattern to the low-frequency variability as seen in observations, i.e. connect it to the NAO-variability. However, there is a large discrepancy between the pattern of this coupled mode (having a monopole structure) and the observed NAO dipole structure. Hence, one tends to conclude that the model is likely to be missing important physics, or that coupled modes play no role at all in the observed variability. Indeed, apart from the crude horizontal and vertical resolution,
other shortcomings of the model are the lack of continents on the eastern and western side of the basin and the lack of representation of the ocean gyres and upwelling. The oceanic Rossby waves propagating westward in the real ocean may have a different spatial and temporal structure.

However, the results from the nonlinear time-dependent flow show that the spatial and temporal structure of the coupled mode is not directly related to the low-frequency variability, but that its role is indirect. From the transient flow computed, it is found that nonlinear interactions of this coupled mode rectify the mean state in such a way that the period of the coupled oscillation decreases to interannual time scale. Furthermore, the rectified mean state becomes susceptible to the smaller-scale instabilities. These secondary instabilities introduce again rectification processes through which the mean state is changed towards its original state. The net result is an (inter-)decadal variability, in which the rectification processes of the mean state generate a weakening/strengthening and a meridional shifting of the atmospheric and oceanic jets. The exact spatial and temporal structure of the coupled mode and of the secondary instabilities is not crucial: one only needs two rectification processes to get the
Low-frequency variability through nonlinear rectification.

Although we have demonstrated this explicitly for only one case, the behavior appears fairly robust. Different initial conditions and nearby parameter values used gave qualitative similar results. Starting at a lower value of the lateral friction coefficient, $A_o = 1.36 \times 10^4 m^2 s^{-1}$, for which the smaller-scale waves are already unstable gives the same low-frequency variability. Starting with $A_o = 1.5 \times 10^4 m^2 s^{-1}$ and $g_1 = 3.0 \times 10^{-4} m^{-1}$ (standard coupling strength) also yields qualitatively the same results. However, in this case a lower percentage of the variance is explained by the changes of the jets (EOF 1 and 2, Fig. 2.14) and a higher percentage accounted for by the propagating wave (EOF 3 and 4, Fig. 2.19). This leads to a less clear extraction of different spatial patterns of variability through the EOF-analysis.

The nonlinear rectification processes lead to patterns of low-frequency variability, which qualitatively resemble the observed dominant patterns of variability (Cayan, 1992; Deser and Blackmon, 1993; Kushnir, 1994; Hurrell, 1995), with a dipole and monopole pattern varying quasi-oscillatory on an (inter-)decadal time scale. However, in our model results, the equivalent barotropic nature of the observed patterns in the atmosphere is not found, which may be
due to the simplified two-layer structure of the atmosphere model. Note that the instability of
the coupled mode is not central to the low-frequency variability: even if the large-scale mode
is stable, its pattern could be exited by atmospheric noise and the corresponding rectification
processes would still occur. This mechanism adds to others suggested from observational
(Deser and Blackmon, 1993; Kushnir, 1994; Sutton and Allen, 1997; Wallace et al., 1990)
and model (Goodman and Marshall, 1999; Latif and Barnett, 1994; Grötzner et al., 1998;
Selten et al., 1999; Cessi, 2000; Primeau and Cessi, 2001)) studies. Certainly, additional
analysis in more elaborated models is needed but the results suggest that nonlinear dynamics
may dramatically change the nature of the variability. Hence, one should be careful in com-
paring linear model results with observations and in using linear explanations of variability
in a complex (nonlinear) GCM.
Chapter 3

Low-frequency variability in the jet-gyre flow

The low-frequency variability in the coupled ocean-atmosphere system is explored, again in a model of intermediate complexity. However, here the idealized oceanic background state is characterized by a double-gyre ocean circulation either in a one- or two-layer ocean configuration. Linear stability analysis on the coupled system with the two-layer ocean model does not show any low-frequency modes. Transient flows show nonlinear rectification processes of intermonthly internal ocean modes leading to generic low-frequency variability. Coupling with the atmosphere damps this low-frequency variability. Both the linear stability analysis and transient flows of the uncoupled system in the one-layer configuration show an internal ocean mode with a period of about a year. From the literature, this mode is known to exhibit a decadal period in multi-layer ocean models. The effect of coupling on this mode is examined but these results are not conclusive.

3.1 Introduction

In the previous chapter it was shown that an atmospheric jet, overlying an oceanic jet can lead to low-frequency variability in the coupled system. The jet-jet system is linearly unstable to a large-scale coupled mode: a mixed barotropic/baroclinic ocean wave interacts with a near-stationary barotropic atmospheric Rossby wave. The decadal period of this mode decreases to interannual time scale in a finite-amplitude, time-dependent flow and also secondary instabilities appear. This nonlinear evolution leads to low-frequency variability, due to different rectification processes of the time mean flow.

One of the obvious idealizations in the model of chapter 2 is the zonally unbounded ocean. It is well known that continental boundaries severely restrict the mean ocean circulation patterns and wave propagation. With the implied wind-stress pattern from the atmospheric jet, a so-called double-gyre circulation appears, instead of the zonal jet. The variability in the jet-gyre system, arising through instabilities of steady flows is the focus of this chapter. The methodology will follow the same path as chapter 2: first the linear stability of the uncoupled
and coupled ocean-atmosphere system is examined and then computation of transients shows the nonlinear evolution. In this analysis the main goal is to address the following questions: (i) can we find a similar coupled low-frequency mode as in the jet-jet system, (ii) can we find low-frequency variability in the jet-gyre system due to rectification of the time mean flow as was found in chapter 2, and (iii) are there additional mechanisms leading to low-frequency variability, not found in the jet-jet system?

The analysis in this chapter is an extension of work done on the stability of double-gyre, oceanic flows (Dijkstra and Katsman, 1997; Katsman et al., 1998; Nauw and Dijkstra, 2001) by including the coupling to the midlatitude atmosphere. In Dijkstra and Katsman (1997) the stability of a double-gyre, quasi-geostrophic ocean circulation is examined in an ocean basin of $1000 \times 1000 km$. Starting from a zero-flow situation in a one-, 1.5- or two-layer ocean, Dijkstra and Katsman (1997) examine how successive bifurcations occur when the lateral friction ($A_o$) is decreased, thereby increasing the strength of the double-gyre flow. In the two-layer model configuration, the most unstable modes are mixed barotropic/baroclinic waves, having an intermonthly time scale. The spatial structures resemble the symmetric and asymmetric modes, found in section 2.3, however, with smaller spatial extension, limited to the double-gyre recirculation area. In the one-layer model configuration, when $A_o$ is decreased, the first bifurcation to occur is a symmetry-breaking pitchfork bifurcation, leading to multiple equilibria. Concerning low-frequency variability, a very interesting low-frequency mode becomes unstable if $A_o$ is decreased further. This interannual mode is called the "gyre" mode, because its structure is strongly related to the asymmetric double gyre structure.

Nauw and Dijkstra (2001) find similar results in a shallow-water model with a larger ocean basin (1000 km in x-direction, 2000 km in y-direction). However, in the shallow-water model, the gyre mode has a period of 12.7 years. Simonnet and Dijkstra (2002) investigate the fundamental physics of the gyre mode and show that the gyre mode results from the conjugate effects of shear- and symmetry-breaking instabilities. The mechanism turns out to be very robust over a hierarchy of models (Simonnet and Dijkstra, 2002). Apart from the coupled mode as described in chapter 2 and the rectification mechanisms, the gyre mode may be another source of low-frequency variability in the midlatitude coupled system.

In Katsman et al. (1998) the time-dependent behavior of the linearly unstable double-gyre circulation described in Dijkstra and Katsman (1997) is examined. Nonlinear interactions of the first intermonthly mode with itself and with the mean state showed a rectification of the mean state, giving a new stationary state, differing from the initial steady state.

Although performed in smaller basins and not coupled to a dynamically active atmosphere, the studies of Dijkstra and Katsman (1997), Katsman et al. (1998) and Nauw and Dijkstra (2001) clearly show directions in which the answers to the main questions stated above might be found. For instance, it has become clear that in the quasi-geostrophic model we use here, both a one-layer and two-layer ocean model configuration must be considered.

First, we will describe the model set-up in section 3.2. The results obtained for a two- and one-layer ocean model are presented in sections 3.3 and 3.4, respectively. In each case, the linear stability of the jet-gyre system is examined and transient flows are analyzed. A summary and discussion of the results follows in section 3.5.
3.2 Model formulation

Similar to the model described in chapter 2, again two dynamically active layers in the atmosphere are coupled with a two-layer ocean, through a constant depth surface layer embedded in the upper ocean layer (Fig. 3.1). A channel set-up having a width \( L = 3000\, \text{km} \) is chosen for the atmosphere. The ocean basin is now closed and has both width and length \( L \). The ocean, surface temperature and atmosphere (thermo-)dynamics are thereby described again by equations (2.1-2.8). Thus, the main difference with the model described in the previous chapter is that here, in the ocean at all lateral boundaries, slip boundary conditions are used. The equations are again solved on a \( 33 \times 33 \) grid, however, because most oceanic features examined here will be localized in the small region where ocean flows are largest, the grid is now stretched both in \( x \)-and \( y \)-direction. The non-equidistant grid \((x, y)\) is calculated from an equidistant grid \((\bar{x}, \bar{y})\) through

\[
x = 1 - \frac{\tanh(q_x (1 - \bar{x}))}{\tanh(q_x)} \tag{3.1a}
\]

\[
y = \frac{1}{2} \left[ 1 + \frac{\tanh(q_y (1 - \bar{y})) - \tanh(q_y (1 - \bar{y}))}{\tanh(q_y)} \right] \tag{3.1b}
\]

using stretching parameters \( q_x = q_y = 2 \). Unless mentioned otherwise, the same parameter values are used as in chapter 2 (Table 2.1).
3.3 Two-layer double-gyre ocean flow

3.3.1 Linear stability

In a similar way as in section 2.3 a basic steady state is constructed. However, in constructing an atmospheric flow which drives a reasonable double-gyre ocean circulation in the upper ocean layer, we have chosen a slightly different shape of the jet. By prescribing

\[ u_{a1}(y) = -u_{a1,m} \cos\left(\frac{2\pi y}{L}\right) \]
\[ u_{a2}(y) = -u_{a2,m} \cos\left(\frac{2\pi y}{L}\right) \]

with \( u_{a1,m} = 20 \, \text{ms}^{-1} \) and \( u_{a2,m} = 10 \, \text{ms}^{-1} \), an atmospheric jet results with westward velocities at the northern and southern boundary and an eastward velocity in the center. This basic state is shown in the Figs. 3.2a and b. Then, using the methods outlined in appendix A, the oceanic steady states are computed as a function of parameters. Again, \( \gamma_3 \) and \( A_o \) determine the shape and amplitude of the upper ocean layer streamfunction. For \( \gamma_3 = 3.5 \times 10^{-8} \, \text{s}^{-1} \) and \( A_o = 1.0 \times 10^4 \, \text{m}^2 \, \text{s}^{-1} \), this streamfunction is plotted in Fig. 3.2c, while the lower ocean layer is unforced and hence motionless (Fig. 3.2d). The SST-pattern for zero flow is constructed as in equation (2.10), giving a constant meridional SST gradient. For a nonzero flow, as shown in Fig. 3.2c, the steady state for SST in the surface layer is shown in Fig. 3.2e, for a value of \( K_T = 3.0 \times 10^3 \, \text{m}^2 \, \text{s}^{-1} \). The latter is a minimum value, restricted by the horizontal resolution of the model.

Using the methods outlined in appendix A, the linear stability of the steady flows is examined using different values for the wind-stress parameter \( \gamma_3 \) and the coupling-strength parameter \( g_1 \). First, we look at the uncoupled system, with \( g_1 = 0 \). By increasing \( \gamma_3 \) from \( 2.7 \times 10^{-8} \, \text{s}^{-1} \) to \( 3.7 \times 10^{-8} \, \text{s}^{-1} \), the maximum zonal velocity in the upper ocean flow increases from 48 to 78 \( \text{cm} \, \text{s}^{-1} \), with an upper ocean streamfunction pattern closely resembling Fig. 3.2c. Due to these increasing upper ocean velocities, the flow becomes unstable to 3 oscillatory modes, which are labeled E1, E2 and E3. The growth rate \( \lambda \) and period \( P \) of these modes are found in Table 3.1 for four different values of \( \gamma_3 \). The first mode becomes unstable if \( \gamma_3 \) is about \( 3.0 \times 10^{-8} \, \text{s}^{-1} \), the second and third if \( \gamma_3 \) is about \( 3.3 \times 10^{-8} \, \text{s}^{-1} \). It is clear from Table 3.1 that these modes all destabilize with increasing wind-stress parameter \( \gamma_3 \). The period of these modes ranges from a few months to 1.4 years. The spatial structures of the modes E1, E2 and E3 for upper and lower ocean layer streamfunction are shown in Figs. 3.3c-d, 3.4c-d and 3.5c-d, respectively (actually, these figures show these modes for \( g_1 = 3.0 \times 10^{-4} \, \text{m}^{-1} \), but the spatial patterns very closely resemble those for \( g_1 = 0 \)). For the case \( g_1 = 0 \), the amplitudes of the atmospheric streamfunction perturbation of E1, E2 and E3 are zero, thereby indicating that these modes are internal ocean modes of variability. The same type of oscillatory modes were found in Dijkstra and Katsman (1997), however, in a much smaller basin. Furthermore, one should note the resemblance of these modes with the “smaller scale modes” found in section 2.3, Fig. 2.7. A clear difference between the type of modes shown in Fig. 2.7 and those found here is that in the latter case the spatial structures are restricted to the western part of the basin, where the ocean velocities are largest. Furthermore, this smaller spatial scale is accompanied by a smaller oscillation time scale. Similarities between the internal ocean modes in the jet-jet
Figure 3.2: Basic steady state of a) $\overline{\psi}_u$, b) $\overline{\psi}_w$, c) $\overline{\psi}_t$, d) $\overline{\psi}_o$ and e) $T$. Fields are scaled with their maximum. Arrows in a) and b) denote direction of flow. Maximum zonal velocities are $20 \text{ m s}^{-1}$, $10 \text{ m s}^{-1}$ and $72 \text{ cm s}^{-1}$ for upper and lower atmosphere layers and upper ocean layer, respectively. Maximum temperature is $18.75 \degree C$. 
these results that internal ocean processes dominate the stability of the first unstable modes.

A small, but robust effect of coupling on the growth rate and period of E1-E3 (Table 3.1). It shows that an increased coupling strength destabilizes E1, but stabilizes E2 and E3. Although a clear difference between E1 on the one hand and E2 and E3 on the other is the (a-)symmetry with respect to the center of the basin (in y-direction), we can not identify the precise physics for the opposite effect of coupling on the stability of the modes. In fact, the large difference in spatial extension between the atmosphere and ocean makes it hard to find a reason for the different behavior of the modes with coupling. However, it can be concluded from these results that internal ocean processes dominate the stability of the first unstable modes.

The spatial difference between the “small-scale” ocean perturbations and the “large-scale” atmospheric perturbations is due to the large difference in oceanic and atmospheric Rossby deformation radius. Examining the stability of this system for 3 different values of wind stress coefficient

Table 3.1: Growth rate \( \lambda (\text{year}^{-1}) \) and period \( P \) (year) for different values of wind stress coefficient \( \gamma_3 (\text{s}^{-1}) \) and coupling strength \( g_1 (m^{-1}) \) for first unstable modes (E1), (E2) and (E3). If \( \lambda < 0 \), this is denoted by "-".

<table>
<thead>
<tr>
<th>( \gamma_3 )</th>
<th>( g_1 )</th>
<th>( \lambda(E1) )</th>
<th>( P(E1) )</th>
<th>( \lambda(E2) )</th>
<th>( P(E2) )</th>
<th>( \lambda(E3) )</th>
<th>( P(E3) )</th>
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<tr>
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<td>7.2</td>
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<td></td>
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<td>7.8</td>
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</tr>
<tr>
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<td>1.6</td>
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<tr>
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Figure 3.3: Perturbation structures of the first unstable mode. a) $\psi_{a1}$, b) $\psi_{a2}$, c) $\psi_{o1}$ d) $\psi_{o2}$ and e) $\bar{T}$. Fields are scaled with their maximum.
Figure 3.4: Perturbation structures of the second unstable mode, a) $\tilde{\psi}_{a1}$, b) $\tilde{\psi}_{a2}$, c) $\tilde{\psi}_{o1}$ d) $\tilde{\psi}_{o2}$ and e) $\tilde{T}$. Fields are scaled with their maximum.
Figure 3.5: Perturbation structures of the third unstable mode. a) $\tilde{\psi}_1$, b) $\tilde{\psi}_2$, c) $\tilde{\psi}_3$, d) $\tilde{\psi}_4$ and e) $\tilde{T}$. Fields are scaled with their maximum.
in the coupled jet-gyre system and that coupling with the atmosphere causes a large-scale atmospheric response, which hardly affects the growth rate and period of the oceanic modes.

When increasing the strength of the ocean circulation (by further increasing $\gamma_3$), many more internal ocean modes become unstable, both for the uncoupled as for the coupled case (not shown). All these modes show a large resemblance with those described above: small-scale ocean and SST pertubations centered in the western part of the basin, resembling the characteristics of Figs. 3.3c-e, 3.4c-e and 3.5c-e, large-scale, baroclinic perturbations in the atmosphere and an oscillation period ranging from a few months to about 1.5 years. Some of these modes stabilize, some destabilize with increased coupling. As the qualitative behavior of all these modes is similar, for brevity we have only shown details of the first 3 unstable modes.

It can be concluded that none of the modes described above resembles the large-scale mode described in section 2.3. First, no large-scale (spatial or temporal) ocean modes were found. Second, none of the modes found here shows an equivalent barotropic structure in the atmosphere. It seems that the zonal extension of the eastward jet at the subtropical/subpolar gyre interface is too small to be unstable to such a large-scale mode. Furthermore, the time scale of the internal ocean modes found here is relatively small, reducing the possibility of spatial and temporal "matching" of the atmospheric and oceanic variability, as described in the previous chapter and by Weng and Neelin (1998) and Neelin and Weng (1999).

### 3.3.2 Transient flows

In the previous subsection, it was shown that in the linear stability analysis no low-frequency modes were found in the coupled jet-gyre system. However, in the previous chapter it was shown that rectification processes of the time-mean state lead to low-frequency variability. Furthermore, as mentioned in section 3.1 Katsman et al. (1998) have shown that in a smaller ocean nonlinear evolution of an intermonthly mode indeed does rectify the mean state. Therefore, it is worthwhile to investigate the nonlinear evolution of the jet-gyre system, even though no low-frequency modes were found in the previous subsection. For this purpose, a 100-year time integration was performed. The initial condition for the time integration is a superposition of the basic state, shown in Fig. 3.2 and the spatial pattern of Fig. 3.3, the latter having a very small amplitude with respect to that of the basic state. In the previous subsection it was shown that on this basic state 3 modes were unstable both for the uncoupled as the coupled system. Therefore, in the first 50 years of the integration, the uncoupled system is examined ($g_1 = 0$) and at year 50, $g_1$ is put to its standard value of $3.0 \times 10^{-4} m^{-1}$, to show the additional effect of coupling. Fig. 3.6 shows the timeseries of the various fields.

In the oceanic upper and lower layer streamfunction and SST, clearly two different types of variability show up: a high-frequency signal on the time scale of months and a low-frequency signal with a period of about 6 years.

In order to explore the details of this variability, an EOF-analysis is performed on years 10-50 (leaving out the irregular first 10 years, in which the quasi-stationary state was reached). PC 1 of the upper ocean layer streamfunction, explaining 62% of the variance, shows that a regular, 6-year period oscillatory signal dominates the variability (Fig. 3.7e). EOF 1 reveals a weakening/strengthening of the double-gyre circulation (Fig. 3.7a). As no such low-frequency behavior was found in the linear stability analysis, we argue that this
variability in the flow is caused by rectification processes, similar to the low-frequency behavior described in section 2.4. Additional support for this comes from EOF 2-4, explaining 14%, 11% and 9%, respectively. In these EOF’s (Fig. 3.7), the spatial and temporal structures of E1 (EOF 3/4) and E3 (EOF 2) are recognized. It is clear from the behavior of PC 2-4 (Figs. 3.7f-h), that the amplitude of the high-frequency modes is directly related to the low-frequency behavior of PC 1 (Fig. 3.7e). The growing unstable modes E1 and E3 (E2 being less important in the time-dependent behavior) seem to rectify the time-mean flow, increasing the double-gyre circulation. On this changed circulation, the stability of the different
The period of the mode captured by EOF 3/4 decreases slightly and the behavior of the mode captured by EOF 2 becomes more irregular (see e.g. years 12.5-15). The changed high-frequency signal causes an opposite rectification of the time-mean flow, pushing the gyre circulation towards its initial state. Thus, where the linear stability analysis did not show low-frequency variability, the time integration does show such behavior, due to rectification processes.

Two major differences can be identified between the results presented here and those obtained in section 2.4:

1) In section 2.4 it seemed that two types of modes on a very different spatial scale caused opposite rectification of the mean flow, giving the low-frequency oscillating signal. However, this scale difference of the modes at work does not seem crucial for the nonlinearly induced low-frequency variability, because in this section only “small-scale” modes are unstable.

2) In section 2.4 the nonlinearly induced variability showed two different characteristics: a meridional shifting and a weakening/strengthening of the eastward jet. Here, we only find the latter and rectification processes due to large-scale perturbations may therefore be crucial for the meridional shifting.

The timeseries of the coupled time integration, years 50-100, are also shown in Fig. 3.6. In the ocean fields, a similar variability as in the first 50 years is present. However, clearly the low-frequency variability changes. The amplitude seems to decrease in time and the period increases from about 6 years (in the first 50 years) to about 10 years. These features are confirmed in the EOF-analysis on the upper ocean layer streamfunction (Fig. 3.8). Simultaneously, the amplitude of the high-frequency oscillatory mode resembling E1 becomes more regular and smaller with time (PC’s 3/4, Figs. 3.8g/h). Thus, the nonlinearly induced low-frequency variability is slowly damped through coupling with the atmosphere and the system evolves towards a stationary mean situation with an additional constant amplitude oscillation, resembling E1. The behavior of E2 (captured by EOF 2 in the years 10-50) becomes rather strange after year 50. Whereas the spatial structure remains almost unchanged (Fig. 3.8b), the period has quickly changed (Fig. 3.8f) to the same frequency as the dominant low-frequency signal (Fig. 3.8e). Phase-locking could be a possible explanation. The linear stability analysis showed that E2 was slightly damped through coupling (Table 3.1) and indeed the time dependent signal of this mode damps out. So, it appears that the small effect of the atmosphere on the internal high-frequency ocean modes in the linear stability analysis becomes crucial in the nonlinear time evolution. By damping out a part of the high-frequency variability it takes out an essential component in the rectification process, thereby damping out the nonlinearly induced low-frequency variability.

EOF-analysis was also performed on the timeseries of the streamfunction of the upper and lower atmosphere layer, which, according to Figs. 3.6a and b, show variability related to the ocean variability, with both a high- and a low-frequency signal. As the atmospheric variability in both layers appeared to be almost purely baroclinic, we only show the results for the upper layer streamfunction. EOF 1 (Fig. 3.9a) shows the low-frequency oscillation, explaining almost all (99%) of the variance. The pattern is a zonally stretched dipole, representing a weakening/strengthening of the eastward jet. Thus, similar to the ocean variability, also the atmospheric low-frequency behavior resembles the results found in the jet-jet-system (again apart from the meridional shifting of the jet). Both the atmospheric dipole structure and the decreasing amplitude in time show that this variability is the atmospheric response to the
internal ocean variability, EOF's 2-4 (not shown) together explain only 1% of the variance, showing that the atmospheric response to the high-frequency ocean variability is very weak.

From the nonlinear time evolution of this double-gyre ocean circulation both uncoupled and coupled to the atmosphere, one can conclude that similar to the linear stability analysis, no large-scale low-frequency coupled ocean-atmosphere mode, resembling the one from section 2.3 is found. Nevertheless, a combination of internal oceanic high-frequency modes

Figure 3.7: Results of EOF analysis on year 10-50 of the timeseries of oceanic upper layer stream-function: a) EOF 1, explaining 62% of the variance, b) EOF 2, (14%), c) EOF 3, (11%), d) EOF 4 (9%), e)-h) PC 1-4. Arbitrary units in a)-d) and on the vertical axis in e)-h). Horizontal axes in e)-h) show time (years).
rectify the time mean uncoupled flow in such way that low-frequency variability is present. However, this low-frequency variability is damped through coupling with the atmosphere. A further discussion on how these results relate to the previous chapter and how they should be interpreted in comparison to observations follows in section 3.5.
Figure 3.8: Results of EOF analysis on year 50-100 of the timeseries of oceanic upper layer streamfunction: a) EOF 1, explaining 40% of the variance, b) EOF 2, (25%), c) EOF 3, (15%), d) EOF 4 (13%), e)-h) PC 1-4. Arbitrary units in a)-d) and on the vertical axis in e)-h). Horizontal axes in e)-h) show time (years).
Figure 3.8 (continued)
Figure 3.9: Results of EOF analysis on year 50-100 of the timeseries of atmospheric upper layer streamfunction: a) EOF 1, explaining 99 % of the variance, b) PC 1. Arbitrary units in a) and on the vertical axis in b). Horizontal axis in b) shows time (years).
3.4 One-layer double-gyre ocean flow

3.4.1 Linear stability

In the particular case where \( \delta^o = 0 \) and in the limit \( \frac{1}{\gamma} = 0 \), in equation (2.1), the ocean model reduces to a one-layer quasi-geostrophic barotropic vorticity equation. The basic steady state is constructed in a similar way as in the previous section. The atmospheric basic state is similar to the one shown in Figs. 3.2a and b. Using \( A_o = 1.5 \times 10^4 m^2 s^{-1} \) and \( \gamma_3 = 7.6 \times 10^{-8} s^{-1} \), the basic state ocean circulation and SST-pattern resemble the spatial structures shown in Figs. 3.2c and e. By decreasing the lateral friction and solving the basic state directly using the methods described in appendix A, the ocean velocities will increase. For subsequent basic states, with increasing ocean velocity, the stability is examined again. The stability characteristics of the system, ocean and atmosphere both uncoupled as coupled, qualitatively strongly resemble those described by Dijkstra and Katsman (1997), see section 3.1. The first transition (for \( A_o = 9.6 \times 10^3 m^2 s^{-1} \)) is a pitchfork bifurcation, breaking the symmetry of the ocean circulation and introducing multiple steady states for lower values of \( A_o \) (Dijkstra and Katsman, 1997). For \( A_o = 5.1 \times 10^3 m^2 s^{-1} \), one of the two multiple steady states is shown in Fig. 3.10, for ocean streamfunction (note that we have only one layer in this case) and SST-pattern (atmospheric basic state is unchanged). For brevity, we only show this basic steady state, which we will call the ”jet-up” situation (referring to the northward shifted eastward jet between both gyres), and do not show the ”jet-down” situation. This choice is

![Figure 3.10: Basic steady state of a) \( \psi_{\omega_1} \) (maximum zonal velocity is 2.7ms\(^{-1}\)) and b) \( T \). Fields are scaled with their maximum.](image-url)
arbitrary because all stability characteristics are similar on both steady states.

The basic state shown in Fig. 3.10 is unstable for three oscillatory modes. Two of these modes are so-called barotropic Rossby basin modes, showing a large spatial scale and a very high frequency (\(< 1\) month). The oceanic spatial structure of these modes is shown in Fig. 3.11. The third oscillatory mode is also found by Dijkstra and Katsman (1997) and is called the ‘gyre’ mode. The perturbation structures for atmosphere, SST and ocean are shown in Fig. 3.12. Figs. 3.12a and b show that the atmospheric structure is large scale. Fig. 3.12c is similar for both an uncoupled and a coupled ocean-atmosphere system, showing that the mode is an internal ocean mode and the atmospheric structures are merely a response to the SST-pattern (Fig. 3.12d). In order to avoid resolution problems in SST, \(K_T = 3.0 \times 10^4\) m\(^2\) s\(^{-1}\) is used for the diffusivity parameter in this and the following analyses.

The effect of ocean-atmosphere coupling on the stability and frequency of the modes described above is examined, but is not very clear. Basin mode 1 stabilizes when the coupling strength increases, but basin mode 2 shows a nonlinear response, stabilizing with small coupling and destabilizing with strong coupling. It must be noted here, that the thermal diffusivity is quite unrealistic and thus the effect of coupling can be blurred by this. Because the basin modes are high-frequency and thus their possible importance for low-frequency variability can only be present in a time-integration (through rectification), we will not further discuss their characteristics in this subsection. Coupling strongly stabilizes the gyre mode, standard coupling already reducing the growth rate from \(0.7\) year\(^{-1}\) to value lower than zero (stable). A mechanism for this stabilizing effect is not found. It is clear that the period of the gyre mode, only about 8 months, is quite small, it is even smaller as the period found by Dijkstra and Katsman (1997). This is probably caused by a slightly different asymmetry of the mean state (Simonnet and Dijkstra, 2002).

From the stability analysis on the one ocean layer quasi-geostrophic model, coupled to

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**Figure 3.11:** Ocean streamfunction perturbation structure \(\bar{\psi}_{ol}\) of a) Rossby basin mode 1 and b) Rossby basin mode 2. Fields are scaled with their maximum.
the two layer atmosphere, it can be concluded that the most unstable modes of variability are caused by internal ocean variability. This ocean variability strongly resembles results of e.g. Dijkstra and Katsman (1997), Nauw and Dijkstra (2001), thereby showing that the size of the basin, only qualitatively changes the stability of this system. Although the spatial and temporal structure of the ocean low-frequency variability is hardly changed by the atmosphere, there is a stabilizing effect through coupling on the stability of the gyre mode. The associated atmospheric anomalies are large scale and show a baroclinic structure.
3.4.2 Transient flows

In order to examine the variability of the double-gyre circulation in the one-layer ocean model, uncoupled and coupled to the atmospheric jet, again a time integration is performed, partly with \( g_1 = 0 \) and partly with \( g_1 = 3.0 \times 10^{-4} \text{m}^{-1} \). Starting with the uncoupled system, the initial condition is a superposition of the basic state shown in Fig. 3.10 and the spatial pattern of Fig. 3.12, the latter having a very small amplitude with respect to the basic state. For the first 30 years of integration in which \( g_1 = 0 \), the timeseries of the upper ocean layer streamfunction is shown in Fig. 3.13. The previous subsection has shown that in this regime, the basic state is susceptible to three oscillatory instabilities, with periods shorter than a year. Therefore, a small time step of \( 1.0 \times 10^{-3} \text{ year} \) is taken. Fig. 3.13 shows that after a few years in which anomalies grow to significant amplitude, the system takes a while to equilibrate but then stays in a regular oscillatory situation. Two time scales of variability are clearly present. The high-frequency variability is caused by the basin modes and the lower frequency variations by the gyre mode (EOF’s are not shown because they closely resemble Figs. 3.11 and 3.12c). Important characteristics become clear from the timeseries:

1) Although the growth rates of the basin modes are higher compared to the gyre mode in the linear stability analysis, the gyre mode dominates the amplitude of the perturbations in the time dependent flow.

2) The characteristics of both the initial steady state as the linearly unstable modes on this steady state do not change significantly during the time integration: the time mean circulation pattern of years 20-30 (not shown) is similar to Fig. 3.10a and both the oscillation period and spatial structure of the basin modes and the gyre mode resemble the results of the linear stability analysis during the time integration. Therefore, it can be concluded that rectification processes are of minor importance here.

3) Nonlinearly induced low-frequency variability as found in the jet-jet system and in the jet-gyre system with a two-layer ocean model can not be found in the uncoupled one-layer model.

Because other studies have shown that the gyre mode is likely to have a multi-year period and the gyre mode is found in a complex GCM in chapter 4, we will further examine its variability in the time integration. One of the important characteristics of the gyre mode

![Figure 3.13: Timeseries of \( \psi_{o1}(10^5 \text{ m}^2\text{s}^{-1}) \) at gridpoint (16,15)](image_url)
is shown to be that it is not a standing oscillation, but that anomalies propagate through the basin (Nauw and Dijkstra, 2001; Simonnet and Dijkstra, 2002). In order to examine this propagation a Multichannel Singular Spectrum Analysis (MSSA) is performed on the timeseries of the upper ocean layer streamfunction. MSSA is a statistical technique designed to detect propagating patterns in noisy, relatively short timeseries (Plaut and Vautard, 1994). Similar to conventional EOF analysis, these patterns are ordered with respect to the amount

Figure 3.14: MSSA mode 4+5 of $\psi_1$, a) - d) are phase plots at 20.2, 20.3, 20.4 en 20.5 year and are characteristic for half a period of the oscillation.
of variance they account for. Mathematically, MSSA is equivalent to Extended EOF (EEOF) analysis (Weare and Nasstrom, 1982), but in MSSA focus is on the temporal structure of the variability, whereas in EEOF analysis it is on spatial variability. Further details on MSSA are described in appendix B. Performing MSSA on years 20-30, it is found that modes 4 and 5 represent the gyre mode. Figs. 3.14a-d show the patterns of the mode at four different time steps, together showing characteristics of half a period of oscillation of the gyre mode. It can be seen that indeed the mode resembles the gyre mode as found in the linear stability analysis (Fig. 3.12). Furthermore it can be seen that in all phases the anomalies are aligned to the north-eastward jet in the basic steady state (Fig. 3.10a). Following the subsequent phase plots in Fig. 3.14 shows that the anomalies propagate northward. A Hörmöller diagram of $\psi_{o1}$ at $x = 550 km$ (Fig. 3.15) clearly confirms this northward propagation of the anomalies.

![Hörmöller diagram at x = 550km](image)

**Figure 3.15**: MSSA mode 4+5 of $\psi_{o1}\left(10^5 m^2 s^{-1}\right)$: Hörmöller diagram at $x = 550 km$.

In the previous subsection it was shown that the basic state of Fig. 3.10 is one of the two multiple equilibria. The stability characteristics of the other basic state, the "jet-down" solution, are qualitatively similar to the results shown here, but the direction of propagation of the anomalies of the gyre mode is southward (not shown). More precisely, the anomalies propagate from the north-western part of the basin south-eastward, and at about $y = 1500 km$ they travel further in a south-westward direction. It might be tempting to relate the realistic double-gyre circulation to the "jet-up" solution here, because of its southwest-northeast orientation, but Schmeits and Dijkstra (2001) have shown that in a more complex model with realistic geometry the multiple steady states are more complicated. Therefore, it can not be concluded from the results whether a southward or northward propagating gyre mode (or perhaps both) will be present in reality.

Again, the effect of ocean-atmosphere coupling is examined. At year 30 the coupling strength $g_1$ is put at its standard value and in Fig. 3.16 the timeseries for the upper ocean streamfunction is shown for years 30-35. First, the coupling seems to have only little effect and the variability looks quite similar to the 10 years before (with $g_1 = 0$). However, then anomalies start to grow rapidly on both intra- and intermonthly time scale. The reason for this
is that the atmospheric response to the ocean anomalies reflects on the ocean mean circulation in such way that it destroys the asymmetry of the time mean circulation. In Fig. 3.17, the mean ocean circulation of year 34.5 to 35 is shown. Although not yet completely symmetric, it is clear that it does not resemble Fig. 3.10. Clearly, the ocean circulation becomes a more inertially driven flow, with totally different stability characteristics (Dijkstra and Katsman, 1997). This behavior is a well-known limitation of such a one-layer quasi-geostrophic ocean model, in which deformation of the thermocline is not allowed. The time integration for standard coupling quickly changes the system into a regime which can no longer be compared to the results of the previous subsection. Furthermore, because of the large flow anomalies in the last part of this integration, the SST-equation can only be solved on this grid with a higher (unrealistic) diffusivity coefficient, which blurs the atmospheric response to the ocean anomalies. In following years the changed regime and the unrealistic SST-patterns lead to a quasi-chaotic timeseries, on which an analysis would be irrelevant for the issues of interest here. Therefore, we do not show further results on the coupled time integration of the one-layer configuration.
3.5 Summary and discussion

In a coupled ocean-atmosphere model of intermediate complexity the variability of the mid-latitude climate system is explored. An eastward atmospheric jet in a two-layer quasi-geostrophic atmosphere model forces a double-gyre ocean circulation in a two- or one-layer quasi-geostrophic ocean model, through wind-stress forcing. Embedded in the upper ocean layer is a surface layer, in which SST couples to the atmosphere. Linear stability analysis shows different stability characteristics for the two- and one-layer ocean model configuration. In the two-layer ocean model internal ocean modes are the most unstable modes and coupling with the atmosphere has only small effects on the stability of these modes with intermonthly period. In the one-layer model asymmetric multiple steady state solutions are unstable to barotropic Rossby basin modes (period smaller than one month) and to the gyre mode, which has a period of about 8 months, but is known to exhibit a decadal character in multi-layer ocean models (Nauw and Dijkstra, 2001; Simonnet and Dijkstra, 2002). Time integration of the two-layer ocean model shows nonlinearly induced low-frequency variability resembling the results of chapter 2; coupling with the atmosphere damps out this low-frequency oscillatory signal. The nonlinear time evolution of the one-layer ocean model shows a minor role for rectification processes and thus the gyre mode is the only source of "low-frequency" variability. Coupling with the atmosphere not only changes the ocean variability, but also destroys the asymmetry of the time mean ocean circulation.

We will discuss these results by subsequently addressing the main questions stated in the introduction of the chapter. The first question (1) "Can we find a similar coupled low-frequency mode as in the jet-jet system?" must be answered with a simple "no". The most unstable modes found in the jet-gyre system are clearly internal ocean modes and show relatively small spatial anomalous patterns, limited to the region where ocean velocities are maximum. This smaller spatial scale of the internal ocean modes is necessarily accompanied by a smaller time scale, if compared to the large-scale coupled mode described in chapter 2. A temporal and spatial scale difference arises between the ocean and atmosphere and therefore oceanic and atmospheric waves do not "match" (Weng and Neelin, 1998; Neelin and Weng, 1999). However, a well-known characteristic of coarse gridded quasi-geostrophic models describing the double-gyre ocean circulation is that the eastward jet hardly penetrates into the ocean interior. The basic oceanic steady states established in our model indeed show the eastward jet to be limited to the western boundary region. It is likely that on a more realistic basic state with a further eastward penetrating jet, the most unstable modes have a large spatial and temporal scale like the coupled mode described in section 2.3, with a large effect of ocean-atmosphere coupling. Therefore, the results of this chapter do not falsify the presence of a large-scale coupled mode as found in chapter 2 in the real climate system.

The second question (2) "Can we find low-frequency variability in the jet-gyre system due to rectification of the time mean flow as was found in chapter 2?" is answered with a "yes", at least for the two-layer ocean model configuration. The nonlinear interaction between intermonthly ocean variability and the time mean state induces interannual to decadal variability. Similar to the jet-jet system, a low-frequency oscillatory weakening/strengthening of the oceanic and atmospheric jet is found. Unlike the jet-jet system, here the role of the atmosphere is only passive and even damps out the nonlinearly induced variability. However, further analysis is needed in more elaborate models, because an additional stochastic atmo-
spheric component might cause an amplification of the nonlinearly induced low-frequency variability, rather than damping it as shown here. Interesting work in this direction has been performed by Dewar (2001). In time integrations in a quasi-geostrophic ocean model underneath a highly idealized atmosphere, he finds low-frequency variability in the ocean and examines the effect of ocean-atmosphere coupling and atmospheric noise on this variability. Returning to question (2), the nonlinearly induced low-frequency oscillatory meridional shifting of the jet, as found in the jet-jet system is not found in the jet-gyre system. In the one-layer ocean configuration the role of rectification processes is of minor importance.

The third question (3) ”Are there additional mechanisms leading to low-frequency variability, not found in the jet-jet system?” is answered by the results of the one-layer model. These results show that the double-gyre circulation is unstable for the gyre mode, described in detail in a number of studies (Jiang et al., 1995; Speich et al., 1995; Dijkstra and Katsman, 1997; Chang et al., 2001; Simonnet and Dijkstra, 2002; Nauw and Dijkstra, 2001; Van der Vaart et al., 2002). These studies have shown that in multi-layer (more realistic) models this gyre mode has a decadal period. In this chapter it is shown that this gyre mode exists in a large ocean basin and some important characteristics have been put forward. However, the effect of the atmosphere on this gyre mode remains unclear. It must be concluded, that the one-layer ocean model is not very well suited to investigate the characteristics of the gyre mode coupled to the atmosphere.

Adding the results of this and the previous chapter, one can conclude that two mechanisms might generate low-frequency variability at midlatitudes: nonlinearly induced variability due to rectification processes and/or low-frequency variability arising through an internal ocean mode, the gyre mode.
Chapter 4

Decadal variability in a coupled ocean-atmosphere GCM

The output of a 300-year simulation with the Climate System Model (version 1.0) is studied with a focus on decadal variability in the North Atlantic. Using Multichannel Singular Spectrum Analysis, a robust decadal oscillatory signal is found with specific spatial patterns of oceanic and atmospheric fields. In the ocean surface and subsurface fields, anomalies propagate from the northwestern Labrador Sea south-eastward along the coast and bend around Newfoundland before they enter the subtropical gyre. The pressure patterns in the atmosphere show a strong resemblance with those of the North Atlantic Oscillation. A comparison of variability in both individual oceanic and atmospheric fields and in combined sets of fields provides strong support that the oceanic temperature anomalies influence both the amplitude and phase of the NAO. Current theories on the physics of such variability are invoked to interpret the patterns of variability, but are found to be insufficient in several aspects. A new framework, in which nonlinear ocean dynamical processes set the decadal time scale, is presented to explain the decadal variability.

4.1 Introduction

As became clear in chapter 1, the dominant pattern of atmospheric low-frequency variability in the Atlantic region is the NAO-pattern, a dipole pressure system with centers near Iceland and the Azores (Rogers, 1990; Cayan, 1992; Hurrell and Van Loon, 1997). Part of the Sea Surface Temperature (SST) variability in the North Atlantic Ocean appears highly correlated to the atmospheric fluctuations (Wallace et al., 1990; Deser and Blackmon, 1993; Deser and Timlin, 1997) and is explained by a dipole structure with centers east of the coast of Florida and near Newfoundland. This pattern may be linked to a lower lobe with a center in the low-latitude mid-Atlantic, together sometimes referred to as the North Atlantic tripole.

Another view of this variability is obtained by looking at propagating patterns. In Sutton and Allen (1997), it was shown that SST-anomalies near Florida propagate to the north-east in successive years, following the Gulf Stream and North Atlantic Current (NAC). These
anomalies turn out to be well correlated with the NAO. In Da Costa and Colin de Verdière (2002), a propagating oscillatory mode (with a period of 7.7 years) is found in 120 years of observed SST data. This mode is highly correlated with an associated propagating oscillatory mode in Sea Level Pressure (SLP), with a spatial pattern resembling that of the NAO. As in Sutton and Allen (1997), a large-scale negative SST anomaly starting at the Florida coast propagates to Newfoundland along the Gulf Stream and is associated with a negative NAO-phase (positive SLP-anomalies in the northern North Atlantic, negative SLP-anomalies in the central North Atlantic). When the SST-anomalies reach the Newfoundland region (and positive SST-anomalies start to grow near Florida and start migrating north-eastward) the SLP-dipole switches sign, resulting in the opposite phase of the NAO. Additional statistical analysis, specifically designed to detect propagating oscillating patterns, has indicated that a significant part of SST-variability in the North Atlantic is explained by such propagating patterns (Hansen and Bezdek, 1996; Moron et al., 1998; Tourre et al., 1999). Zorita and Frankignoul (1997) and Frankignoul et al. (2000) analyzed combined surface fields of ocean and atmosphere in the output of the multi-century ECHAM1/LSG coupled simulation, also using “propagation detecting analyses”. Both studies find mainly an oceanic response to atmospheric forcing and decadal coupled modes only weakly differ from red noise. Frankignoul et al. (2000) find a weak positive feedback between the atmosphere, SST and the strength of the double-gyre ocean circulation. Also our results of the analyses on the idealized models in chapters 2 and 3 have shown propagating patterns, explaining a significant part of the variability. Both the small-scale and large-scale internal ocean modes found in sections 2.3 and 3.3 propagate eastward (westward compared to the stationary flow). The gyre mode shown in section 3.4 travels either north- or southward, depending on the stationary state. However, the nonlinearly induced low-frequency variability found in sections 2.4 and 3.3 is best described by a standing oscillation, without significant propagation.

Many of the studies on coupled ocean-atmosphere low-frequency variability in the Atlantic region described in chapter 1, indicate that coupling between the ocean and atmosphere is most effective at the thermal front located at the midlatitude gyre boundary. During a positive phase of the NAO, when the westerlies accelerate, the gyre circulation in the ocean will respond with a delay which is set by the propagation speed of long Rossby waves. The anomalous circulation will induce SST anomalies near the thermal front. If the gyres amplify asymmetrically, then the mean ocean current shifts which results in anomalous advection of heat over the thermal front and decreases the meridional SST-gradient. Reflection of this pattern in the atmospheric surface temperature leads, through thermal wind, to a negative feedback (Marshall et al., 2001). If this front is sharpened because the strength of both gyres increases, then because of the enhanced meridional temperature gradient in the atmosphere, a positive feedback and decadal coupled modes may result (Cessi, 2000; Primeau and Cessi, 2001). Apart from baroclinic Rossby waves, the delayed response in the ocean may also be caused by adjustment of the thermohaline circulation (Marshall et al., 2001).

In this chapter we analyze a 300-year simulation of a state-of-the-art coupled ocean-atmosphere model, namely the NCAR Climate Systems Model (CSM). Again, the focus will be on low-frequency variability. Can we find significant modes of variability, how do they look like and can we relate them to the fundamental mechanisms described in the previous chapters. In order to answer these questions we apply Multichannel Singular Spectrum Analysis (MSSA), a technique designed to find propagating patterns that are optimal in the
representation of the variance. Our analysis shows that robust decadal variability exists in both oceanic and atmospheric fields on a time scale of 80-120 months and is associated with propagating patterns. A comparison of variability in both individual oceanic and atmospheric fields and in combinations of fields provides strong support that the oceanic temperature anomalies influence both the amplitude and phase of the NAO. The results are subsequently analyzed and current theories on the physics of this variability, as described above, are invoked to interpret the decadal variability. The data set and the statistical methods are shortly described in section 4.2, with details in appendix B. Section 4.3 contains the main results on MSSA modes in both individual ocean and atmosphere fields as well as in joint data sets. The physics of these modes is analyzed in section 4.4 and a summary and discussion of the results follows in section 4.5.

4.2 CSM Output and Analysis Methods

All data sets used here are output of a 300-year run of the NCAR Climate Systems Model (CSM), version 1.0. More details on the model, its performance and this particular simulation can be found in a special issue of the Journal of Climate (Volume 11, no. 6, (1998)).

The ocean model equations are solved (Gent et al., 1998) on a 152×111 longitude-latitude grid, with a uniform longitudinal resolution of 2.4° and a latitudinal resolution varying from 1.2° near the equator and high latitudes (> 60°) to 2.3° at 20°N and 20°S. In the vertical, there are 45 non-equidistant levels with a distance varying from about 12.5 to 250 meters. From this global grid, a 53×41 subgrid is extracted over the domain 19.8°N-80.4°N, 90.0°W-30°E, capturing the extra-tropical Atlantic. In the analysis of the oceanic fields, the first 27 years of the simulation were not used, leaving 273×12 = 3276 monthly mean fields of data. The annual cycle was removed by subtracting the monthly mean from the values of that particular month; this filtering does not remove all seasonal variability. Since this chapter deals with interannual to interdecadal variability and the model run contains some significant drifts in deep ocean properties (Boville and Gent, 1998) and has an unrealistically large spatial extent of sea-ice in the second centennial period, a cosine weighted high-pass filter is performed on the data, with a threshold of 720 months. Finally, for SST, ocean grid points where sea-ice occurred were excluded from the analysis.

The atmospheric equations are solved on a 128×64 longitude-latitude grid, with a T42 spherical harmonic truncation, having a nominal horizontal resolution of 1.8° and 18 levels in the vertical (Kiehl et al., 1998). Atmospheric fields, defined on a 44×22 subgrid over the domain 20.9°-79.5°N, 92.8°W-28.1°E were analyzed. Again the first 27 years were excluded and the same filtering (excluding ultra-low frequency variability and eliminating the annual cycle) as on the ocean data was performed. Furthermore, an additional cosine low-pass filter (threshold 10 months) has been applied to exclude intermonthly variability.

The timeseries of the oceanic and atmospheric data fields are analyzed using MSSA (Plaut and Vautard, 1994). MSSA is a statistical technique designed to detect propagating patterns in noisy, relatively short timeseries. In order to reduce the number of spatial dimensions, without the loss of essential information, initially an EOF-analysis is performed and the Principal Components (PC’s) are used as input for the MSSA-algorithm. The number (L) of PC’s that were used in the MSSA-analysis was chosen such that over 80% of the variance was captured
and depends on the variable which is analyzed. Further details on MSSA such as the choice of the window length \((M)\), the determination of the oscillatory patterns as reconstructed components (RC’s) and on the test for their statistical significance is provided in appendix B.

Apart from applying MSSA on the individual oceanic- or atmospheric data fields, also combined sets of fields were analyzed. Before different data fields were combined, the individual fields were normalized with respect to the whole (individual) data set. Although this normalization of the data could blur the relative importance of the individual fields in their mutual interaction, it is the most objective way to deal with it. The individual fields were put together as being one field and then EOF-analysis was performed on this extended data set; the PC’s again form the input channels for MSSA.

### 4.3 Results

The oceanic and atmospheric fields which were analyzed are shown in Table 4.1, together with the acronyms which are used in the chapter. In the first subsection below, we compare the time-mean state and first EOF in the output of the CSM with those in observations and other GCM simulations. In the next subsection, MSSA is performed to identify dominant propagating patterns of variability in individual fields as well as in combined fields.

<table>
<thead>
<tr>
<th>Field</th>
<th>Acronym</th>
</tr>
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<tbody>
<tr>
<td>Sea Surface Temperature</td>
<td>SST</td>
</tr>
<tr>
<td>Sea Surface Zonal Velocity</td>
<td>SSU</td>
</tr>
<tr>
<td>Sea Surface Meridional Velocity</td>
<td>SSV</td>
</tr>
<tr>
<td>Potential Temperature (at 262m depth)</td>
<td>PT262</td>
</tr>
<tr>
<td>Zonal Velocity (at 262m depth)</td>
<td>UO262</td>
</tr>
<tr>
<td>Meridional Velocity (at 262m depth)</td>
<td>VO262</td>
</tr>
<tr>
<td>Potential Temperature (at 711m depth)</td>
<td>PT711</td>
</tr>
<tr>
<td>Zonal Velocity (at 711m depth)</td>
<td>UO711</td>
</tr>
<tr>
<td>Meridional Velocity (at 711m depth)</td>
<td>VO711</td>
</tr>
<tr>
<td>Barotropic Streamfunction</td>
<td>PSI</td>
</tr>
<tr>
<td>Sea Level Pressure</td>
<td>SLP</td>
</tr>
<tr>
<td>Zonal Velocity at 850hPa</td>
<td>UA850</td>
</tr>
</tbody>
</table>

Table 4.1: Ocean and atmosphere fields analyzed in the CSM output and the acronyms used.
4.3.1 Time-mean state and first EOF

Since the time-mean state plays an important role in the analysis of the variability in subsequent sections, relevant mean patterns are shown in Fig. 4.1. Comparing the oceanic surface velocity and temperature (Figs. 4.1a and b, respectively) with observations (Levitus, 1982; Meehl, 1982) it is clear that the meridional temperature gradient as well as the gyre velocities are underestimated, both due to the coarse horizontal resolution in the CSM. The model mean fields in the atmosphere show the well-known dipole structure in SLP (Fig. 4.1c) and the associated atmospheric jet at 850hPa (Fig. 4.1d).

Because in many previous studies of North Atlantic variability, prime focus has been on EOF's, we show the patterns of the leading EOF in the CSM simulation. The first EOF of SSU shows a large-scale dipole pattern (Fig. 4.2a), with centers west of Spain and south of Greenland. This dipole pattern represents a meridional shift of the mean eastward ocean current. The first EOF of SST (Fig. 4.2b) resembles the well-known dipole pattern (Deser and Blackmon, 1993; Wallace et al., 1990). Compared to observations, the amplitude of the southern center of the dipole is weaker relative to the northern center. In the first EOF of SLP (Fig. 4.2c), a typical NAO-pattern is found, with a corresponding zonal velocity field (Fig. 4.2d). The northern center of the dipole structure in SLP is located too far to the northeast, when compared to observations (e.g. Fig. 5 in Cayan (1992)). Both results on mean state and the first EOF for both oceanic and atmospheric fields give confidence in the quality of this CSM simulation, as far as North Atlantic variability is concerned.
Figure 4.2: EOF 1 of a) Sea surface zonal velocity (explaining 22% of the variance), b) Sea surface temperature (16%), c) Sea level pressure (43%) and d) Zonal velocity at 850hPa (36%).

4.3.2 Propagating patterns of variability

From the MSSA-analysis, a decadal mode was found in all quantities, except SST and SLP, either as MSSA-mode 1+2 (for PSI and UA850) or as MSSA-mode 3+4 (SSU, SSV, PT262, UO262, VO262, PT711, UO711 and VO711). A windowlength of 150 months was used in the analyses, but the decadal mode was robust to variation of the windowlength over a range $M = 100$ to $M = 300$ months. The spatial/temporal structure of the modes is found from the Reconstructed Components (RC’s) and the Space-Time Principal Components (ST-PC’s). The latter is the timeseries of an individual mode, of which a pair forms the RC. Patterns of the RC’s will be shown below at different phases of the oscillation, giving the characteristics of the oscillating part of the total signal. Animations of the different modes can be found at http://www.phys.uu.nl/~wwwimau/avoird/article/RCs.html.

Ocean

In all individual ocean fields, except SST, an MSSA-mode with a period of 116 months is found. Results from the significance test (see appendix B) for PSI are shown in Fig. 4.3. All open circles above the drawn curve indicate modes, for which it is unlikely that these are caused by red-noise processes. The 116-month mode is marked by the arrow in Fig. 4.3 and it is significant for all other ocean quantities (except SST). The reason that the 116-month mode does not show up in SST is probably because the second century of the integration showed a rather odd sea-ice distribution (Weatherly et al., 1998), which strongly affects SST-variability.
Figure 4.3: Monte Carlo significance test of the Barotropic Streamfunction (PSI), using 20 PC’s from the conventional EOF analysis. Shown are projections of PSI-data onto the AR(1) null-hypothesis basis. Open circles are the data eigenvalues, plotted against the dominant period of the corresponding ST-PC. The drawn line shows the 99% confidence level computed from 1000 realizations of a noise model consisting of 20 independent AR(1) processes with the same variance and lag-1 autocorrelation as the input data channels. The arrow points at the 116-month mode described in the text.

The RC of the MSSA mode 3+4 of the combined fields SSU and SSV (also having a period of 116 months) is plotted as a vector-plot in Fig. 4.4, showing half a period of the oscillation. Following the subsequent phases in Figs. 4.4a-h, a clear propagating signal is apparent. At phase $t = 0$ (January of year 67 of the integration), a large anomalous anti-cyclonic circulation east of Newfoundland (Fig. 4.4a), pushes the eastward ocean jet (Gulf Stream/NAC) in a more northerly position. About 6 months later, the anti-cyclonic circulation has moved slightly to the southwest (Fig. 4.4b), a direction of propagation which is continued in the following years (and throughout the entire integration). Anomalies are clearly centered along the American east-coast and have a small amplitude in the rest of the basin. In the phases pictured in Figs. 4.4b and 4.4c, there is a signature of an anti-cyclonic ’intergyre gyre’ (Marshall et al., 2001). In Fig. 4.4d, the anomalous anti-cyclonic circulation strengthens the subtropical gyre. With time, as the anti-cyclonic circulation keeps on propagating towards Florida and slowly decays a new, now cyclonic, circulation develops east of Newfoundland (Figs. 4.4e-f). In Figs. 4.4g-h this cyclonic circulation now pushes the ocean jet into a more southerly position. This picture of south-westward propagation of subsequent anomalies of positive and negative relative vorticity is characteristic of the decadal variability in the model simulation. Although in Figs. 4.4e-f there is some anomalous circulation in the Labrador Sea, this region does not significantly contribute to the variance accounted for by this mode.
Figure 4.4: MSSA mode 3+4 of combined SSU and SSV (cm s\(^{-1}\)): a) - h) are phase plots characteristic for about half a period of oscillation (actual months: 469, 475, 481, 487, 493, 499, 505, 511).
Figure 4.5: MSSA mode 3+4 of combined fields at 711 meters depth: PT711 (°C). a) - h) as in Fig. 4.4.
Figure 4.6: a) MSSA mode 3+4 of combined fields at 711 meters depth. a) The path 1-25 along which the Hörmüller diagram of PT711 (°C) drawn in b) is calculated. The time interval 210-260 years is characteristic for the whole integration.

In the combined fields of horizontal velocities (UO711 and VO711) and potential temperature (PT711) at 711 m depth, also a dominant mode (MSSA mode 3+4) with a period of 116 months is found. The pattern of PT711 of the RC of this mode indicates (Fig. 4.5) that anomalies within the Gulf Stream region follow the same upstream propagation as the 116-month mode in surface velocities. Contrary to the surface velocity signal, there is also substantial amplitude in the subsurface Labrador Sea. In Fig. 4.5a, there are negative PT711 anomalies in the Labrador Sea, accompanied by positive anomalies southeast of New Foundland. In the following years, the negative anomalies in the Labrador Sea travel south-eastward into the Atlantic, bend around Newfoundland and then propagate in a south-westward direction (Fig. 4.5d-h). Meanwhile, a positive PT711 anomaly appears in the Labrador Sea (Fig. 4.5d-f) which also moves towards the Atlantic region. This replaces the negative anomaly, giving the anomalous pattern of opposite sign after about half a period of oscillation (in Fig. 4.5h). A more detailed view on the propagation of anomalies of PT711, can be obtained from a Hörmüller diagram along the path plotted in Fig. 4.6a. The anomalies propagate along this path with a phase velocity of about 2 cms\(^{-1}\) in the Labrador Sea, slowing down to about 1 cms\(^{-1}\) east of New Foundland and accelerating to about 5 cms\(^{-1}\) in the subtropical gyre (Fig. 4.6b).

When MSSA is performed on the combined fields SSU, SSV and SST, the 116-month mode turns out to be significant, with the RC of SST (not shown) resembling that of PT711, except in the Labrador Sea (due to the presence of sea-ice). The patterns of the ocean surface velocity and temperature anomalies resemble those found by Watanabe et al. (1999) in GCM-output, although the dominant time-scale they find (9.9 years) is slightly different. In
the CSM, the amplitude of the temperature anomalies in the Labrador Sea is also significant and comparable to that in the central North-Atlantic. Furthermore, the subsurface temperature anomalies indicate that the phenomenon extends towards the deeper ocean. In fact, the 116-month mode has a significant barotropic signature (Fig. 4.7), involving the entire gyre circulation. For the barotropic streamfunction (PSI), the mode is not only significant (Fig. 4.3), but it also explains 7.6% of the variance over the whole basin and southeast of Newfoundland, the correlation coefficient of the RC and the "raw data" (annual cycle and longterm trend removed) exceeds 0.67. Strong anomalies are apparent in the Labrador Sea (Fig. 4.7c) and they travel south, around Newfoundland, into the subtropical gyre, influencing the ocean gyre circulation. Characteristic for the PSI-signature of the mode is that, similar to Figs. 4.4 and 4.5, the amplitude is largest in the Newfoundland region. At t=0 (same phase as in Figs. 4.4-4.5), the anomalous streamfunction weakens the mean circulation, the subtropical as well as the subpolar gyre (Fig. 4.7a). At later times, the positive anomalous barotropic streamfunction induces an anti-cyclonic geostrophic circulation in the region where the subtropical and subpolar gyre interfere (Fig. 4.7b-d). In this situation, there appears a barotropic anti-cyclonic ‘intergyre’ gyre (Marshall et al., 2001) and the mean Gulf Stream axis is clearly shifted northwards. As the anomaly moves south-westwards towards Florida (Figs. 4.7f-h), it strengthens the mean circulation.

**Atmosphere**

For the atmospheric fields considered, no atmospheric modes were found to be significant under a red-noise (which is essentially white noise if the coefficient of the AR(1) process is small, which holds for the atmospheric GCM output) null-hypothesis. However, in the results of MSSA on UA850, an 86-month mode was found, RC 17+18, which after applying the low-pass filter as described in section 4.2 became RC 1+2, explaining 7.4% of the total variance. The characteristics of this 86-month mode in UA850 are shown in Fig. 4.8. At t=0 (again the initial phase coincides with that in Figs. 4.4, 4.5 and 4.7), a strong dipole structure is present. If UA850 is assumed to be representative for the atmospheric jetstream, this pattern corresponds to a northward shift of the mean atmospheric jet axis (cf. Fig. 4.1d). The anomalous UA850-pattern in Fig. 4.8a is related to the positive phase of the NAO pattern (cf. Fig. 4.2c). As time progresses, the positive northern center seems to propagate southward and slowly decays when reaching the central mean jet-axis (Figs. 4.8b-d). A negative zonal velocity anomaly appears north of Iceland and replaces the initial positive center (Figs. 4.8d-h). Meanwhile, the initial positive anomaly propagates southward and regains strength as it moves away from the mean jet-axis.
Figure 4.7: MSSA mode 1+2 of PSI (Sv). a) - h) as in Fig. 4.4.
Figure 4.8: MSSA mode 1+2 of UA850 (m s$^{-1}$). a) - h) as in Fig. 4.4.
Combined oceanic and atmospheric fields

To determine patterns of co-variability in oceanic and atmospheric fields, MSSA was performed on the combined fields of PSI, PT262, UA850 and SLP. The variables are chosen such that they are assumed to represent the major fields involved in ocean-atmosphere interaction. PT262 is chosen because of a possible response of the atmosphere on oceanic surface anomalies. SST would be a better quantity for this reason but, as mentioned before, the SST output of CSM was ‘polluted’ by sea-ice anomalies. The 116-month mode in PT262 of the combined analysis of PT262, U262 and V262 is highly correlated to the 116-month mode in SST of the combined analysis of SSU, SSV and SST. Hence, PT262 is representative for SST in the results below.

Mode 3+4 of MSSA on the four combined fields revealed a significant 85-month mode and four different phases of the RC’s are shown in Figs. 4.9, 4.10, 4.11 and 4.12. Note that for PSI and UA850, the plots are at actual months which are a subset of those plotted for the individual fields. The patterns in Figs. 4.9 and 4.11 show a very good resemblance with the RC’s of the 116-month individual PSI mode (Fig. 4.7) and the 86-month individual UA850 mode (Fig. 4.8), respectively. Consistent with the patterns of the individual 116-month mode for PT711, also in the patterns of PT262 of the combined mode (Fig. 4.10) large-scale anomalies propagate from the Labrador Sea south-eastward, bend around Newfoundland and move

Figure 4.9: MSSA mode 3+4 of the combined fields of PSI, PT262, UA850 and SLP: PSI (Sv). a) - d) are phase plots characteristic for about half a period of oscillation (actual months: 469, 481, 493, 505).
south-westward towards Florida. In the pattern for SLP, a positive NAO-index situation (amplified meridional pressure gradient) is recognized in Fig. 4.12a, with anomalously high zonal velocities north of the jet-axis (Fig. 4.11a). This dipole structure moves south, weakens (Figs. 4.12b-c) and then strengthens again giving the reversed NAO signature (Fig. 4.12d). Plots of the variance field of UA850 and SLP of this combined mode (Fig. 4.13) clearly show that the dominant atmospheric variability is associated with the NAO pattern (Figs. 4.2c-d).

The results indicate that, although in oceanic quantities the dominant time scale is 116-month, the combined mode has a dominant time scale of 85 months, more like the time scale found in the non-significant UA850 mode. To clarify this difference, the timeseries of the total kinetic energy ($\epsilon$) is plotted in Fig. 4.14a for both the 116-month mode in the individual PSI field as well as for the 85-month mode from the combined fields. Here, $\epsilon(t)$ is calculated as

$$\epsilon(t) = \sum_{i,j} \frac{1}{2} (\bar{u}_{i,j}^2(t) + \bar{v}_{i,j}^2(t))$$

(4.1a)

$$\bar{u}_{i,j} = -\frac{\partial PSI_{i,j}}{\partial y} ; \quad \bar{v}_{i,j} = \frac{\partial PSI_{i,j}}{\partial x} ;$$

(4.1b)

where $\bar{u}$ and $\bar{v}$ are the depth integrated zonal and meridional velocity and the summation is over the entire grid. Here, PSI is the sum of the mean value of PSI (averaged over the whole integration) and the anomalous PSI due to the MSSA-mode. The drawn (dashed) curve in
Fig. 4.14a shows $\epsilon(t)$ due to the individual 116-month MSSA mode (85-month combined MSSA mode). Although the dominant time scale is different, the timeseries in Fig. 4.14a correlate very well; the correlation coefficient is 0.69. In periods where the amplitude of both modes is relatively large, roughly from year 28-90 and year 200-300, this correlation is much higher. Both modes are damped in the middle part of the integration, which is the interval for which the spatial extension of sea-ice is overestimated (Weatherly et al., 1998). Another interesting feature in Fig. 4.14a is that, tracing the months in which $\epsilon$ is at its maximum, the period of the oscillation varies in the range from about 80-120 months. Hence, although the dominant time scale is different, individual 116-month MSSA mode and the 85-month combined MSSA mode are very similar and hence they represent the same preferred spatial/temporal pattern.

The atmospheric temporal behavior is shown as anomalies of UA850 (Fig. 4.14b) of either the individual 86-month MSSA mode 1+2 or the combined 85-month MSSA mode 3+4. The particular gridpoint (60°N,34°W) is taken at the point of highest variance. In Fig.4.14b, the signal is largest and most regular in the first and last part of the integration and the period ranges from 80 to 120 months. The correlation coefficient between the two timeseries shown in Fig.4.14b is 0.91, again confirming that the atmospheric modes displayed above relate to the same spatial/temporal pattern. Because the dashed curves in Fig.4.14a and b are highly correlated, the oceanic and atmospheric mode are essentially one coupled mode. Although the spectrum of the atmosphere itself can not be distinguished from a white noise one, there is actually a decadal mode hidden in this noise. This decadal mode is related to oceanic variability which does significantly exceed the red-noise power spectrum.
Figure 4.12: MSSA mode 3+4 of the combined fields of PSI, PT262, UA850 and SLP: SLP (Pa). (a) - (d) as in Fig. 4.9.

Figure 4.13: Square root of the variance of RC 3+4, determined from MSSA on the combined fields of PSI, PT262, SLP and UA850: (a) UA850 (ms⁻¹), (b) SLP (Pa).
Figure 4.14: Timeseries of a) Drawn: Kinetic Energy ($m^2 s^{-2}$) of the sum of RC 1+2 and the mean field of PSI. RC 1+2 determined from MSSA on PSI. Dashed: Kinetic Energy ($m^2 s^{-2}$) of the sum of RC 3+4 and the mean field of PSI. RC 3+4 determined from MSSA on the combined fields of PSI, PT262, SLP and UA850. b) Drawn: UA850 ($ms^{-1}$) at (60°N, 34°W) of RC 1+2, determined from MSSA on the UA850. Dashed: UA850 ($ms^{-1}$) at (60°N, 34°W) of RC 3+4, determined from MSSA on the combined fields of PSI, PT262, SLP and UA850.
4.4 Evaluation of current theories

Having identified the properties of this decadal mode, we now turn to the physical processes involved. The results from recent studies on this variability (Cessi, 2000; Primeau and Cessi, 2001; Marshall et al., 2001) motivate to look at:

1) How do the wind anomalies in the extreme NAO phases influence the gyre circulation?
2) How does the gyre response influence the meridional temperature gradient near the northern American east-coast?
3) Can the ocean adjustment processes be identified which determine the decadal time scale?

We will address each of these issues in turn below.

For the initial positive phase of the NAO situation in Fig. 4.12a, the atmospheric jetstream is shifted northwards (Fig. 4.11a), in agreement with the view in Marshall et al. (2001). The effect of this anomalous atmospheric circulation on the ocean is investigated by plotting the strength of the atmospheric jet north of the mean jet-axis (drawn curve in Fig. 4.15), and the total kinetic energy $\epsilon$ (dotted curve in Fig. 4.15) of the 85-month mode. For clarity, we have only plotted the first part of the integration, being characteristic for the whole period. The kinetic energy of the ocean circulation lags the atmospheric jetstream, and a lag-correlation (over the entire 273 years) analysis shows a maximum correlation coefficient of 0.93 if $\epsilon$ lags UA850 by about 30 months.

The high-energy flow pattern which follows a positive phase of the NAO is determined by making a composite pattern (from the raw data) of the months where $\epsilon$ is maximal, resulting in Fig. 4.16. A state of high energy corresponds to a situation where both subpolar and subtropical gyre are amplified, but also where the subtropical gyre is extended northwards.

![Figure 4.15: Drawn: Timeseries of Kinetic Energy (m$^2$s$^{-2}$) of the sum of RC 3+4 and the mean field of PSI. Dashed: UA850 (m$^2$s$^{-1}$) at (60°N,34°W) of RC 3+4. RC 3+4 for both lines determined from MSSA on the combined fields of PSI, PT262, SLP and UA850.](image)
Hence a positive phase of the NAO is followed (about 30 months later) by an anomalous barotropic ocean gyre circulation which pushes the Gulf Stream into a more northerly position. An observational study by Taylor and Stephens (1998) also shows a northward shift of the northern wall of the Gulf Stream, lagging the NAO by 2 years. The amplification of both oceanic gyres after a positive NAO phase is in agreement with the view of Cessi (2000). However, as the amplification is asymmetrical, it also agrees with the anti-cyclonic ‘intergyre gyre’ view of Marshall et al. (2001).

The crucial point is now to determine the effect of the changes of the gyre circulation on the temperature field near the region where the mean meridional temperature gradient is large (cf. Fig. 1b). In Marshall et al. (2001), the anomalous intergyre gyre circulation advects warm water northward thereby decreasing the meridional temperature gradient, which in turn weakens the atmospheric jetstream. On this weakened jetstream fewer depressions arise and the NAO shifts to a negative phase, indicating a negative feedback mechanism. On the other hand, Cessi (2000) find that amplification of both gyres increases the temperature gradient at the thermal front, thereby amplifying the atmospheric jetstream, resulting in a positive feedback. In the results of the CSM, these two opposite effects on the temperature gradient near the thermal front are both present. A third effect is the direct thermal response on the atmospheric forcing (Deser and Timlin, 1997), increasing the meridional temperature gradient. All these effects combined give an increased meridional temperature gradient east of Newfoundland (Fig. 4.10c/d). In Fig. 4.17, the timeseries of the anomalous PT262 gradient east of Newfoundland and the anomalous UA850 at the center of the atmospheric jet has been plotted (note that this timeseries of UA850 is at another gridpoint as that of Figs. 4.14 and 4.15). The correlation coefficient between both series is 0.76 and indicates that the atmospheric jet responds instantaneously to the local changes in the temperature gradient. These results combined seem to favor the view that a positive ocean-atmosphere feedback is involved in the variability (Latif and Barnett, 1994; Cessi, 2000; Watanabe and Kimoto, 2000). However, it should be noted that some important features blur the clear signal of a positive
feedback. Because of the time it takes for the ocean circulation to adjust to an anomalous atmospheric forcing, its effect on SST can not simply be added to the response due to the noise.

Next issue is the processes which determine the time scale of the variability. In current theories a baroclinic ocean wave, involved in ocean adjustment, is suggested to control the time scale. No signatures of such a baroclinic wave are seen in the results. The crossing time scale $\tau_c$ of such a wave over a length $L$ in a background field with Rossby deformation radius $L_D$ on a midlatitude $\beta$-plane is

$$\tau_c \approx \frac{L}{\beta L_D^2}$$

In the CSM, with about 1-2° horizontal resolution, the deformation of isopycnals is only captured on a scale larger than 100 km. This leads at 45°N to an estimate of $\tau_c$ of the order of about a year, using for $L$ the horizontal extent of the patterns of the temperature anomalies. Hence, although baroclinic adjustment may control the 30 months lag between wind and gyre response, it is not responsible for the decadal time scale in the CSM.

The latter is also consistent with the strong barotropic component of the decadal mode, which appears to control the heat transport over the thermal front. While slow barotropic adjustment through Rossby waves is unlikely, because of their fast propagation, in chapter 3 it has become clear that slow barotropic modes exist in the gyre circulation. The gyre mode found in section 3.4 showed barotropically controlled anomalies to propagate either north- or southward. This mode was shown to exhibit a decadal time scale in a number of multi-level models (Jiang et al., 1995; Speich et al., 1995; Chang et al., 2001; Simonnet and Dijkstra,
Decadal variability in a coupled ocean-atmosphere GCM (2002; Nauw and Dijkstra, 2001; Van der Vaart et al., 2002). The physics of this gyre mode has recently become transparent: it propagates due to a symmetry breaking instability and grows on the horizontal shear of the background flow (Simonnet and Dijkstra, 2002). In idealized basins, its signature upon the flow is through the kinetic energy, which oscillates between high- and low energy states (McCalpin and Haidvogel, 1996; Nauw and Dijkstra, 2001). Depending on the steady state on which the instability grows, the direction of propagation of anomalies is either northward or southward, from one recirculation gyre (amplified/weakened double-gyre circulation) via the eastward jet (cyclonic/anti-cyclonic intergyre gyre) in between towards the other recirculation gyre (weakened/amplified double-gyre circulation). Although obviously many differences between the CSM and our model used in chapter 3 cause many differences in the appearance of the gyre mode, there are clearly similarities in the spatial characteristics of Figs 3.14 and 4.9.

Support for the presence of the gyre mode in the CSM is (i) the barotropic characteristics of the coupled mode, (ii) the appearance of high/low kinetic energies in the mode, (iii) the localization of the pattern of the mode around the Gulf Stream axis and (iv) the north-south propagation. Hence, such an internal mode may be responsible for the decadal time scale of variability in the CSM. The ocean temperature anomalies caused by this internal ocean mode then become involved in a coupling with the atmosphere, as described above.

4.5 Summary and discussion

Multichannel Singular Spectrum Analysis performed on a 300-year simulation of the Climate Systems Model of NCAR has revealed a 80-120 month coupled ocean-atmosphere mode. The mode is statistically significant (with respect to a red-noise null-hypothesis) in the oceanic quantities analyzed. The anomalies in the ocean circulation and heat budget support a positive feedback with the atmosphere. Although the feedback causes the mode to be present in the atmosphere as well, it is not strong enough to raise the atmospheric decadal variability above the white noise. Main characteristics of this decadal coupled mode are the barotropic nature and southward propagation of the oceanic anomalies. Therefore, we suggest that an internal oceanic mode, the gyre mode, may be responsible for the decadal oscillation. Within this framework, the physical mechanism of the decadal mode in the coupled ocean-atmosphere system can be described as follows. Due to the gyre mode, large-scale anomalies in all ocean quantities propagate through the Labrador Sea, travel south, rounding Newfoundland and then propagate towards Florida. When a positive anomaly of the barotropic streamfunction is situated in the region where the subtropical and subpolar gyre interfere, both gyres are amplified and the Gulf Stream is pushed northwards. This results in warm temperature anomalies southeast of Newfoundland and cold temperature anomalies northeast of Newfoundland and thus an amplified meridional temperature gradient. This anomalous temperature field increases the atmospheric jet through thermal wind. Such an amplified jet favors the development of depressions on the northern jet exit, deepening the Icelandic Low; on the southern exit of the jet, high pressure systems are favored. This results in a positive phase of the NAO, which in turn forces a stronger and more northward ocean gyre circulation, thus amplifying the initial anomalous ocean circulation. Such a positive feedback fits in the views of Latif and Barnett (1994), Cessi (2000) and Watanabe and Kimoto (2000). The adjustment of the ocean
circulation lags the atmospheric forcing by about 2.5 years. The ocean anomalies propagate southward due to the gyre mode, and the atmospheric anomalies seem to follow this southward migration (Fig. 4.18). It can be seen that indeed the amplitude is largest on the northern and southern side of the mean atmospheric jet-axis. Because of this alternating amplitude of the atmospheric anomalies, the positive ocean-atmosphere feedback described above also prefers certain phases of the total decadal oscillation.

Current theories (Watanabe et al., 1999; Cessi, 2000; Primeau and Cessi, 2001; Marshall et al., 2001; Da Costa and Colin de Verdière, 2002) show that the decadal time scale in coupled ocean-atmosphere oscillations is set by the baroclinic ocean adjustment to changes in the atmospheric forcing. The characteristics of the 80-120 month mode do not support this mechanism to be responsible for the decadal variability in the CSM. The oceanic anomalies of the mode clearly exhibit a barotropic nature and in addition, they propagate southward instead of westward.

Because of the clear temperature signal in the Labrador Sea, one might be tempted to relate the decadal mode to thermohaline and/or sea-ice driven variability. Also here, we think that this possibility is not very realistic because of the barotropic nature of the mode. Furthermore, although the mode is damped in the middle period of the CSM-integration (with large spatial extend of sea-ice, Weatherly et al. (1998)) it remains present in both atmosphere and ocean, even though the ocean-atmosphere interaction is negligible in the Northern Atlantic sinking regions east and west of Greenland.

Whereas Frankignoul et al. (2000), using both lagged SVD and MSSA, only find a weak decadal coupled signal in the extended ECHAM1/LSG coupled simulation, we find a much

Figure 4.18: MSSA mode 3+4 of the combined fields of PSI, PT262, UA850 and SLP: Hovmöller diagram of UA850 (m s⁻¹) at 34°W.
stronger signal. First, this could be due to a different resolution, because Frankignoul et al. (2000) do not resolve a proper Labrador Sea ocean circulation, which seems to be rather important in our study. Second, Frankignoul et al. (2000) only analyze surface quantities, whereas the decadal mode found here has a large amplitude in the ocean interior as well as the atmospheric jet. Frankignoul et al. (2000) also find indications of a weak positive ocean-atmosphere feedback.

Comparing the results with observational studies shows some interesting similarities. First of all, the dominant pattern of atmospheric variability of the decadal mode resembles the observed NAO-pattern (Cayan, 1992). Furthermore, this NAO variability changes the path of the Gulf Stream with a similar lag as in observations (Taylor and Stephens, 1998). The fact that processes in the Newfoundland region control the phase of the NAO through SST anomalies has been observed in Da Costa and Colin de Verdière (2002). In a statistical analysis rather similar to MSSA they find atmospheric patterns and a time scale very much resembling the results presented here. In the ocean, however, Da Costa and Colin de Verdière (2002) find anomalies propagating northeast in contradiction to our results. It is not clear how the direction of propagation of anomalies would change if the model Gulf Stream/NAC would be stronger, as in reality. In the idealized model described in chapter 3, the direction of propagation of the gyre mode is either southward or northward, depending on the steady state flow. The study on the relation between the gyre mode found in the idealized model and the coupled decadal mode found in this chapter must be examined further in future studies.

Finally, we want to comment on a question stated in the introduction of this chapter, whether we can relate the variability in the CSM to the fundamental mechanisms of low-frequency variability in the previous chapters. It is clear that one of these mechanisms, the gyre mode, is related to the variability found here, but we have not detected nonlinearly induced variability through rectification processes. The MSSA is not a good tool to find standing oscillations, but before applying MSSA an EOF-analysis was performed. EOF’s resembling those described in sections 2.4 and 3.3 have not been found as dominant EOF’s in the analysis of the CSM output. An obvious reason for this is that the resolution of the CSM is too coarse to resolve small-scale modes, which were found to be crucial for rectification processes. A more detailed discussion on the comparison between the results of the different chapters follows in chapter 5.

Summarizing the results of this chapter, the analysis here indicates (1) a decadal mode exists in the ocean and atmosphere in the CSM, that (2) a (localized) positive ocean-atmosphere feedback is involved in certain phases of its propagation and that (3) the time scale of the mode is solely caused by internal ocean dynamics through the gyre mode.
Chapter 5

Discussion

In this chapter, first the link between the different chapters will be discussed. Chapters 2-4 form the systematic approach as described in section 1.3. Can we indeed use fundamental physical mechanisms of low-frequency variability deduced from intermediate coupled models to interpret variability observed in more the complex climate system? Then, the basic hypotheses from section 1.2 are discussed, using the knowledge obtained in this thesis. Do our results favor the idea of internal atmospheric, internal ocean, or coupled variability leading to the observed low-frequency variability? Subsequently, we will put the results of this thesis in context with the literature on the subject to form a personal view on the physics of low-frequency variability at midlatitudes. Finally, a comment is made on how the results of this thesis relate to the issue of predictability of the NAO.

The approach of this thesis

The main goal of this thesis was to explore the physical mechanisms responsible for low-frequency variability at midlatitudes in a systematic way. Highly idealized models were used to identify basic mechanisms which may cause low-frequency variability. The starting situation was an atmospheric eastward jet overlying an eastward ocean jet. This jet-jet system appeared to be unstable to a large-scale coupled low-frequency mode, with a large-scale baroclinic ocean wave coupling with a barotropic atmospheric wave. The mode resembled a mode found by Goodman and Marshall (1999) and recently, Ferreira et al. (2001) also showed a rather similar mode to appear in a more complex model. With its long decadal time scale, this mode could be a possible prototype to explain part of the midlatitude coupled low-frequency variability. However, our systematic approach towards more complex models revealed that this coupled mode is not very robust. In transient flows in the jet-jet system the period of the mode decreased dramatically to a few years and its amplitude became very irregular, both features due to rectification processes. When zonal boundaries of the ocean basin were present, the large-scale coupled mode disappeared. In this jet-gyre system, the ocean variability was limited to a small region where velocities were largest and the small spatial scale of the modes of variability was accompanied by a small time scale. In this
way, a spatial and temporal scale difference between ocean and atmosphere decrease the possibility of significant interaction, which is necessary to destabilize the large-scale coupled mode. However, the eastward jet in the jet-gyre system did not penetrate very well into the ocean interior, compared to reality. Therefore, it is not yet possible to exclude the possibility of a large-scale coupled mode in the real jet-gyre system. However, in the output of the coupled GCM in chapter 4, no sign of the large-scale coupled mode was found. MSSA could have sorted out such an eastward propagating signal, but no significant statistical modes were found resembling the characteristics found in chapter 2. Although the large-scale mode might be present within the noise, it is not likely to explain a large part of the NAO-like variability.

From the transient flows in the jet-jet system it became clear that in this system a large part of the low-frequency variability was explained by rectification processes. Nonlinearly induced variations showed a weakening/strengthening and a meridional shifting of both ocean and atmosphere jet. The large-scale coupled mode on one hand and secondary small-scale internal ocean modes on the other hand seemed to have an opposite effect on the time-mean circulation and because of a varying amplitude of both processes the time-mean flow slowly changed, resulting in generic low-frequency variability. This mechanism appeared to be robust to the change from jet-jet to jet-gyre system. In the latter configuration it was shown that the presence of the large-scale coupled mode is not crucial for the nonlinearly induced variations. The weakening/strengthening of the eastward jets showed up to be an internal ocean process, because it also existed in the uncoupled system without atmospheric variability. Ocean-atmosphere coupling even seemed to damp out the nonlinearly induced variability, but it must be noted that introducing stochastic atmospheric noise might significantly change this picture. Because of the striking resemblance of the atmospheric spatial (a dipole north-south pressure pattern) and temporal (irregular decadal to interdecadal oscillations) behavior in the jet-jet, the jet-gyre and the real climate system, this mechanism is a serious candidate to explain a significant part of the NAO-like variability. The low-frequency meridional shifting of the eastward jet was not found in the jet-gyre system. This feature might have been a result of the large-scale mode, which was not present in the jet-gyre system. Unfortunately, a proper identification of this mechanism of nonlinearly induced low-frequency variability was not possible in the complex CSM. The fact that in the analysis of the CSM such a feature was not found, is not conclusive, because the grid of the CSM is too coarse to resolve small-scale internal ocean modes, which in chapters 2 and 3 have been shown to be crucial for the mechanism. Thus, for this mechanism, the systematic approach in this thesis is not completed yet: a next step has to be taken, namely increasing the resolution in more elaborate models.

Another fundamental mechanism, which was found to explain part of the low-frequency variability in the jet-gyre model of intermediate complexity, namely the gyre mode, was also found in the CSM. In the jet-gyre system of chapter 3, a barotropic internal ocean mode appeared in both linear stability analysis and (uncoupled) time integration. Whereas the time scale was rather small (about 8 months), its origin has been proven to be low-frequency (Simonnet and Dijkstra, 2002) and many studies have shown that this mode exhibits a decadal time scale in multi-layer models, in which its signature was equivalent barotropic, rather than purely barotropic. Apart from the time scale, the barotropic structure, its direction of propagation and the localization around the eastward jet resemble the characteristics of the gyre mode described in the literature (Jiang et al., 1995; Speich et al., 1995; Dijkstra and Katsman, 1997; Chang et al., 2001; Simonnet and Dijkstra, 2002; Nauw and Dijkstra, 2001).
Exactly these characteristics were found in a decadal mode extracted with MSSA in the 300-year simulation in the CSM. Thus, this internal ocean mode, shown to be a fundamental, robust feature in idealized models, also explains a significant part of the variability in a model of high complexity. Concerning this mechanism which generates low-frequency variability at midlatitudes, the systematic approach has shown promising results. Nevertheless, many aspects of the connection between the gyre mode as found in the idealized models and the identification in reality must be submitted to an additional investigation.

**Mechanisms of low-frequency NAO variability**

Having discussed the systematic approach used in this thesis, we now address the following question: How do the results found in chapters 2-4 relate to the basic hypotheses stated in chapter 1? The first hypothesis assumes that NAO-like low-frequency variability is generated within the atmosphere and that the oceanic NAO variability is simply a direct response. The results of chapters 2 and 3 do not contribute any insight concerning this hypothesis. A frictional term in the atmospheric equations damps out all high-frequency variability and therefore excludes the possibility of nonlinear interactions of high-frequency modes leading to low-frequency variability. The analysis on the atmospheric variables of the CSM output did not show any significant low-frequency variability. The only atmospheric mode found within the white noise was clearly related to internal ocean variability. However, because of the coarse grid in the CSM, these results can not be conclusive to decide whether part of the NAO variability is generated internally in the atmosphere.

The second hypothesis in section 1.2 stated that the atmospheric low-frequency variability was a pure response to internal ocean variability. In the following discussion of this hypothesis we directly discuss also the third hypothesis, which states that ocean-atmosphere feedbacks are crucial for the generation of significant atmospheric low-frequency variability. The first basic mechanism we found in chapters 2-3, namely the rectification processes leading to low-frequency variability, seems to favor the second hypothesis. Although in chapter 2 coupling seemed to be crucial for the generation of a large-scale coupled mode which was an essential component in generating the nonlinearly induced low-frequency variations, chapter 3 showed that such variability could already be explained by ocean processes alone. The NAO-like atmospheric variability in this model clearly was a direct response to SST-anomalies. The second basic mechanism we found in chapters 3 and 4, namely the gyre mode, is an internal ocean mode, thereby also consistent with the second hypothesis. However, chapter 4 showed that the atmosphere did not only respond to the low-frequency forcing from the ocean, but also became involved in a (weak) positive feedback. Probably due to this feedback, an atmospheric reflection of the gyre mode could be found in statistical analysis of the atmosphere alone. Therefore, also the third hypothesis is favored by these results.

**Personal view**

Recalling from section 1.2, Bladé (1997) and Saravanan (1998) examined the qualitative role of the coupling between the ocean and atmosphere on the low-frequency variability. They find that this coupling does not change the low-frequency spatial patterns of variability, but does modify its amplitude and temporal character. In chapter 4 we find similar behavior. If
the gyre mode would not have existed in the ocean, the atmospheric variability in the 300-year simulation of the CSM would be purely white noise, with a dominant spatial pattern resembling the NAO. However, coupling with the ocean revealed an atmospheric mode with a decadal time scale and NAO spatial pattern. Thus, coupling did not change the dominant spatial pattern, but did change its amplitude and temporal character. A decadal mode found by Selten et al. (1999), explained by a different physical mechanism also revealed similar behavior. Furthermore, atmospheric GCM’s always produce a NAO atmospheric spatial pattern of variability, but its amplitude and temporal behavior improve (compared to observations) if the lower boundary is forced with observed SST (Rodwell et al., 1999). Clearly, the NAO pattern is not only the dominant pattern of internal atmospheric variability, but also a pattern which is the most unstable to external forcings. And because many processes have shown to produce internally generated ocean modes on various time scales, many processes will offer lower boundary for the atmosphere changing the amplitude and temporal character of the NAO. Other studies (e.g. Lau and Nath (1990), Cessi (2000), Da Costa and Colin de Verdière (2002)), section 2.2.3 and chapter 4 have shown that the region east of Newfoundland is crucial for the atmospheric response to SST anomalies and thus internal ocean modes active in this region will be responsible for changing the NAO. The decadal mode described in chapter 4 explained over 65% of the variance of the barotropic streamfunction in this region and thus makes it a serious candidate for explaining that part of the NAO that can be explained (still the largest part being white noise).

Predictability of the NAO

Finally, we want to comment on the primary goal of the NAO-research: increasing the interseasonal to interannual predictability of the climate in the Atlantic region. The mechanism of rectification processes leading to low-frequency variability obviously must be examined in greater detail to be able to contribute to an increased predictability. The results presented here are very qualitative and amplitude and temporal behavior must be investigated in high-resolution elaborate models. But even then, it is a well-known fact that nonlinear processes in the climate system are very difficult to predict, especially if the process is subject to a very low signal-to-noise-ratio. The mechanism of the gyre mode leading to low-frequency atmospheric variability might eventually lead to increasing the predictability of the Atlantic climate system. It was shown that ocean characteristics in the Labrador Sea lead the NAO-pattern, which offers a clue to ocean monitoring and then to predicting the NAO. However, this will always remain a "the odds are 55% to 45%" kind of prediction that the NAO will show a certain phase. This randomly guessed 55% will not exceed the 50%-level very much, because the reflection of the gyre mode on the NAO did not reject the white noise null-hypothesis and furthermore only explains a small part of the total variability.

Furthermore, it is very likely that the two mechanisms both account for a part of the total atmospheric low-frequency variability and that other mechanisms do so as well. As all these mechanisms certainly will have influence on each others amplitude and phase, it will be a very hard job to come to a satisfying level of predictability of the Atlantic climate system.
Appendix A

Numerical methods

The equations and boundary conditions in chapters 2 and 3 are implemented within a continuation code as described in Dijkstra and Katsman (1997). The set of partial differential equations (2.1) to (2.10) is discretized on a $N \times M$ grid. After discretization a system of nonlinear algebraic equations results, which can be written as

$$M \frac{du}{dt} = F(u, \lambda) \quad (A.1)$$

Here $u$ is the $d$-dimensional state vector, consisting of the unknowns $(\Psi_1, \Psi_2, \psi_1, \psi_2, T)$ at the grid points, $\lambda$ is the $\lambda$-dimensional vector of parameters, $F$ is a nonlinear mapping from $\mathbb{R}^d \times \mathbb{R}^\lambda \to \mathbb{R}^d$ and $M$ is a linear operator. Stationary solutions satisfy the equation

$$F(u, \lambda) = 0$$

which is a system of $d$ nonlinear algebraic equations. To compute a branch of stationary solutions in a control parameter, say $\mu$, a pseudo-arclength method (Keller, 1977) is used. The branches of stationary solutions $(u(s), \mu(s))$ are parameterized by an ‘arclength’ parameter $s$. Since an additional equation is needed, the tangent is normalized along the branch,

$$\dot{u}_0^T (u - u_0) + \dot{\mu}_0 (\mu - \mu_0) - \Delta s = 0 \quad (A.2)$$

where $\Delta s$ is the step length and a dot indicates differentiation to $s$. Once the tangent $(\dot{u}, \dot{\mu}_0)$ is known, an initial guess $(u, \mu)$ along this tangent can be made. The Newton-Raphson method is used to converge to the branch of stationary solutions. This method finds isolated steady solutions, regardless of their stability.

When a steady state is determined, the linear stability analysis amounts to solving a generalized eigenvalue problem of the form

$$Ax = \sigma Bx \quad (A.3)$$

where $A$ is the Jacobian matrix (the derivative $F_u$) and $B = -M$, which are in general non-symmetric matrices. The problem (A.3) is solved by the Simultaneous Iteration method
Numerical methods

(S Stewart and Jennings, 1981). With this method, one can compute several, say \( m \), eigenvalues and optionally eigenvectors near a specified target.

A nice spin-off of steady state solvers is the immediate availability of an implicit time integration scheme. Using a time step \( \Delta t \), and a time index \( n \), a class of two-level schemes can be written as

\[
M \frac{u^{n+1} - u^n}{\Delta t} + \Theta F(u^{n+1}) + (1 - \Theta) F(u^n) = 0
\]  
(A.4)

For \( \Theta = 1/2 \), the Crank-Nicholson scheme is obtained. The equations for \( u^{n+1} \) are solved by the Newton-Raphson technique and lead to large systems of nonlinear algebraic equations, similar to that for the steady state computation. It is well-known that the Crank-Nicholson scheme is unconditionally stable for linear equations. This does not mean that one can take any time step, since this quantity is still constrained by accuracy. Although the scheme is second order accurate in time, large discretization errors occur when the time step is too large. Second limitation on the time step is the convergence domain of the Newton-Raphson process, which does not necessarily converge for every time step. For the model here, much larger time steps can be taken than with explicit time discretization.
Appendix B

Multi-channel Singular Spectrum Analysis (MSSA)

After applying the conventional EOF-analysis, the PC’s are used as input for MSSA. The reduced data set X now consists of a multichannel timeseries \( X_{l,i}, i = 1, \ldots, N; l = 1, \ldots, L \), where \( i \) represents time and \( l \) the channel number (PC). Making \( M \) lagged copies of \( X \), the state vector at time \( i \) is given by

\[
(X_{1,i+1}, X_{1,i+2}, \ldots, X_{1,i+M}, X_{2,i+1}, \ldots, X_{2,i+M}, \ldots, X_{L,i+1}, \ldots, X_{L,i+M}), \tag{B.1}
\]

where \( M \) is the window length. The cross-covariance matrix \( T \) for a chosen window length \( M \) has a general block-Toeplitz form in which each block \( T_{ll'} \) is the lag covariance matrix (with maximum lag \( M \)) between channel \( l \) and channel \( l' \). The \( L \times M \) real eigenvalues \( \lambda_k \) of the symmetric matrix \( T \) are sorted in decreasing order where an eigenvector (referred to as a ST-EOF) \( E^k \) is associated with the \( k \)th eigenvalue \( \lambda_k \). The \( E^k \) are \( M \)-long time sequences of vectors, describing space-time patterns of decreasing importance as their order \( k \) increases. A space-time principal component (referred to as a ST-PC) \( a^k \) can be computed by projecting \( X \) onto \( E^k \); \( \lambda_k \) is the variance in \( a^k \). In this way, the MSSA expansion of the original data series is given by

\[
X_{l,i+j} = \sum_{k=1}^{L \times M} a^k E^k_{ij}, j = 1, \ldots, M \tag{B.2}
\]

When two consecutive eigenvalues are nearly equal and the two corresponding \( E^k \) as well as the associated \( a^k \) are in quadrature, then the data possesses an oscillation with the period given by that of \( a^k \) and with the spatial pattern \( E^k \) (Plaut and Vautard, 1994). The sum in the right-hand side of (B.2), restricted to one or several terms, describes the part of the signal behaving as the corresponding \( E^k \). The components constructed in this way are called reconstructed components (RC’s). In this way, the part of the signal involved with an oscillation can be isolated.
As Allen and Robertson (1996) have pointed out, the presence of an eigenvalue pair is not sufficient ground to conclude that the data exhibit an oscillation. Moreover, low-frequency eigenvalue pairs, which are entirely due to red noise, will appear high in the eigenvalue rank-order. A Monte Carlo red-noise significance test for MSSA was therefore constructed (Allen and Robertson, 1996). This is an objective hypothesis test for the presence of oscillations at low signal-to-noise ratios in multivariate data. Rejection of the red-noise null hypothesis using the test should be considered a necessary condition for MSSA to have detected an oscillation, although in certain situations non-oscillatory processes might also lead to rejection. The test is built up as follows. Surrogate data segments are constructed by superposing \( L \) uncorrelated AR(1) processes having the same variance and lag-1 autocorrelation as the PC’s of the data set. Data and surrogate data segments are then projected onto a ST-PC basis of rank \( N - M + 1 \). This basis is derived from the AR(1) process cross-covariance (referred to as the null-hypothesis basis). The matrix of projections \( \Lambda \) is

\[
\Lambda = P^T YY^T P, \quad (B.3)
\]

where \( Y \) is the augmented data matrix of either red-noise surrogate data, or the sample time series, and \( P \) is the ST-PC basis. The method is described in more detail by Allen and Robertson (1996).


Summary

The winter-mean climate over the North Atlantic ocean shows a low pressure center near Iceland and a large high pressure center ranging from Florida to Spain. Between these centers, an eastward jet prevails, causing cold and dry winter conditions in Northeast America and moderate conditions in Western Europe. On this mean situation, variability shows up on various time scales. The dominant pattern of variability expresses as a weakening and strengthening of the mean pressure centers and is called the North Atlantic Oscillation (NAO). The impact of the NAO on the American and European winter climate is large and the understanding of a possible mechanism which drives the NAO from one phase into another is therefore of great importance. In the North Atlantic climate system, multiple subsystems appear to be highly related to the NAO. For instance, high correlations have been found between specific North Atlantic sea-surface temperature (SST) patterns of variability and the NAO. Therefore, it has often been suggested that the NAO is both an atmospheric and an oceanic phenomenon with possibly a crucial role for ocean-atmosphere coupling. Although the NAO fluctuates on all time scales, there seems to be a preference for low-frequency variability, ranging from a few years to a few decades.

In this thesis, we focus on the NAO at the low-frequency time scale. The central question is: How do oceanic and/or atmospheric processes cause low-frequency fluctuations at midlatitudes?

In literature, this central question has been approached in different ways:

(i) Several studies have analyzed observed data, from all kinds of atmospheric and oceanic measurements.

(ii) Analysis has been performed on very complex state-of-the-art coupled ocean-atmosphere models, which simulate the real climate system very well in many ways.

(iii) Studies have isolated those characteristics of the climate system, which are thought to be of crucial importance for low-frequency variability. have put these in a simple, or intermediate model and investigated the fundamental processes leading to NAO-like behavior.

By using observed data (i) or output of high-complexity models (ii), it is difficult to isolate the problem under investigation (the signal) from all the variability which is not related to this problem (the noise). By using simple or intermediate models (iii), it might be difficult
to relate the findings to the real, complex climate system. Therefore, in this thesis, a combination of (ii) and (iii) is applied. First, intermediate models are analyzed using a dynamical systems approach, showing the fundamental processes which lead to low-frequency variability at midlatitudes. Then, output of a complex coupled General Circulation Model (GCM) is analyzed, in order to explore whether these fundamental processes can be identified and understood in the more complex climate system.

The intermediate-complexity model used in chapter 2 assumes quasi-geostrophic dynamics in both ocean and atmosphere. The latter are coupled through processes controlling the sea-surface temperature within a surface layer of constant depth. Linear stability analysis of an idealized background state, characterized by zonal jets both in the atmosphere and ocean and a simplified sea-surface temperature profile, reveals that a low-frequency, large-scale mode may destabilize once the coupling strength is large enough. The mode corresponds to a near-stationary barotropic Rossby wave in the atmospherecoupled to a large-scale baroclinic oceanic Rossby wave. Computations of transient flows in the highly nonlinear regime show that nonlinear rectification processes of this coupled mode change the mean state in such a way that it becomes susceptible to high-frequency instabilities which, again through rectification processes, stabilize the coupled mode. The (generic) low-frequency variability which arises from these rectification processes is associated with amplification/weakening and North/South shifting of the zonal jets, very much resembling observed patterns of variability at midlatitudes.

One of the obvious idealizations in the model of chapter 2 is the zonally unbounded ocean. It is well known that continental boundaries severely restrict the mean ocean circulation patterns and wave propagation. Therefore, in the model used in chapter 3, the ocean is bounded and its idealized ocean background state is characterized by a double-gyre ocean circulation caused by the implied wind-stress pattern from the atmospheric jet. Linear stability analysis on the coupled system with a two-layer ocean model does not show any low-frequency modes. Transient flows show nonlinear rectification processes of intermonthly internal ocean modes leading to generic low-frequency variability. Coupling with the atmosphere damps this low-frequency variability. Both the linear stability analysis and transient flows of the uncoupled system in a one-layer configuration show an internal ocean mode, the so-called gyre mode, with a period of about a year. From the literature, this mode is known to exhibit a decadal period in multi-layer ocean models. The effect of coupling on this mode is examined but these results are not conclusive.

Adding the results of chapters 2 and 3, one can conclude that two mechanisms might generate low-frequency variability at midlatitudes: nonlinearly induced variability due to rectification processes and/or low-frequency variability arising through an internal ocean mode, the gyre mode.

In chapter 4, the output of a 300-year simulation with a state-of-the-art coupled GCM, namely the Climate System Model (version 1.0) is studied, again with a focus on decadal variability in the North Atlantic. Using Multichannel Singular Spectrum Analysis, a robust decadal oscillatory signal is found with specific spatial patterns of oceanic and atmospheric fields. In the ocean surface and subsurface fields, anomalies propagate from the northwestern Labrador Sea south-eastward along the coast and bend around Newfoundland before they enter the subtropical gyre. The pressure patterns in the atmosphere show a strong resemblance with those of the North Atlantic Oscillation. A comparison of the variability in the individual
oceanic and atmospheric fields and in combined sets of fields provides strong support that the oceanic temperature anomalies influence both the amplitude and phase of the NAO. Current theories on the physics of such variability are invoked to interpret the patterns of variability, but are found to be insufficient in several aspects. A new framework, in which the gyre mode found in chapter 3 is suggested to set the decadal time scale, is presented to explain the decadal variability.

The discussion of this thesis in chapter 5 starts with the fundamental processes leading to low-frequency variability at midlatitudes. The NAO-like signal due to rectification processes found in both models described in chapters 2 and 3, is not found in the output of the GCM, analyzed in chapter 4. However, because of the coarse resolution in the latter model, this comparison is not conclusive. Rectification processes might lead to NAO-like variability and must be examined in future studies. The internal ocean mode, the gyre mode, found in chapter 3, appeared to be of great importance in the GCM in chapter 4. It is a serious candidate to explain a significant part of the low-frequency variability and therefore must be submitted to additional analysis.

The results of this thesis and the literature on the NAO lead to a general (personal) view on the NAO: the low-frequency variations of the NAO may very well be caused by low-frequency variability in the ocean. It is shown that ocean-atmosphere coupling does not modify the spatial pattern of the NAO, but it does influence the amplitude of the NAO-pattern and its temporal character. This influence is mainly due to ocean-atmosphere interactions in the Newfoundland region. Because the ocean mode identified in chapter 4 dominates the decadal variability of SST in the Newfoundland region, it is a serious candidate to explain part of the low-frequency variations of the NAO.
Samenvatting

Het winterklimaat van West-Europa en Noordoost-Amerika wordt voornamelijk bepaald door het luchtdrucksysteem boven de Noord-Atlantische Oceaan. Gemiddeld ligt er een lagedrukgebied bij IJsland en een uitgestrekt hogedrukgebied van Florida tot Spanje. Hiertussen stroomt een oostwaartse straalstroom (een zogenaamde "jet"), die zorgt voor koude, droge lucht in Noordoost-Amerika en zachte omstandigheden in West-Europa. Dit dipoolpatroon in de luchtdruk vertoont fluctuaties, die zorgen voor een versterking dan wel verzwakking van de jet, en wordt de Noord-Atlantische Oscillatie (NAO) genoemd. De NAO heeft grote invloed op het Europese en Noord-Amerikaanse winterklimaat en het is daarom belangrijk om het mechanisme te begrijpen, dat de overgang van een sterke jet (positieve NAO-fase genoemd) naar een zwakke jet (negatieve NAO-fase) verklaart.

De variabiliteit van de NAO is gerelateerd aan fluctuaties in atmosfeer en oceaan. Zo is er bijvoorbeeld een hoge correlatie gevonden tussen de Noord-Atlantische luchtdruk op zeeniveau en de oppervlaktetemperatuur van het zeewater. De NAO wordt daarom tegenwoordig vaak gezien als zowel een atmosfeer- als een oceaanfenomeen met mogelijk een cruciale rol voor de oceaan-atmosfeerkoppeling. Alhoewel de NAO fluctueert op alle tijdschalen, blijkt er een voorkeur te zijn voor de laagfrequente variaties, van enkele jaren tot enkele decades.

In dit proefschrift ligt de nadruk op de variaties van de NAO op deze lange tijdschaal. De centrale vraagstelling is: Hoe kunnen oceaan- en/of atmosfeerprocessen laagfrequente variabiliteit op gematigde breedte veroorzaken? Bij de behandeling van deze vraag in de literatuur zijn er verschillen in de aanpak:

(i) Diverse studies hebben waargenomen data van verschillende atmosfeer- en oceaanmetingen geanalyseerd.

(ii) Er zijn analyses verricht op de output van zeer gecompliceerde, "state-of-the-art" gekoppelde oceaan-atmosfeer computermodellen, die vele facetten van het klimaatsysteem goed simuleren.

(iii) In diverse studies worden die karakteristieken van het klimaatsysteem die belangrijk worden geacht voor laagfrequente variabiliteit geïsoleerd. Deze worden vervolgens geïmplementeerd in relatief eenvoudige modellen, waarmee de fundamentele processen die NAO-achtige variabiliteit genereren worden onderzocht.

In het gebruik van waarnemingen (i) of output van complexe modellen (ii) is het altijd moeilijk om het te onderzoeken signaal te onderscheiden van de rest van de variabiliteit (de
Samenvatting


Het relatief eenvoudige model in hoofdstuk 2 gaat uit van de quasi-geostrofe benadering, zowel in oceaan als atmosfeer. De oceaan-atmosfeerkoppeling vindt plaats via de zeewater-temperatuur in een oppervlakte laag met constante diepte. De geïdealiseerde achtergrondstoestand van het model wordt gekarakteriseerd door zonale jets in oceaan en atmosfeer en een toenemende oppervlaktetemperatuur van noord naar zuid. Lineaire stabiliteitsanalyse op deze achtergrondstoestand laat zien dat een grootschalige, laagfrequente "mode" wordt gede-stabiliseerd als de oceaan-atmosfeerkoppeling sterk genoeg is. De mode bestaat uit een bijna stationaire, barotrope Rossby-golf in de atmosfeer, gekoppeld aan een grootschalige, baroklino Rossby-golf in de oceaan. Lineaire stabiliteitsanalyse laat zien dat deze hoogfrequente instabiliteit genereert die, wederom door rectificatieprocessen, de gekoppelde mode stabiliseren. De laagfrequente variabiliteit die door deze verschillende rectificatieprocessen ontstaat, veranderen, uiteindelijk de gemiddelde toestand en leiden tot grootschalige instabiliteiten. Deze variabiliteit wordt gedempt door een sterker koppeling van oceaan en atmosfeer. Een interne laagfrequente oceaanmode, de zogenaamde "gyre mode", ontstaat in de lineaire stabiliteitsanalyse en in tijdsintegraties, in een modelconfiguratie met een enkele oceaanlaag. Het effect van oceaan-atmosfeerkoppeling op deze mode wordt bekeken in hoofdstuk 3, maar de resultaten zijn onduidelijk.

Het gecombineerde resultaat van hoofdstuk 2 en 3 leidt tot twee fundamentele mechanismen die laagfrequente variabiliteit veroorzaken: niet-lineaire rectificatieprocessen en variabiliteit door een interne oceaanmode, de gyre mode.

In hoofdstuk 4 wordt de output van een 300-jaar simulatie met een "state-of-the-art", gekoppeld klimaatmodel, namelijk het Climate Systems Model, bestudeerd. Met "Multichannel Singular Spectrum Analysis" wordt een robuust, decadaal, oscillatorisch signaal gevonden met specifieke patronen in atmosfeer en oceaan. In de oceaan propageren anomalieën vanuit de noordwestelijke Labradorzee naar zuidoostwaarts, langs de Noord-Amerikaanse kust, rondom
Newfoundland, richting Florida. De anomalie luchtdrukpatronen lijken sterk op het fluctuerende dipoolpatroon van de NAO. Een vergelijking van de variabiliteit in de oceaan- en atmosfeervelden wijst op een sturende rol voor de oppervlaktetemperatuur van de oceaan, voor zowel de fase als de amplitude van het NAO-achtige signaal. De ruimtelijke patronen van het decadale signaal worden verklaard met behulp van theorieën uit recente studies. We suggereren dat de gyre mode de tijdschaal verklaart.

In hoofdstuk 5 worden eerst de fundamentele processen die leiden tot laagfrequent variabiliteit op gematigde breedte bediscussieerd. Het NAO-achtige signaal uit hoofdstuk 2 en 3, veroorzaakt door niet-lineaire rectificatieprocessen, wordt niet gevonden in de output van het model in hoofdstuk 4. Echter, deze vergelijking leidt niet tot een definitieve conclusie, vanwege de grove resolutie van het complexe model. De laagfrequent gyre mode, gevonden in hoofdstuk 3, speelt ook een belangrijke rol in het decadale signaal uit hoofdstuk 4.

De resultaten die worden gepresenteerd in dit proefschrift leiden, samen met de bestaande literatuur over dit onderwerp, tot een algemene, persoonlijke visie op de NAO: de laagfrequente variaties van de NAO zouden goed veroorzaakt kunnen worden door processen in de oceaan. De koppeling van oceaan en atmosfeer verandert niet het ruimtelijke patroon van de NAO, maar beïnvloedt wel de amplitude van het patroon en het tijdsafhankelijke gedrag. Deze beïnvloeding komt vooral door oceaan-atmosfeer interactie in het gebied bij Newfoundland en omdat de decadale mode uit hoofdstuk 4 de variaties van de zeewaterstemperatuur in dit gebied sterk bepaalt, is deze mode een belangrijke kandidaat voor het verklaren van een gedeelte van de laagfrequent variabiliteit van de NAO.
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