Climatic Variability over the North Atlantic

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"Meteorology at the Millennium": Contributions from Invited Speakers

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Abstract

This paper summarizes an invited presentation given at the historic "Meteorology at the Millennium" Conference in July 2000, which marked the 150th anniversary of the Royal Meteorological Society. It begins with a broad review of the North Atlantic Oscillation (NAO) and the mechanisms that might influence its phase and amplitude on decadal and longer time scales. New results are presented which suggest an important role for tropical ocean forcing of the unprecedented trend in the wintertime NAO index over the past several decades. We conclude with a brief discussion of a recent significant change in the pattern of the summertime atmospheric circulation over the North Atlantic.

1. Introduction

The climate of the Atlantic sector and surrounding continents exhibits considerable variability on a wide range of time scales. It is manifested as coherent fluctuations in ocean and land temperature, rainfall and surface pressure with a myriad of impacts on society and the environment. One of the best-known examples of this variability is the North Atlantic Oscillation (NAO). Over the past several decades, an index of the wintertime NAO has exhibited a strong upward trend. This trend is associated with a strengthening of the middle latitude westerlies and anomalously low (high) surface pressure over the subpolar (subtropical) North Atlantic. Given the large impact of NAO variability on regional temperatures and precipitation, as well as on Northern Hemisphere (NH) temperature changes in general, understanding the physical mechanisms that govern the NAO and its variability on all time scales is of high priority.

In this paper we present a review of the NAO and the processes that might influence its phase and amplitude, especially on decadal and longer time scales. In addition, some new results that suggest a link between the recent upward trend in the wintertime NAO index and warming of the tropical oceans will be highlighted. Concluding comments are made that focus attention on decadal variability of the extratropical North Atlantic climate during northern summer.   The discussion is aimed at a scientifically diverse audience, such as the one that made the "Meteorology at the Millennium" conference a large success.

2. What is the NAO and how does it impact regional climate?

Monthly mean surface pressures vary markedly about the long-term mean sea level pressure (SLP) distribution. This variability occurs in well-defined spatial patterns,

particularly during boreal winter over the Northern Hemisphere when the atmosphere is dynamically the most active. These recurrent patterns are commonly referred to as "teleconnections" in the meteorological literature, since they result in simultaneous variations in weather and climate over widely separated points on earth. One of the most prominent patterns is the NAO. Meteorologists have noted its pronounced influence on the climate of the Atlantic basin for more than two centuries.

The NAO refers to a north-south oscillation in atmospheric mass between the Icelandic subpolar low- and the Azores subtropical high-pressure centers. It is most clearly identified when time-averaged data (monthly or seasonal) are examined, since time averaging reduces the "noise" of small-scale and transient meteorological phenomena not related to large-scale climate variability. The spatial signature and temporal variability of the NAO are usually defined through the regional SLP field, for which some of the longest instrumental records exist, although it is also readily apparent in meteorological data through the lower stratosphere.

The spatial structure and a time series (or index) of more than 100 years of NAO variability are shown in **Fig. 1**[1]. Although the NAO is present throughout much of the year, this plot illustrates conditions during northern winter when the NAO accounts for more than one-third of the total variance in SLP over the North Atlantic, far more than any other pattern of variability. Differences of more than 15 hPa in SLP occur across the North Atlantic between the two phases of the NAO in winter. In the so-called positive phase, higher than normal surface pressures south of 55°N combine with a broad region of anomalously low pressure throughout the Arctic (**Fig. 1a**). Consequently, this phase of

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[1] More sophisticated and objective statistical techniques, such as eigenvector analysis, yield time

the oscillation is associated with stronger-than-average westerly winds across the middle latitudes of the Atlantic onto Europe, with anomalous southerly flow over the eastern United States and anomalous northerly flow across western Greenland, the Canadian Arctic, and the Mediterranean. The easterly trade winds over the subtropical North Atlantic are also enhanced during the positive phase of the oscillation.

This anomalous flow across the North Atlantic during winter transports anomalous warm (and moist) maritime air over much of Europe and far downstream across Asia during the positive phase of the oscillation, while the southward transport of polar air decreases land and sea surface temperatures (SSTs) over the northwest Atlantic. Anomalous temperature variations over North Africa and the Middle East (cooling), as well as North America (warming), associated with the stronger clockwise flow around the subtropical Atlantic high-pressure center are also notable during high-index NAO winters.

The changes in the mean circulation patterns over the North Atlantic during extreme phases of the NAO are accompanied by changes in the intensity and number of storms, their paths, and their associated weather. The details of changes in storminess differ depending on the analysis method and whether one focuses on surface or upper-air features. Generally, however, positive NAO index winters are associated with a northward shift in the Atlantic storm activity, with enhanced storminess from southern Greenland across Iceland into northern Europe and a modest decrease in activity to the south. The latter is most noticeable from the Azores across the Iberian Peninsula and the

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series and spatial patterns of average winter SLP variability very similar to those shown in **Fig. 1.**

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Mediterranean. Positive NAO winters are also typified by more intense and frequent storms in the vicinity of Iceland and the Norwegian Sea.

Changes in the mean flow and storminess associated with swings in the NAO change the transport and convergence of atmospheric moisture over the North Atlantic and, thus, the distribution of precipitation. Drier-than-average conditions prevail over much of Greenland and the Canadian Arctic during high NAO index winters, as well as over much of central and southern Europe, the Mediterranean and parts of the Middle East. In contrast, more precipitation than normal falls from Iceland through Scandinavia.

Given this large impact on regional climate, improved understanding of the process or processes that govern variability of the NAO is an important goal, especially in the context of global climate change. At present there is little consensus on the mechanisms that produce NAO variability, especially on decadal and longer time scales. It is quite possible one or more of the mechanisms described below affects this variability.

3. What are the Mechanisms that Govern NAO Variability?

Atmospheric processes

Atmospheric general circulation models (AGCMs) provide strong evidence that the basic structure of the NAO results from the internal, nonlinear dynamics of the atmosphere. The observed spatial pattern and amplitude of the NAO are well simulated in AGCMs forced with climatological annual cycles of solar insolation and SST, as well as fixed atmospheric trace-gas composition. The governing dynamical mechanisms are interactions between the time-mean flow and the departures from that flow. Such intrinsic atmospheric variability exhibits little temporal coherence and, indeed, the time scales of

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observed NAO variability do not differ significantly from this reference. Large changes in the atmospheric circulation over the North Atlantic occur from one winter to the next, and there is also a considerable amount of variability within a given winter season. In this context, any low frequency NAO variations in the relatively short instrumental record could simply reflect finite sampling of a purely random process.

A possible exception to this interpretation is the strong trend toward the positive index polarity of the NAO over the past 30 years (**Fig. 1b**). This trend exhibits a high degree of statistical significance relative to the background interannual variability in the observed record; moreover, multi-century AGCM experiments forced with climatological SSTs and fixed atmospheric trace-gas composition do not reproduce interdecadal changes of comparable magnitude.

The equivalent barotropic vertical structure of the NAO, reaching into the stratosphere, suggests that it could be influenced by the strength of the atmospheric circulation in the lower stratosphere. The leading pattern of geopotential height variability in the lower stratosphere is also characterized by a seesaw in mass between the polar cap and the middle latitudes, but with a much more zonally symmetric (or annular) structure than in the troposphere. When heights over the polar region are lower than normal, heights at nearly all longitudes in middle latitudes are higher than normal. In this phase, the stratospheric westerly winds that encircle the pole are enhanced and the polar vortex is "strong" and anomalously cold. It is this annular mode of variability that has been termed

the Arctic Oscillation (AO)[2] , and the aforementioned conditions describe its positive phase.

During winters when the stratospheric AO is positive, the NAO tends to be in its positive phase. There is a considerable body of evidence to support the notion that variability in the troposphere can drive variability in the stratosphere. New observational and modeling evidence, however, suggests that some stratospheric control of the troposphere may also be occurring.

The atmospheric response to strong tropical volcanic eruptions provides some evidence for a stratospheric influence on the earth's surface climate. Volcanic aerosols act to enhance north-south temperature gradients in the lower stratosphere by absorbing solar radiation in lower latitudes. In the troposphere, the aerosols exert only a very small direct influence. Yet, the observed response following eruptions is not only lower geopotential heights over the pole with stronger stratospheric westerlies, but also a strong, positive NAO-like signal in the tropospheric circulation.

Reductions in stratospheric ozone and increases in greenhouse gas concentrations also appear to enhance the meridional temperature gradient in the lower stratosphere, leading to a stronger polar vortex. It is possible, therefore, that the upward trend in the NAO index in recent decades (**Fig. 1b**) is associated with trends in either or both of these quantities. Indeed, a decline in the amount of ozone poleward of 40°N has been observed

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[2] The signature of the stratospheric AO in winter SLP data, however, looks very much like the anomalies associated with the NAO with centers of action over the Arctic and the Atlantic. The "annular" character of the AO in the troposphere, therefore, reflects the vertically coherent fluctuations throughout the Arctic more than any coordinated behavior in the middle latitudes outside of the Atlantic basin. That the NAO and AO reflect essentially the same mode of tropospheric variability is emphasized by the similarity of their time series, with differences depending mostly on the details of the analysis procedure.

during the last two decades, and the stratospheric polar vortex has become colder and stronger.

Ocean forcing of the atmosphere

Over the North Atlantic, the leading pattern of SST variability during winter consists of a tri-pole. It is marked, in one phase, by a cold anomaly in the subpolar North Atlantic, warmer-than-average SSTs in the middle latitudes centered off of Cape Hatteras, and a cold subtropical anomaly between the equator and 30°N. This structure suggests that the SST anomalies are primarily driven by changes in the surface wind and air-sea heat exchanges associated with NAO variations. Indeed, the relationship is strongest when the NAO index leads an index of the SST variability by several weeks. Over longer periods, persistent SST anomalies also appear to be related to persistent anomalous patterns of SLP (including the NAO), although the mechanisms which produce SST changes on decadal and longer time scales remain unclear. Such fluctuations could primarily be the local oceanic response to atmospheric decadal variability. On the other hand, non-local dynamical processes in the ocean could also be contributing to the SST variations and, perhaps, the ocean anomalies could be modifying the atmospheric circulation.

A key and long-standing issue in this regard has been the extent to which anomalous extratropical SST and upper ocean heat content anomalies feed back to affect the atmosphere. Most evidence suggests this effect is quite small compared to internal atmospheric variability; for instance, AGCM studies typically exhibit weak responses to extratropical SST anomalies, with sometimes-contradictory results. Yet, some AGCMs, when forced with the time history of observed, global SSTs and sea ice concentrations over the past 50 years or so, show modest skill in reproducing aspects of the observed

NAO behavior, especially its interdecadal fluctuations. One example is given in **Fig. 2**.

In this plot, the NAO is defined as the spatial structure function corresponding to the first empirical orthogonal function (EOF) of monthly 500 hPa geopotential height anomalies. The principal component time series have been smoothed with a 73-month running mean filter in order to emphasize the low frequency NAO variations. The model data are from a 12-member ensemble of simulations performed with an AGCM (version 3.6 of the NCAR Community Climate Model, known as CCM3) over 1950-1994. Ensemble experiments are analyzed because any individual simulation is dominated by internal atmospheric variability in the extratropics. Since the time history of observed SSTs is specified as a lower boundary condition at all ocean gridpoints, experiments of this type are often referred to as Global Ocean Global Atmosphere (GOGA) integrations.

The key point in **Fig. 2** is that the low-frequency ensemble-mean NAO time series correlates with the observed at 0.76, which is significant at the 5% level, and the overall upward trend is captured. The amplitude of the simulated NAO variability is about one-half that of the observed. This is a consequence of the ensemble averaging, as the simulated NAO in individual members of the ensemble have realistic amplitudes. These results, which have been found in other recent studies as well, suggest that a low frequency component of the NAO over this period is not merely stochastic climate noise, but a response to variations in global SSTs. They do not necessarily imply, however, that the extratropical ocean is behaving in anything other than a passive manner. It could be, for instance, that long-term changes in tropical SSTs force a remote atmospheric response over the North Atlantic, which in turn drives changes in extratropical SSTs and sea ice. Indeed, additional experiments with CCM3 add merit to this viewpoint.

To isolate the role of changes in tropical SSTs, another ensemble of simulations was performed. In these experiments, the time history of observed SSTs was specified over tropical latitudes (30°S-30°N), but CCM3 was forced by monthly climatological values of SST and sea ice at higher latitudes. These experiments are known as Tropical Ocean Global Atmosphere (TOGA) integrations. Finally, a third set of integrations (Tropical Atlantic Global Atmosphere, or TAGA) were designed to separate the role of tropical Atlantic SST variability by forcing CCM3 with climatological SSTs everywhere outside of the tropical Atlantic. The TOGA and TAGA results shown below are from 5-member ensembles.

To illustrate the spatial pattern of the dominant trends in NH wintertime atmospheric circulation over the past several decades, shown are the differences in 500 hPa geopotential heights between 1977-94 and 1950-76 (**Fig. 3**). The observed changes not only reflect the upward trend in the NAO, but are also characterized by significantly lower heights over the central North Pacific and higher heights over western Canada[3]. Results from the GOGA ensemble show that CCM3 captures many, but not all, aspects of these hemispheric changes in the wintertime circulation. For instance, over the North Atlantic the height changes associated with the trend in the NAO are well captured (as

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[3] The changes over the Pacific correspond to an enhancement of another prominent mode of atmospheric circulation variability known as the Pacific-North American (PNA) teleconnection pattern.

expected from **Fig. 2**), but the model extends the negative height anomalies too far to the east.

Perhaps the most striking result in **Fig. 3,** however, is that the low-frequency changes in 500 hPa heights over the North Atlantic are recoverable from tropical SST forcing alone. The spatial pattern of height change in the TOGA ensemble closely matches that of the GOGA ensemble (the slightly larger amplitudes of the former probably reflect the smaller ensemble size). Tropical (30°S-30°N) SSTs exhibit a warming trend of about 0.2° C since 1950 and this trend is common to all ocean basins. A linear regression of CCM3 500 hPa TOGA ensemble-mean heights onto a low-pass filtered version of the tropical SST time series projects strongly onto the spatial structure of the leading EOF in **Fig. 2**, which suggests that, at least for this AGCM, low frequency variations in the NAO originate from the tropics. Under this scenario, the link between the extratropical Atlantic SSTs and the NAO could be largely diagnostic in the sense that the externally-forced NAO variations are themselves contributing to North Atlantic SST variations. The process may be analogous to the atmospheric bridge mechanism that has been invoked for the seasonal variability of North Pacific SSTs driven by teleconnections from the tropical Pacific.

Some recent studies using other AGCMs have suggested variations in tropical Atlantic SSTs significantly modulate NAO variability. The TAGA results in **Fig. 3,** however, do not suggest a strong influence by the tropical Atlantic. The simulated low frequency height changes project only weakly onto the spatial structure of the NAO, and the principal component time series of the leading EOF from the TAGA ensemble (not shown) has little correlation with the observed time series (**Fig. 2)**, unlike the GOGA and

TOGA results**.**

The pattern of changes in tropical rainfall between 1977-94 and 1950-76 are shown in **Fig. 4** for the TOGA and TAGA ensemble means. The warming trend in tropical SSTs is associated with increases in (TOGA) simulated rainfall (and thus latent heating) over the western and central tropical Pacific, as well as over the tropical Indian Ocean. Over the Caribbean and northern South America, decreases in rainfall are simulated, but these decreases are much weaker in the TAGA ensemble. This suggests that the heating anomalies in these regions are primarily in response to tropic-wide changes in circulation associated with SST forcing outside of the tropical Atlantic. We are currently working to firmly establish the dynamics of this tropical teleconnection and its possible impact on North Atlantic climate variability.

The response of the extratropical North Atlantic atmosphere to changes in tropical and extratropical SST distributions, and the role of land processes and sea ice in producing atmospheric variability, are problems which are currently being addressed by the community. Until these are better understood, it is difficult to evaluate the realism of more complicated scenarios that rely on truly coupled interactions between the atmosphere, ocean, land and sea ice to produce North Atlantic climate variability. It is also difficult to evaluate the extent to which interannual and longer-term variations of the NAO might be predictable.

4. Concluding comments on other aspects of North Atlantic climate variability

While the NAO behavior shown in **Fig. 1** explains a substantial portion of the low-frequency climate variability over the North Atlantic, the fact that most NAO variability is concentrated at higher (weekly to monthly) frequencies means that there are significant monthly-to-interannual departures from the long-term NAO trend. The winter of 1996, for instance, was characterized by a very negative NAO index (in sharp contrast to the long-term trend toward positive index values), and winter conditions over much of Europe were severe. However, the anomalous anticyclonic circulation was located well to the east of the canonical NAO pattern, with positive SLP anomalies of more than 9 hPa centered over Scandinavia. Such persistent high pressure anomalies are a typical feature of the North Atlantic climate and are referred to as "blocks". Longitudes near 15°W are particularly favored locations for the development of blocking highs, which occur when the westerly flow across the middle latitudes of the NH is weak and typified by an exaggerated wave pattern.   North Atlantic blocking is most typical during boreal spring and early summer, but it occurs throughout the year.

Over recent decades, for instance, increased anticyclonicity and easterly flows from the relatively warm, summer continent have brought anomalously warm and dry conditions to much of the United Kingdom. This change in circulation is illustrated in **Fig. 5**, which shows the leading EOF of summer SLP over the North Atlantic. The associated principal component time series is remarkable for its low-frequency variations, including the persistence of high-pressure anomalies over northern Europe during most summers since the late 1960s.   Changes in summertime circulation are especially important from the perspective of droughts and heat waves, which have large societal impacts. Indeed, the

drying during July-August is particularly evident in England and Wales rainfall, which has exhibited a strong downward trend since about 1960 (not shown). Some of the driest summers over the United Kingdom in recorded history have occurred during this most recent period.

The need consider the possible links between the summertime climate anomalies noted above to changes in other, more remote, regions of the globe was first illustrated by the work of Chris Folland and colleagues nearly 15 years ago. Their studies noted a strong statistical association between climate variations over the UK region and rainfall over the Sahel region during northern summer. The association is particularly prominent on decadal time scales and it appears to have existed over much of the twentieth century. This intriguing link has not been explained to date. As such, it is also a focus of our current work.

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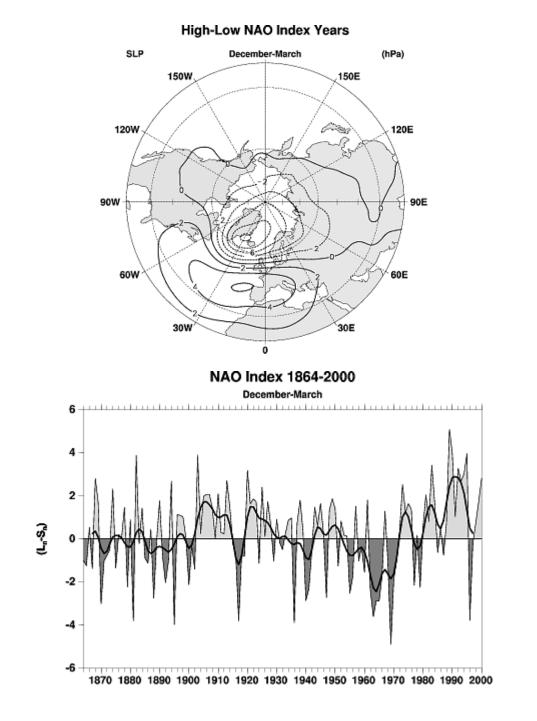
Figures and Captions

High-Low NAO Index Years

NAO Index 1864-2000

Figure 1. (top) Difference in sea level pressure between winters (December-March)

with an NAO index value > 1.0 and those with an index value < −1.0 (high minus low index winters) since 1899. The contour increment is 2 hPa and negative values are dashed. The index (bottom) is based on the difference of normalized sea level pressure between Lisbon, Portugal and Stykkisholmur/Reykjavik, Iceland from 1864 through 2000. The heavy solid line represents the index smoothed to remove fluctuations with periods less than 4 years.
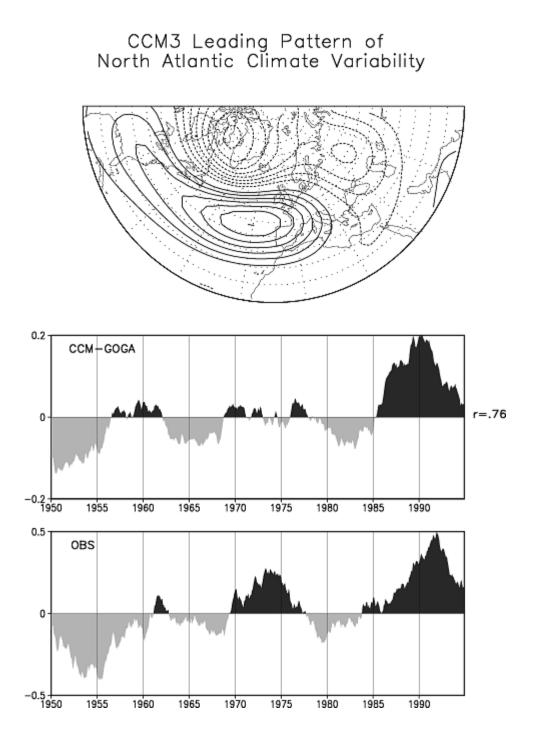
Figure 2. Leading empirical orthogonal function of the CCM3 GOGA ensemble-mean monthly 500 hPa geopotential height for the region 100°W-80°E, 20-90°N (top, see text for details). The associated principal component time series (middle), smoothed with a 73-month running mean, correlates with the observed (bottom) at 0.76.
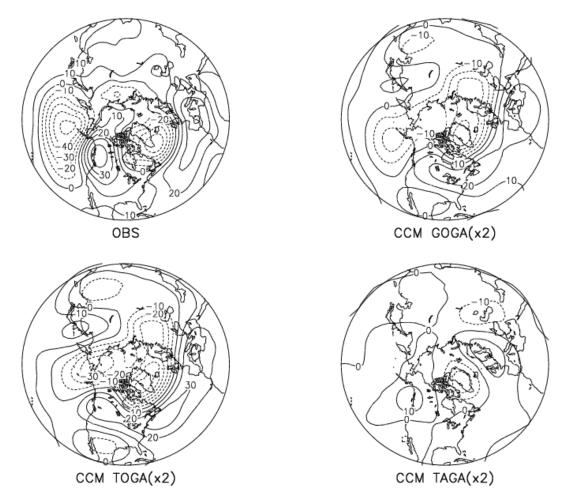
Figure 3. Changes in 500 hPa geopotential height (m), between 1977-94 and 1950-76, for the observations and the CCM3 ensemble-mean experiments described in the text. The model changes have been multiplied by a factor of 2, and only winter (December-March) conditions are illustrated. The contour increment is 10 m.
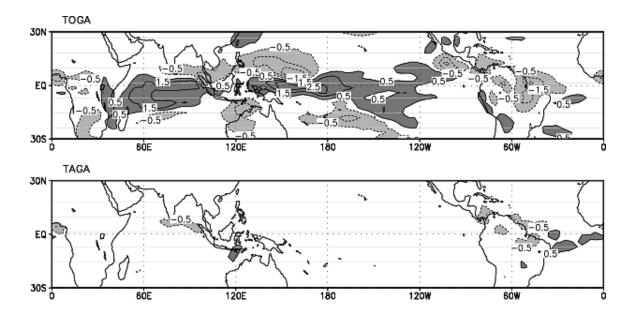
Figure 4. Changes in winter (December-March) total precipitation (mm day^{-1}), between 1977-94 and 1950-76, for the TOGA (top) and TAGA (bottom) ensemble means (see text for details). The contour increment is 0.5 mm day^{-1}.
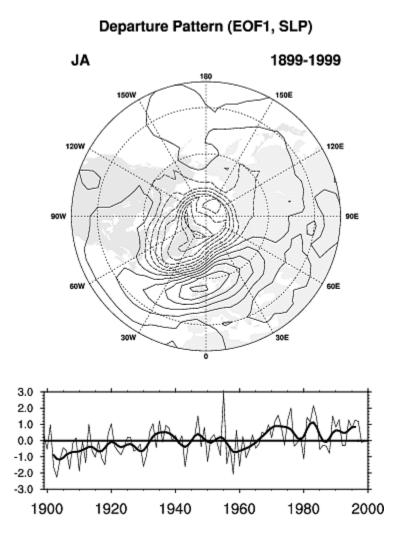
Figure 5. The leading empirical orthogonal function of July-August sea level pressure (top) and the associated principal component time series (bottom) over the domain 60°W-30°E, 30-80°N and period 1899-1999. The contour increment is 0.4 hPa, and the contours correspond to the change in sea level pressure associated with a one standard deviation change of the principal component time series. Negative contours are dashed.