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VARIABILITY OF THE NORTH ATLANTIC SUBTROPICAL HIGH AND ASSOCIATIONS WITH TROPICAL SEA-SURFACE TEMPERATURE

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ABSTRACT

Based on generated time series of the central pressure of the subtropical high, the behaviour of this atmospheric centre of action has been examined since 1950 with regard to the inter-annual variations, persistence, linear trends, abrupt change, spectral analysis and interactions. The year-to-year variations in the central pressure of the subtropical high are considerable. The pronounced strengthening of the subtropical high during approximately the last 20 years is its most remarkable feature. Variations in the subtropical high's intensity seem to follow a cyclic pattern. Statistically significant abrupt onset events are found, with the majority occurring in winter. Increasing and decreasing episodes have occurred in the late 1960s and early 1980s. According to spectral analysis, it can be assumed that the centre of action of the subtropical high is characterized by non-periodic behaviour. The peaks occur only at the lowest frequency. The quasi-biennial oscillation may affect the subtropical high in the winter season, whereas the North Atlantic Ocean may affect the subtropical high in summer. There is no detectable connection between El Niño–southern oscillation and the subtropical high in either season. The subtropical high has been affected by sea-surface temperatures of the north and south tropical Atlantic, with negative and positive signs respectively in the winter season. Interactions between the subtropical high and sea-surface temperatures over the tropical Atlantic Ocean are absent in the summer season.

KEY WORDS: subtropical high; SST; North Atlantic Ocean; South Atlantic Ocean; tropical Atlantic Ocean; ENSO; QBO; long-term variations

1. INTRODUCTION

Within the context of the discussion on climate change, the diagnostics of long-term circulation variability play an important role, particularly as the Intergovernmental Panel on Climate Change in 1992 termed this problem a 'key topic for future research' (Houghton *et al.*, 1992). Studies based on long-term observations are still rare. In spite of the abundant research, only a few investigations can be mentioned that are of relevance to this study. A study by Walker and Bliss (1932), which describes the phenomenon of the North Atlantic oscillation (NAO), is particularly prominent. In subsequent years several researchers have sought to characterize the variability and intensity of the North Atlantic or northern hemispheric circulation by means of differently defined indices, such as the zonal index or index cycle. Similarly, the pressure differences between stations in the Azores and Iceland are used to characterize the variability of the NAO (e.g. Rossby, 1939; Rossby and Willett, 1948; Namias, 1950; Lorenz, 1951; Bjerknes, 1964; Rogers, 1984; Kidson, 1985; Wallace and Hsu, 1985; Moses *et al.*, 1987; Robinson, 1991).

Recent modelling studies demonstrate the possibility that an ocean–atmosphere feedback mode may produce such anti-symmetries on the decadal time scale (e.g., Chang *et al.*, 1997; Xie, 1999). The existence of such a

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dipole is controversial, however. Dommenget and Latif (2000) suggest that tropical North Atlantic and tropical South Atlantic sea-surface temperature (SST) anomalies occur independently and dipole configurations are not ubiquitous. Nevertheless, the index formed by the SST anomaly difference between the tropical North Atlantic and tropical South Atlantic (Servain, 1991), sometimes referred to as a dipole index, has proved to be an invaluable indicator of the meridional gradient in SST anomaly that is correlated with strong climate anomalies over the surrounding land regions. Moreover, much evidence exists for an ocean–atmosphere coupling — stronger in the tropical North Atlantic region — that causes reinforcement and persistence of meridional gradient anomalies (Xie, 1999). This meridional anti-symmetry appears to work preferentially on the decadal time scale (Enfield *et al.*, 1999) and it had been thought that it explains little of the total SST anomaly variance.

Despite the subtropical high being the dominant atmospheric circulation system in the lower troposphere and controlling the whole of the east/west and mid-Atlantic (Figure 1), it has not attracted much attention. This is despite the evidence showing that the subtropical high pressure also exerts a powerful influence on climate over middle latitudes. The aim of this paper is to investigate the variability of the subtropical high centre of action and its relationship with the SSTs over the tropical Atlantic and east equatorial Pacific El Niño–southern oscillation (ENSO) in winter (December, January, and February) and summer (June, July, and August) seasons of the period 1950–2002. This paper is organized as follows: after this introduction the data and the methods used for the evaluation of the central pressure action of the subtropical high are described in sections 2 and 3 respectively. Section 4 is focused on the statistical characteristics of the standardized anomaly behaviour of the centre, of the persistence of these standardized anomalies, and of long-term trends on an interannual scale. In Section 5 the interaction between the index of the subtropical high's central intensity (SHCI) and SSTs is discussed. The conclusions of this investigation are summarized in Section 6.

2. DATA

The major data source in this study is the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis fields from January 1950 to December 2002. The NCEP–NCAR reanalysis fields use a state-of-the-art global data assimilation system on a 2.5° longitude by 2.5° latitude grid. Monthly sea-level pressure (SLP) and monthly SST data were used in this study. Figure 1 shows the climatology means of SLP during the winter (December to February) and summer (June to August) over the area bounded by 80 °W–40 °E and 15–50 °N. The most pronounced feature is that the surface circulation is dominated by a huge subtropical high-pressure atmospheric centre of action. This strong anticyclone circulation system is centred over the eastern Atlantic in winter and over the western Atlantic in summer (figure 1(a) and (b) respectively). The quantitative index of the SHCI is defined as the regional mean SLP averaged over the area 15–35 °W and 27–37 °N in winter and 28–45 °W and 30–38 °N in summer; this provides a measure of the strength of the subtropical high. These rectangular areas generally cover the central regions of the anticyclone where the pressure is generally greater than 1021 hPa in winter and 1024 hPa in summer.

SST indices were obtained from the Climate Prediction Center (National Oceanic and Atmospheric Administration (NOAA), USA). These are the NINO3 index (5 °N–5 °S, 150–90 °W), a widely used ENSO indicator (Camberlin *et al.*, 2001); an equatorial South Atlantic index (SATL; 0–20 °S, 30 °W–10 °E); a tropical North Atlantic index (NATL; 5–20 °N, 60–30 °W); and a tropical Atlantic index (TATL; 10 °S–10 °N, 0–360°).

3. METHODS

A general overview of the central pressure is given by deriving a time series of standardized pressure anomalies (1950–2002). To visualize the decadal and interdecadal fluctuations or 'persistence' in the behaviour of the circulation centre, the cumulative annual means method is used (Pavia and Graef, 2002), because this advantageously reveals time-varying structures in a time series. The cumulative annual mean time series can

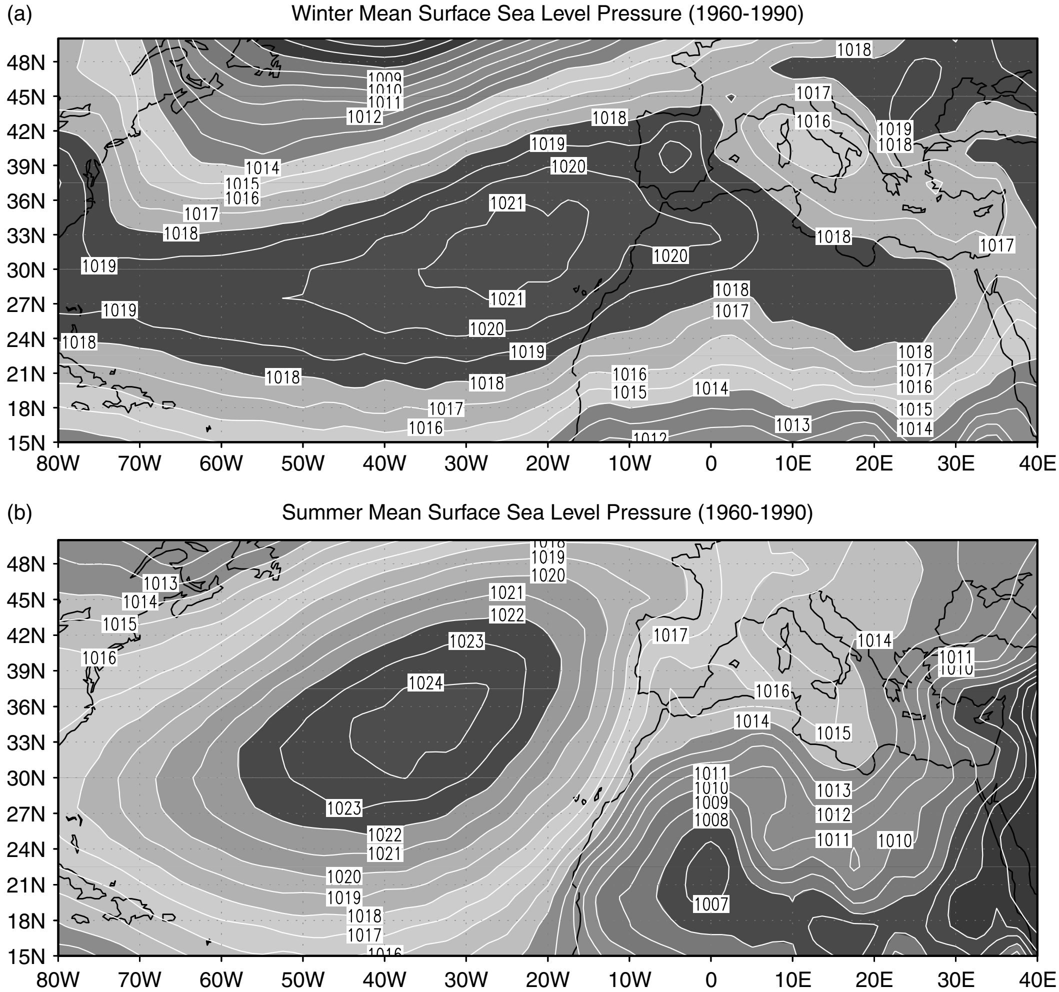

Figure 1. Mean SLP climatology (reference period is 1960–90) in (a) winter and (b) summer

be defined as

$$y_j = \frac{1}{j}\sum_{i=1}^{j} x_i \qquad j = 1, 2, \ldots, N \tag{1}$$

where x is the total annual SLP and N is the number of years of data used. Of course, $y_{j=N} = \overline{x}(N)$.

The evaluation of the trend analysis is based on the Hu *et al.* (1998) method. The 11 year running mean is used as a filtering method. It removes variations with periods shorter than 10 years in a time series and retains variations on interdecadal time scales, which are the focus of this study. To preserve as much information of the time series in the central position (year) of the 11 year running window as possible, different weights are assigned to each of the 11 years in the running mean. These weights, from the first through to the 11th year in the running window, are in order, 1/24, 1/24, 1/12, 1/8, 1/8, 1/6, 1/8, 1/8, 1/12, 1/24, and 1/24. The symmetry of the weight distribution guarantees no phase shift of the variations in the time series after the filter is applied. The response function of the running mean is similar to that of an ordinary filter (e.g. Shapiro, 1975). Also, it has a small effect on variations whose frequencies are lower than the cutoff frequency of the filter but it has a large effect on variations of frequency near its cutoff frequency, e.g. the 12 year variation.

The identification of an abrupt climatic change can be made by using the sequential version of the Mann–Kendall rank statistic (Sneyers, 1975, 1990). This test seems to be the most appropriate method for analysing climatic changes in climatological time series (Gossens and Berger, 1986). To use this test we need to consider only the relative values of all the terms in series X_i under analysis. For this reason, it is necessary before applying the test to replace the X_i by their ranks Y_i arranged in increasing order. For each element Y_i, the number n_i of elements Y_j preceding it ($i > j$) is calculated such that $Y_i > Y_j$.

The statistical test t_i is then given by

$$t_i = n_1 + n_2 + \ldots + n_i \tag{2}$$

and its distribution function under the null hypothesis is asymptotically normal, with mean and variance given by

$$E(t_i) = \frac{i(i-1)}{4} \quad \text{and} \quad \mathrm{var}(t_i) = \frac{i(i-1)(2i+5)}{72} \tag{3}$$

It is clear that, in the absence of any assumptions regarding the existence of a trend in a given direction, the test is correct only if its two-sided form is adopted, i.e. if the null hypothesis is rejected for large values of $|t_i|$. In these conditions, after calculating t_i, it is useful to determine the probability α_1, by the means of the standard normal distribution table, such that

$$\alpha_1 = p(|u| > |u(t_i)|) \tag{4}$$

where

$$u(t_i) = \frac{t_i - E(t_i)}{\sqrt{\mathrm{var}(t_i)}} \tag{5}$$

The null hypothesis is accepted or rejected at the level (e.g. 0.05) depending on whether we have $\alpha_1 > \alpha_0$ or $\alpha_1 < \alpha_0$.

When the values of $u(t_i)$ are significant, an increasing or decreasing trend can be observed depending on whether $u(t_i)$ is increasing or decreasing. However, when a series shows a significant trend, we may wish to locate the start of the phenomenon by means of a sequential analysis. In this case, it can be usefully extended to the reversed series. Therefore, we calculate the number n'_i of Y_j terms for each Y_i term, such that $Y_i > Y_j$ with $i < j$, which gives a check on the first calculation, since we have

$$n'_i = Y_i - 1 - n_i \tag{6}$$

so that

$$i' = (N+1) - i \tag{7}$$

where N is the total number of the series. Therefore, the values of $u'(t'_i)$ for the reversed series can be calculated similar to $u(t_i)$ as mentioned above.

In the absence of any trend in the series, the graphical representation of u and u' in terms of i generally gives curves which overlap several times. However, in the case of a significant trend, the intersection of these curves enables the start of the phenomenon to be located approximately.

The power spectrum of the SHCI time series was computed using auto-correlation spectral analysis (ASA; Mitchell *et al*., 1966) ASA is smoother and more accurate than fast Fourier transformation (FFT), but the amplitude relationship is poorly reflected (Padmanabhan, 1991). Our discussion will focus on practical aspects of the functional form for the ASA power spectrum B_i:

$$B_i = \frac{r_0}{m} + \frac{2}{m}\sum_{L=1}^{m-1} r_L \cos\left(\frac{360}{2m} i L\right) + \frac{r_m}{m}(-1)^i \tag{8}$$

where r_L is the auto-correlation coefficient of lag L. In the case of B_0 and B_m the coefficients resulting from the formula must be divided by two. Spectral estimates S_i are then computed by smoothing the raw estimates with a three-term weighted average. In the 'Hamming' method, the formula is

$$S_i = 0.25B_{i-1} + 0.5B_i + 0.25B_{i+1} \tag{9}$$

Variance spectra of all the time series have been estimated using ASA. The statistical confidence of the power spectra is tested using Markov red noise theory and χ-tests (Mitchell *et al.*, 1966) or Monte Carlo methods (Junk, 1983).

Finally, the classical correlation method was used to detect the relationship between SHCI and SSTs over the tropical Atlantic and eastern Pacific (ENSO). Programs for all computations and statistical analyses were prepared using the FORTRAN programming language.

4. VARIABILITY OF THE SUBTROPICAL HIGH

4.1. SHCI

Based on the NCEP–NCAR reanalysis SLP data sets, the SHCI is established for the period 1950–2002. It is a well-known fact that the variations of the location and intensity of the quasi-permanent circulation centres are partly seasonally driven and that these variations are modified by the differential heating of ocean and land areas. Figure 2 shows two time series of the SHCI, one from the present work (as described in Section 2) and the other from Hameed *et al.* (1995). Our SHCI depends mainly on the mean SLP data, which is obtained from NCEP–NCAR reanalysis. The two SHCI indices agree very well, especially in the winter season. The correlation coefficient between the two SHCIs is $r = 0.923$ in winter and $r = 0.825$ in summer. Thus, these SHCIs are both reliable and representative for describing the interannual variations of the subtropical high.

4.2. Cumulative annual means

To understand how the mean has changed over time, we used the cumulative annual means method (Pavia and Graef, 2002). Persistent phases of alternating higher or lower intensity of the circulation centre, which vary in length, are recognizable in the time series of the central pressure (Figure 3). In Figure 3, the most important feature here is the change, from below $\overline{y}$ to above $\overline{y}$, during the late 1960s and early 1970s in the winter and summer seasons. The pattern of SHCI in winter seems to be a cyclic wave, but in summer the SHCI seems to be an exponential function. In the winter season, the SHCI oscillates by about 3 hPa around the mean value, but it only oscillates by about 1.2 hPa in the summer season. The climatological mean values of SHCI have increased above the average of the whole period in the last 20 years in both seasons. These results agree with those of Mächel *et al.* (1998), where variations between positive and negative anomalies over longer time spells can be observed in the tropical pressure minimum.

4.3. Trend analysis

The evaluation of the trend analysis is based on the Hu *et al.* (1998) method. In addition to a large amount of interannual variability, there have been several periods when the subtropical high has persisted in a strong or weak state over many years. Apart from short-term variations, the subtropical high in the period between about 1900 and the 1960s is predominantly located south of its mean position (Mächel *et al.*, 1998). Mächel

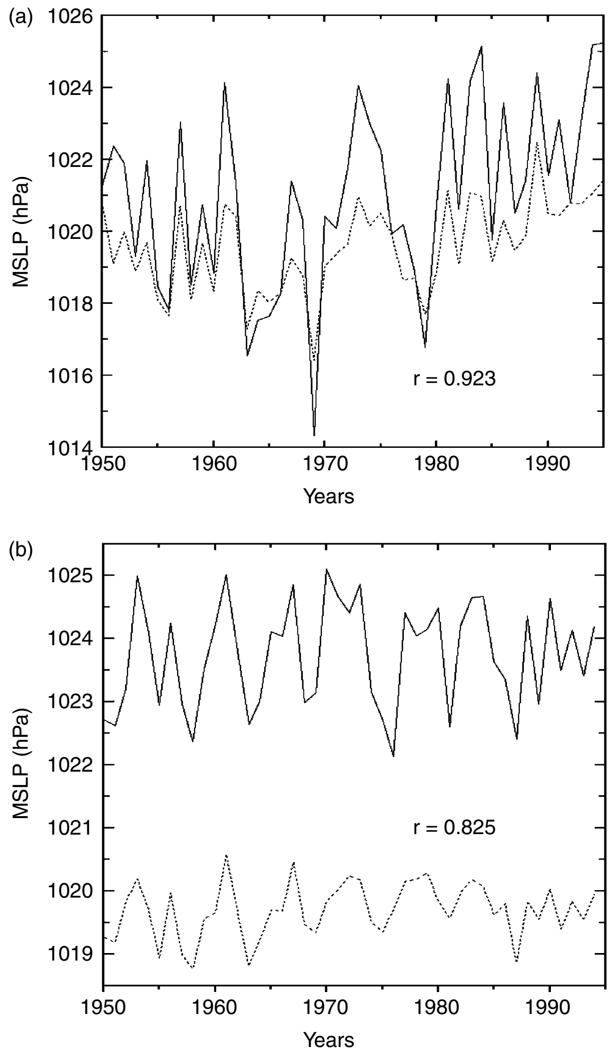

Figure 2. Comparison between two SHCI time series, one from the present work (solid line) and the other from Hameed *et al.* (1995) (dashed line) in (a) winter and (b) summer

et al. (1998) also noted that the central pressure of the subtropical high rises around the turn of the century, falls from the 1930s onwards and rises again in the late 1980s.

The SHCIs for the winter and summer seasons, as well as the fitting for the period 1950–2002, are shown in Figure 4(a) and (b). In winter (Figure 4(a)), the striking features are the low values of the fitting (below the mean) during the period from 1953 up to the late 1960s and the relatively high values of the fitting (above the mean) in the last 20 years. The 1970s was a period where SHCI fitting fluctuated around the mean value. The trend pattern in the winter season is different than the trend pattern in the summer season. In the summer season (Figure 4(b)), the SHCI fitting fluctuates around the mean in the period from mid 1950 up to the early 1990s. From 1995 up to the end of the period under study the fitting is seen to be increasing. The variations of the SHCI seem to be cyclic fluctuations. By using the Mann–Kendall rank statistic test for non-linear trend (Sneyers, 1990; Huth, 1999), the magnitude of trend for SHCI is equal to 0.25 and 0.10 in the winter and summer seasons respectively. Upon using the t-test, the trend in the winter season is found to be significant, but it is not significant in summer.

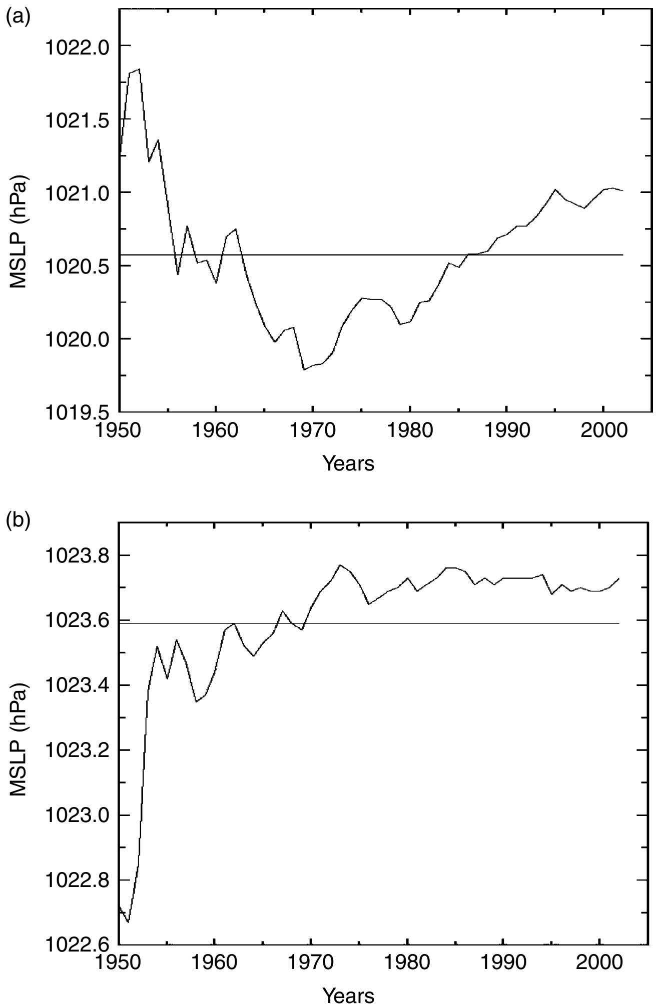

Figure 3. The cumulative annual mean (CAM) time series and the averaged CAM in (a) winter and (b) summer

4.4. Abrupt change

The concept of abrupt change is, by now, well established. For example, Rogers (1985) described the marked changes in SLP patterns of the North Atlantic during the 1920s. Flohn (1986) expanded the concept of abrupt climate change to include both singular events and catastrophes such as the extreme El Niño of 1982–83. Trenberth (1990) and Miller *et al.* (1994), among others, documented a rather abrupt shift of SST and atmospheric circulation features in the Northern Hemisphere in the mid-1970s.

Figure 5 shows the Mann–Kendall *t* test for the SHCI in the winter and summer seasons. This shows that for both seasons an abrupt climatic change took place. In the winter season (Figure 5(a)), negative and positive values (decreasing and increasing SLP) occur in periods starting from the late 1960s and from the early 1980s. A change toward increasing SLP (positive sign) is shown in 1972 and 1982 and change toward decreasing SLP (negative sign) is shown in 1968 and 1979. These seems to be a cycle with 10–11 years in this period. In the summer season (Figure 5(b)), a change toward increasing SLP (positive sign) is seen in 1953.

The episodic or abrupt changes of extra-tropical changes in circulation pattern have been documented in many observational studies (e.g. Namias 1988, 1990; Zeng *et al.*, 1994), but the overall characterization, let alone understanding, of abrupt changes remains unresolved. Fu *et al.* (1999) suggested that a major change

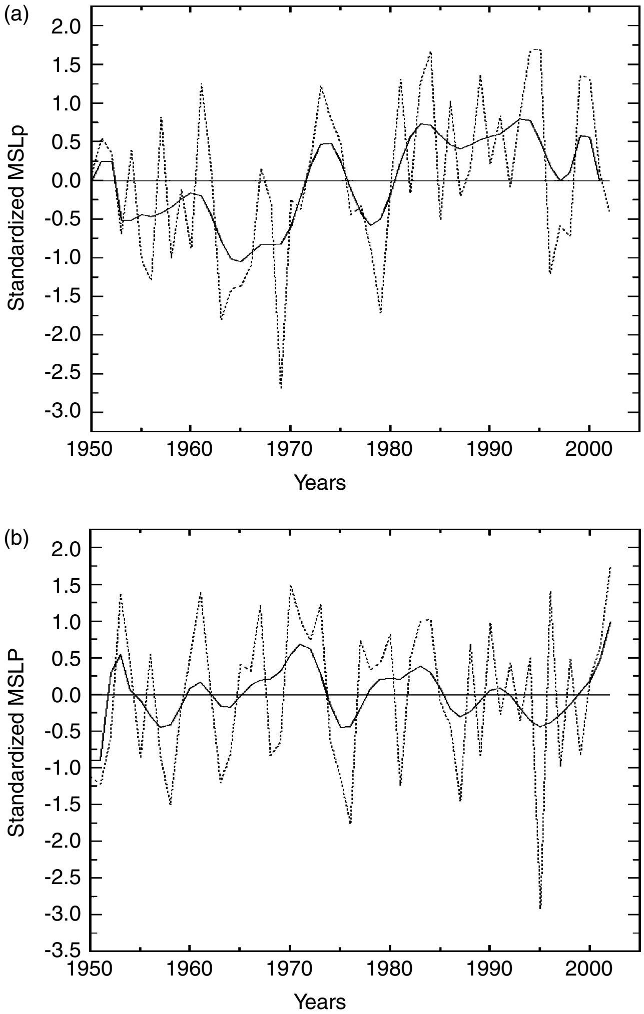

Figure 4. The fluctuations of the SHCI in (a) winter and (b) summer

of atmospheric circulation occurred at roughly the same time as an abrupt change throughout the northern oceanic subtropics. They also noted that the change in baroclinicity resulting from a reduced level of available potential energy leads to changes in position and strength of the subtropical highs.

4.5. Spectral analysis

The graphs of the cumulative annual means, linear trend, and abrupt change of central pressure suggest the existence of 'persistent' alternating spells between positive and negative anomalies. More detailed information can be given by a spectral variance analysis of these time series and the analysis of its periodic behaviour, which is based on Mitchell *et al*. (1966) and Jenkins and Watts (1968).

Figure 6 shows the ASA of the SHCI in winter and summer. The comparison of the smoothed spectra of these time series with the 'white noise' continuum reveals that the peaks above the 95% confidence level occur only at the lowest frequency (total length of the time series) in the central pressure spectra of the subtropical high in winter and summer. The results from applying the ASA techniques to SHCI reveal peaks of around 2.3 and 2.6 years in winter and a peak at around 7 year in summer that is highly significant at the 95% confidence level. A physical explanation of the periodicities mentioned above is that the shorter waves

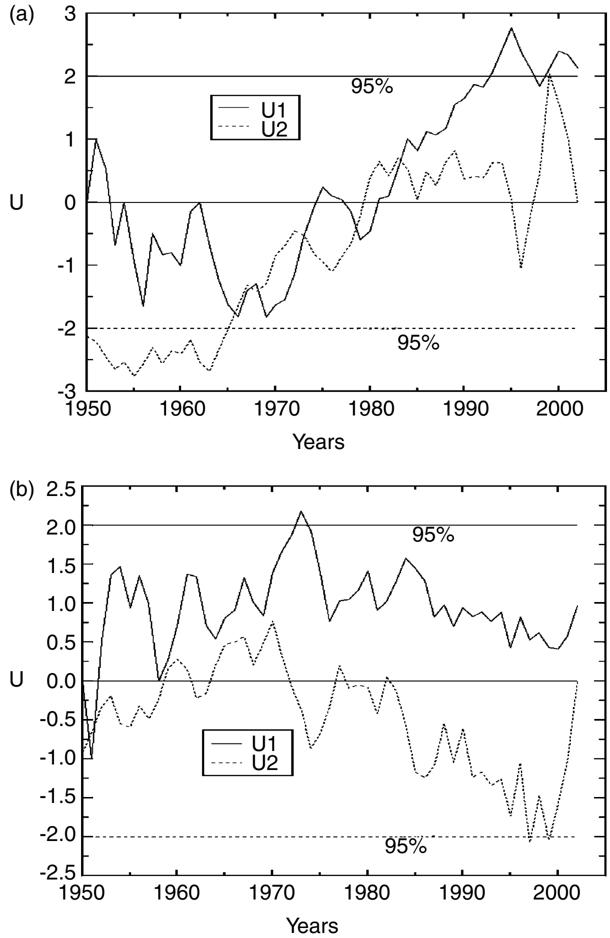

Figure 5. Abrupt change for SHCI time series in (a) winter and (b) summer as derived from sequential version of the Mann–Kendall test. (U1 forward sequential statistic and U2 backward sequential statistic)

(2.0–3.0 years) seem to be associated with the quasi-biennial oscillation (QBO). This connection has been mentioned by Angell *et al.* (1966), Scherhag (1967) and Lamb (1972). Lamb (1972) noted that the QBO is related to the southern oscillation, which is the strength of the subtropical high belt in both Northern and Southern hemispheres. Here, the correlation coefficient between SHCI and QBO (not shown) is weak but statistically significant. The ocean may cause oscillations of between 3 and 8 years (WMO, 1985; Malcher and Schonwiese, 1987). The 7 years periodicity could be associated with the dynamics of the atmosphere interacting with the North Atlantic Ocean (Garcia *et al.*, 2002). According to the spectral analysis, it can also be assumed that the centre of action of the subtropical high is characterized by non-periodic behaviour. This coincides with the results of Sahsamanoglou (1990) and Mächel *et al.* (1998) for the spectral analysis of the time series of the central pressure and position of the Icelandic low and subtropical high.

5. RELATIONSHIP BETWEEN SHCI AND SST OVER THE TROPICAL ATLANTIC VERSUS ENSO

Atlantic climate variability shows many important phenomena on different time scales. Unlike the tropical Pacific, the seasonal cycle dominates the ocean–atmosphere signal in the tropical Atlantic. A phenomenon similar to, but weaker than, the Pacific El Niño also occurs in the Atlantic (Latif and Grotzner, 2000). During a warm phase, trade winds in the equatorial western Atlantic are weak and SST is high in the equatorial eastern Atlantic. The converse occurs during a cold phase. This phenomenon is called the Atlantic zonal equatorial mode (or the Atlantic El Niño) (Wang, 2002), although SST anomalies in the tropical Atlantic are weaker than those associated with ENSO (Marshall *et al.*, 2001).

The analysis of the interactions between the SHCI and SSTs over the tropical Atlantic and east equatorial Pacific (ENSO) can be obtained from a correlation analysis. Tables I and II show the correlation coefficients

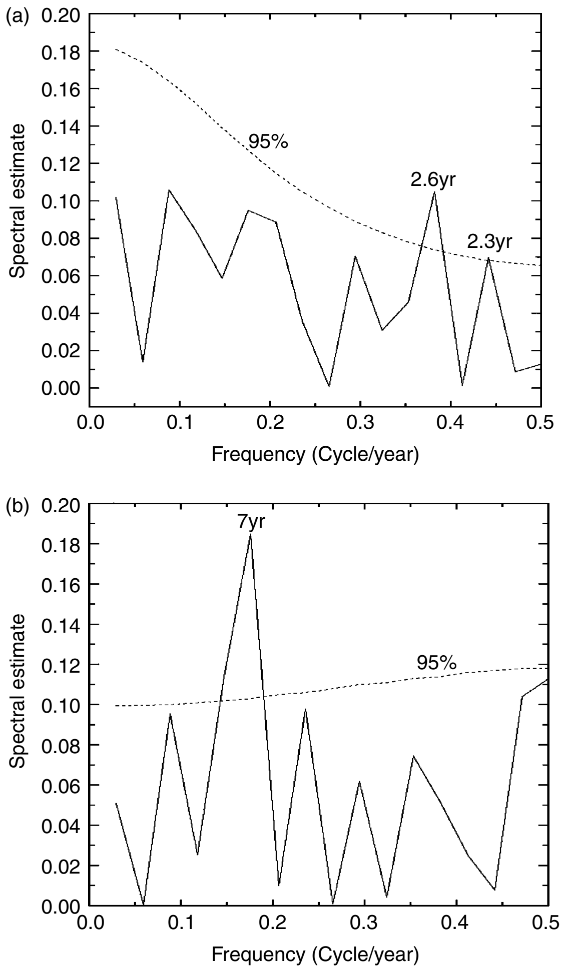

Figure 6. Power spectra of SHCI using ASA in (a) winter and (b) summer

Table I. Correlation coefficient between winter SHCI with seasonal and annual ENSO, NATL, SATL, and TATL indices

	ENSO	NATL	SATL	TATL
Winter	−0.04	−0.43**	0.25*	0.06
Spring	−0.05	−0.39**	0.26*	0.03
Summer	−0.13	−0.23*	0.26*	0.06
Autumn	−0.17	−0.22*	0.29*	0.02
Annual	−0.15	−0.39**	0.32*	0.03

* Significant at 95% confidence level.
** Significant at 99% confidence level.

Table II. Correlation coefficient between summer SHCI with seasonal and annual ENSO, NATL, SATL, and TATL indices

	ENSO	NATL	SATL	TATL
Winter	0.18	0.26*	0.10	0.21*
Spring	0.12	0.06	0.11	0.18
Summer	−0.18	−0.15	0.05	−0.03
Autumn	−0.13	−0.28*	0.04	−0.04
Annual	−0.05	−0.08	0.08	0.07

* Significant at 95% confidence level.
** Significant at 99% confidence level.

between the seasonal and annual NATL, SATL, TATL and ENSO with winter and summer SHCI. In winter, although the correlation coefficients are generally weak (Table I), but statistically significant, for SSTs of the tropical North Atlantic and tropical South Atlantic, they still indicate the relationships between the pressure centre and the SSTs. Strong negative relationships between the NATL index and SHCI and positive relationships with the SATL index are found. Consequently, cold SSTs over the tropical North Atlantic (5–20 °N, 60–30 °W) lead to an increase in SLP of the subtropical high, and vice versa. Also, warm (cold) SSTs of the tropical South Atlantic (0–20 °S, 30 °W–10 °E) lead to an increase (decrease) in SLP of the subtropical high. Changes in the Atlantic subtropical high induce variations in the northeast trade winds on its southern flank and then affect the tropical north Atlantic SST anomalies. The atmospheric circulation cell changes result in anomalous ascending motion in the tropical North Atlantic that decreases SLP and pushes the subtropical anticyclone northward; then decreases the northeast trade winds and latent heat flux; and thus increases the tropical North Atlantic SST anomalies (Wang, 2002). Wang (2000) also noted that the tropical Atlantic meridional gradient mode is associated with the variations in the Northern Hemisphere Hadley circulation in the tropical North Atlantic and tropical South Atlantic.

The relationships between ENSO and SHCI are weak and non-significant in both seasons. In general, the relationship between SHCI and SST over the tropical Atlantic and east equatorial Pacific is weak in the summer season, but it is strong in winter and autumn in the tropical North Atlantic and in winter in the tropical Atlantic.

6. CONCLUSIONS

Variability in the SHCI over the 53 year period from 1950 to 2002 was investigated. According to cumulative annual means analysis, the most important feature is the change of time mean from below to above average during the late 1960s and early 1970s in the winter and summer seasons. Fluctuations of 3 hPa and 1.2 hPa around the mean value were observed in the winter and summer season respectively. The climatological mean values of SHCI have increased above the average for the whole period in the last 20 years in both seasons.

Regarding the fluctuations in SHCI, the results show some striking features: there are low values during the period from 1953 up to the late 1960s and relatively high values in the last 20 years in winter. In the 1970s the SHCI fluctuated around the mean value. In the summer season the SHCI has fluctuated around the mean in the period from the mid 1950s up to the early 1990s, and it has increased from 1995 up to the end of the period under study. The variations in the SHCI seem to be cyclic fluctuations. The magnitude of trend in the winter season is equal to 0.25, which is significant at the 95% confidence level.

By examining the SHCI time series it was shown that a significant abrupt climatic change was observed in the winter and summer seasons. Increasing and decreasing episodes have occurred in the periods starting from the late 1960s and from the early 1980s. Increasing episodes occur in winter SHCI in 1972 and 1982 and in summer SHCI in 1953. On the other hand, decreasing episodes took place in winter SHCI in 1968 and 1979.

The results of the spectral analysis on the SHCI show peaks at around 2.3 and 2.6 years in winter, and there is a peak at around 7 years in the summer season that is highly significant at the 95% confidence level. So, a QBO (2.0–3.0 years) may affect the SHCI in the winter season, and the North Atlantic Ocean (7 years) may affect the SHCI in the summer season. The QBO is related to the southern oscillation, which is related to the strength of the subtropical high belt in both the Northern and Southern Hemispheres. According to the spectral analysis, it can also be assumed that the centre of action of the subtropical high is characterized by non-periodic behaviour. A comparison of smoothed spectra of these time series with the 'white noise' continuum reveals that the peaks above the 95% confidence level occur only at the lowest frequency (total length of the time series) in the central pressure spectra of the subtropical high in winter and summer.

The study of the interactions between the SHCI and SSTs over the tropical Atlantic and east equatorial Pacific (ENSO) shows a strong negative (positive) relationship between NATL (SATL) with SHCI in the winter season. These relationships between NATL and SATL with SHCI suggest that a change in the Atlantic subtropical high is induced by tropical North and South Atlantic SSTs. The relationships between ENSO and SHCI are weak and non-significant in both seasons. In the summer season, the relationship between SHCI and SSTs over the tropical Atlantic and east equatorial Pacific (ENSO) is weak.

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