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Coastal and oceanic SST variability along the western Iberian Peninsula

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ABSTRACT

The inter-annual variability of the sea surface temperature (SST) was analyzed along the western Iberian Peninsula in the region ranging from 9.5 °W to 21.5 °W and from 37.5 °N to 42.5 °N with a spatial resolution of $1^\circ \times 1^\circ$ from 1900 to 2008. Both coastal and oceanic SST showed an overall increase with warming and cooling cycles similar to those observed in the North Atlantic region and in previous regional studies. In addition, the evolution of coastal and ocean water has been observed to be different. In general, ocean water is more affected by the different warming–cooling cycles than coastal water. In spite of coast and ocean are highly influenced by global changes affecting the whole North Atlantic region, near shore SST has been observed to be correlated with local wind regime, which is itself a manifestation of the Eastern Atlantic (EA) teleconnection pattern.

1. Introduction

The temperature of the sea surface water (SST) is a fundamental parameter in the ocean-atmosphere heat exchange and hence in the climatic regulation. In addition, SST is influenced by climatic, meteorological, hydrodynamic and topographic parameters. During the last century, a great effort has been devoted to develop reliable SST series with global coverage, first by means of measurements from voluntary observation ships, drifters and moored buoys (Brohan et al., 2006; Smith et al., 2008) and then by means of satellite-derived data. Great efforts were also devoted to correct uncertainties in the SST data due to several factors as: changes in the ship routes after the opening of Panama and Suez Canals, sampling sparseness during the world wars, differences in water collection and more recently, uncertainties due to the presence of aerosols and clouds, which can cause a cool bias, and to the fact that satellite instruments record skin temperature instead of near-surface temperature (for a complete understanding of the different bias and the methods to correct them see: Kushnir, 1994; Folland and Parker, 1995; Kaplan et al., 1998; Smith and Reynolds, 2002, 2003, 2004, 2005; Worley et al., 2005; Kent and Berry, 2005; Kent and Challenor, 2006; Kent and Taylor, 2006; Brohan et al., 2006; Smith et al., 2008).

Numerous studies have tried to quantify trends in SST, which have shown to be extremely dependent on spatial and temporal scales being possible to observe opposite trends when considering different periods of time (Parker et al., 1994; Smith et al., 1994; Casey and Cornillon, 2001). Despite these differences, most of the studies carried out during the last decade concluded that a considerable global warming in SST has occurred over the last century no matter the considered data set (Folland et al., 1984; Folland et al., 1992; Parker et al., 1994; Nicholls et al., 1996; Casey and Cornillon, 2001). In addition, similar trends were observed in wind (Caires et al., 2003; Gillet and Thompson, 2003; Chelton et al., 2004), cloud coverage (Wiley et al., 2002; Roderick and Farquhar, 2002) and humidity (Flohn et al., 1990).

Global warming is far from being uniform in time and space. On the one hand, global warming is not spatially uniformly distributed all over the world's oceans since there are some regions where the warming is faster or slower than the global average (Levitus et al., 2000; Palttridge and Woodruff, 1981). In particular, the Atlantic Ocean contributes most to the increase of the heat content (Nerem et al., 1999; Levitus et al., 2000; Strong et al., 2000). On the other hand, according to the Intergovernmental Panel on Climate Change (2007), global SST time series show distinct warming-cooling periods during the last century. In particular, some authors (Garcia-Soto et al., 2002; deCastro et al., 2009; Gómez-Gesteira et al., 2011) have pointed out the existence of warming-cooling periods at several locations in the North Atlantic, which reflect the changes exhibited by the North Atlantic Ocean. Nevertheless, as far as know, the different response of coastal and ocean water to global warming-cooling cycles had not been considered in previous research.

Regional differences in the warming rate could be explained in terms of local and remote forcing factors (Cole et al., 2000; Lemos and Pires, 2004; Ginzburg et al., 2004; Santos et al., 2005; Gómez-Gesteira et al., 2008; deCastro et al., 2008). Among the remote factors, the Thermohaline Circulation (THC) highly influences SST features in the North Atlantic region carrying warm water from the tropics to northern latitudes. This circulation has been often

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stated as the main reason why Western Europe is so temperate compared to the same latitude in Eastern America. THC can be analyzed in terms of the Atlantic Multidecadal Oscillation index (AMO). AMO is a coherent pattern of multidecadal variability in SST centered on the North Atlantic Ocean with a cycle ranging from 35 to 80 years depending on the author (Delworth et al., 1993; Timmermann et al., 1998; Kerr, 2000; Dima and Lohmann, 2007). Following Trenberth and Shea (2006), the magnitude of the AMO signal is modest; the range is less than 0.4 °C. The AMO has been linked with the variability in Northeast Brazilian rainfall (Folland et al., 2001), North American climate (Sutton and Hodson, 2005) and U.S rainfall and river flows (Enfield et al., 2001). In addition, the AMO also affects the number of Atlantic hurricanes and the tropical storms (Goldenberng et al., 2001; Trenberth and Shea, 2006). Delworth and Mann (2000), suggested a link between the AMO and the variability of the THC as the mean THC transports sufficient heat northward (Ganachaud and Wunsch, 2000) to warm the Northern Hemisphere by several degrees (Vellinga and Wood, 2002). More recently, (Knight et al., 2005), by means of a 1400 year simulation with the HADCM3 climate model (Gordon et al., 2000), were able to simulate the observed AMO pattern and amplitude from measurements dating back to the nineteenth century. The results imply that the AMO is a genuine quasi-periodic cycle of internal climate variability persisting for many centuries, and is related to variability in the oceanic THC.

Upwelling forcing on SST is possibly the most important oceanographic feature in the so called EBUEs (Eastern Boundary Upwelling Ecosystems) since it involves the replacement of warmer surface water by cooler subsurface water. Even, according to some authors, changes in the thermal gradient between land and ocean can be responsible of changes observed in upwelling intensity (Bakun, 1990; Mendelssohn and Schwing, 2002; McGregor et al., 2007).

The western coast of the Iberian Peninsula (37 °N to 43 °N) may be regarded as the northern boundary of influence of a broader upwelling system (Eastern North Atlantic Upwelling System), that acts all along the northwest coast of Africa and the Atlantic coast of the Iberian Peninsula (Nykjaer and Van Camp, 1994; Santos et al., 2005; Álvarez et al., 2008a). These previous studies have shown that upwelling is mainly a seasonal event that occurs with higher probability from April to September.

The aim of the present study is to describe the differences in SST evolution during the last century at coastal and oceanic locations along western Iberian Peninsula. The SST variability will be analyzed in terms of upwelling and THC intensity.

2. Data and processing

SST was obtained from the UK Meteorological office, Hadley Center HadISST1.1-Global sea-Ice coverage and SST (http://badc.nerc.ac.uk/data/hadisst) (Rayner et al., 2003). Data are available from 1870 to nowadays, with monthly periodicity on a $1^\circ \times 1^\circ$ grid with global coverage. In the present study, 78 data points in front of the western Iberian Peninsula (WIP) coast were considered from 1900 to 2008. These points range from 9 °W to 21.5 °W and from 37.5 °N to 42.5 °N, (Fig. 1). Monthly SST data were seasonally and annually averaged.

The SST difference between coast and ocean was calculated as:

$$\Delta SST = SST_{ocean} - SST_{coast} \quad (1)$$

Twelve points were considered for this purpose. Points located at 9.5 °W (17.5 °W) are representative of coastal (oceanic) conditions (Fig. 1, circles). Differences were calculated between each

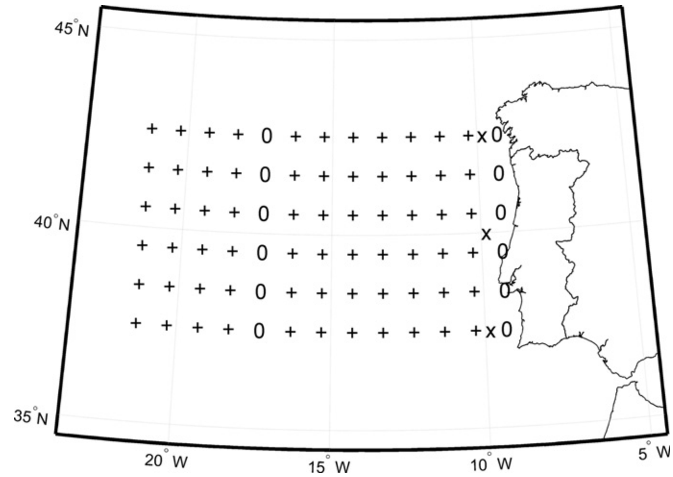

Fig. 1. Study area. SST points (+) are placed on a $1^\circ \times 1^\circ$ grid from 9.5 °W to 21.5 °W and from 42.5 °N to 37.5 °N. Circles (0) represent coastal (9.5 °W) an oceanic (17.5 °W) reference SST points. Crosses represent the location of wind data (located at 10.0 °W, 37.5 °N, 40.0 °N and 42.5 °N).

pair of points located at the same latitude and then meridionaly averaged.

The AMO index, which characterizes the large-scale pattern of multidecadal variability of SST, was calculated as the SST anomaly averaged for the North Atlantic region. This index has been traditionally calculated as the average of SST anomaly for the North Atlantic north of the equator (Enfield et al., 2001). For practical purposes the grid used in the present study covers from 7.5 °W to 75.5 °W and from 0 to 59.5 °N. The SST northern limit was kept at 59.5 °N to avoid problems with sea ice changes.

Wind data were obtained from the National Center of Atmospheric Research/National Center for Environmental Prediction (NCEP/NCAR) (http://www.esrl.noaa.gov/psd/data/reanalysis/reanalysis.shtml). Reanalysis Data Archive are available from 1948 onwards, with a global coverage and a spatial resolution of $2.5 \times 2.5^\circ$. In this study, points located in front of the WIP at 10 °W and 42.5 °N, 40 °N and 37.5 °N, respectively, were considered from 1948 to 2008 (Fig. 1, crosses). Time series obtained at the different points showed to be well correlated ($R > 0.8$) allowing a meridional average. The upwelling index (UI) can be defined as minus the zonal component of Ekman transport. Note that the shore line is macroscopically perpendicular to the Equator along the WIP:

$$UI = -Q_x = -\frac{\tau_y}{\rho_w f} \quad (2)$$

where

$$\tau_y = \rho_a C_d (W_x^2 + W_y^2)^{1/2} W_y \quad (3)$$

being τ_y the meridional wind stress, W the wind speed near surface, ρ_w the sea water density (1025 kg m^{-3}), C_d a dimensionless drag coefficient, (1.4×10^{-3}), ρ_a the air density (1.22 kg m^{-3}) and f is the Coriolis parameter defined as twice the vertical component of the Earth's angular velocity, Ω, about the local vertical given by $f = 2\Omega \sin(\theta)$ at latitude θ. Finally, x subscript corresponds to the zonal component and the y subscript to the meridional one. Negative (positive) τ_y values result in positive (negative) UI values, which correspond to upwelling favorable (unfavorable) conditions.

According to previous research (deCastro et al., 2008), the Eastern Atlantic mode (EA) shows a significant negative correlation with upwelling along the entire western coast of the Iberian Peninsula. The EA pattern consists of a north–south dipole that spans the entire North Atlantic Ocean with centers near 55 °N, 20