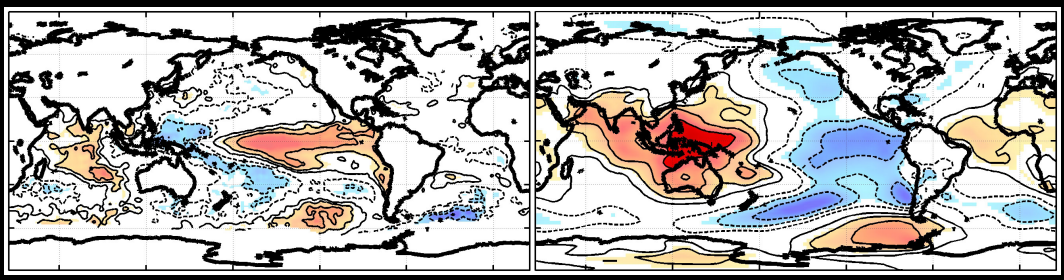
Climate predictability: theory and impacts at seasonal and climate change timescales

Predictibilitat climàtica: teoria i impactes a l'escala estacional i a la del canvi climàtic



Joan Ballester Claramunt

Barcelona, 2011

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Abstract

The atmosphere, as any dynamical system highly sensitive to initial conditions, has a finite limit of predictability. This threshold affects our ability to anticipate the evolution of the climate system at all timescales, from the forecast of interannual anomalies some months ahead to the study of the sensitivity of climate statistics to slow-varying changes associated with global warming. This dissertation particularly focus on these two topics. On the one hand, a new precursory signal is introduced to improve the forecast and our understanding of El Niño-Southern Oscillation phenomenon, which represents the most prominent modulator of atmospheric variability worldwide and the major driver of climate teleconnections. On the other hand, the sensitivity of the distribution of European temperatures to the increasing concentration of atmospheric greenhouse gases is described, and used to infer projections of temperature-related mortality.

The role of the oceanic RossBell dipole as a new premonitory signal for the onset of recent eastward-propagating El Niño events is established. This extratropical tracer is followed by El Niño events around 9 months later, but at the same time, it occurs a year after the development of a warm oceanic area in the western tropical Pacific region. This initial anomaly generates an anomalous atmospheric wavetrain extending eastward and poleward in the southern hemisphere, which is associated with the RossBell feature. Changes in the atmospheric circulation lead to warm ocean anomalies in the central tropical Pacific, being later enhanced by suppressed equatorial easterlies. These processes are linked to an eastward shift in the tropical convection, and thus a weakening of the Walker circulation, setting up the positive Bjerknes feedback that exponentially grows on top of the incipient warming and leads to the mature phase of El Niño.

Changes in the distribution of European temperatures in a context of global warming are described from two different points of view. On the one hand, it is shown that the increasing intensity of the most damaging summer heat waves is mostly due to higher seasonal mean temperatures in summer, and not to specific changes in deseasonalized anomalies during extreme events. This result appears to be compatible with previous studies stating that future heat waves will be more intense, more frequent and longer lasting as a result of temperature rise. On the other hand, a simplified methodology is presented in order to reproduce the simulated changes in the temperature distribution from a relatively small set of parameters. Thus, the change in frequency, length and intensity of warm, neutral and cold temperatures is here derived from the evolution in only three central statistics of the temperature distribution, the mean, standard deviation and skewness.

This simplified methodology is verified by testing the link between daily temperature and mortality data in nearly 200 European regions representing more than 400 million people. These relationships are later used to infer projections of mortality

under greenhouse gas emission scenario simulations from an ensemble of state-of-the-art high-resolution climate models. Analyses point to a change in the seasonality of mortality, with maximum monthly incidence shifting from winter to summer. Results also show that the rise in heat-related mortality will indeed compensate the reduction of deaths from cold at the end of the century, unless a substantial degree of acclimatization to warm temperatures takes place, stressing the need for adaptation and mitigation strategies.

Què és el més important de la vida?

M'agradaria expressar la meva sincera gratitud al doctor Xavier Rodó, per haver cregut en mi, i per la seva infinita perseverança i paciència. Res del que hi ha escrit en aquesta tesi doctoral hauria estat possible sense ell. Acabo aquesta etapa havent après moltes coses a nivell professional i humà gràcies a ell.

També desitjaria agrair a tots aquells companys de feina, d'aquí i d'allà, que han participat de forma més o menys directa en el disseny, elaboració, anàlisi o discussió dels resultats aquí exposats. La riquesa d'aquesta tesi també prové d'ells.

No em voldria oblidar dels companys de l'IC3, que no només humanitzen el meu dia a dia, sinó que també m'humanitzen a mi.

Però sobretot vull agrair a aquelles poques persones que conformen el nucli dur del que sóc, sense el suport, afecte i amor dels quals em seria molt difícil emprendre els grans projectes de la meva vida. A la meva família, per l'amor incondicional que mai m'ha faltat. No solem valorar allò que de forma immutable sempre hi és, i per això em sento especialment orgullós de saber apreciar una cosa de la qual he pogut i segueixo podent gaudir en abundància.

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List of Abbreviations

ACW	Antarctic Circumpolar Wave
CEOF	Complex Empirical Orthogonal Function
EN	El Niño
ENSO	El Niño-Southern Oscillation
GHG	GreenHouse Gases
IPCC	Intergovernmental Panel on Climate Change
LN	La Niña
N34	Niño3.4
PDF	Probability Distribution Function
RB	RossBell
SLP	Sea Level Pressure
SST	Sea Surface Temperature
WPAC	Western tropical PACific

Contents

Chapter 1

State of the art

1.1 Climate predictability

1.1.1 Definition and characterization of predictability

The influence of human beings on the dynamical evolution of the atmosphere is negligible at meteorological timescales, and therefore the set of equations governing the motion of air are formally *deterministic*, which means that any future state is completely and univocally determined by its present state, or more generally, by a set of present and past states. Against this notion of determinism, a process is said to be *stochastic* if present and past states merely determine the Probability Distribution Function (PDF) of future states, and *completely stochastic* if they determine nothing. During the first half of the twentieth century, scientists believed that the deterministic equations describing the motion of air could be indefinitely integrated forward in time, so that improvements in the description of the atmosphere would ultimately lead to a continuous and unlimited increase in the forecast skill.

The atmosphere is however a deterministic chaotic system, and as such, our ability to predict future states is limited by the instabilities of the system. In this sense, a realization of a deterministic process is said to be *stable* if the change in its future evolution remains small after introducing a slight perturbation in the initial state (i.e. it is not highly sensitive to initial conditions), and *unstable* otherwise. Nonetheless, the way in which we implement the equations governing the motion of air is stochastic. The property of determinism does not hold when, for example, a number is rounded in a computational operation of a climate model, since nearly identical states become indistinguishable (find further details and examples in Section 1.1.2 below).

Formally speaking, *predictability* is the extent to which events can be known in advance. This simple and generic definition however conceals in itself a heterogeneous range of events, which prevents scientists from a unified framework. The most prominent objective frontier in this sense is imposed by the chaotic nature of the atmosphere, which defines a theoretical threshold in the prediction lead time of up two weeks (cf. weather and climate forecasting; Section 1.1.2). As a result of for example this heterogeneity, no unique and universal criterion is used to determine to which extent an event of a dynamical system is predictable.

In the classical approach (i.e. using analogues), the differential equations describing weather

and climate,

$$\frac{d}{dt}x_i = f_i(x_1, \ldots, x_n), \qquad i = 1, \ldots, n$$

are solved for a sequence of time steps with a pair of slightly different initial states, and then the difference between these realizations is compared with regard to two states of the system chosen at random (Somerville 1987). Lorenz (1984) proposed to evaluate the error growth rate by studying the set of linear differential equations

$$\frac{d}{dt}\delta x_i = \sum_{j=1}^{n} A_{ij}\delta x_j, \qquad i = 1, \ldots, n,$$

where (A_{ij}) is the Jacobian matrix of $f = (f_1, \ldots, f_n)$. Note that these equations describe the evolution of errors for the initial time period when they remain small, since otherwise the error would grow infinitely in the unstable case. These linearized equations define the Lyapunov characteristic exponents, which in turn determine whether the system is stable or not and the directions of expansion or contraction in the phase state (Fraedrich 1987).

Alternatives to this classical error-based approach have been proposed, for example, in information theory, where the predictability of an event is derived from the notion of uncertainty. Given that the state of the system is not completely known, it is typically represented by a PDF describing the relative probability of each possible state. Thus, in information theory, the larger is the information that is gained in a prediction, the lower is the uncertainty that remains in the PDF. In this way, the *information* of an event X with probability $p(x)$ is defined as

$$h(X = x) = log\frac{1}{p(x)}.$$

Therefore, the larger is the probability of an event x, the lower is the gain in information after it is observed. Accordingly, the *entropy* is defined as the weighted average of information

$$H(X) = \int p(x)h(x)dx = -\int p(x)log(p(x))dx,$$

which is a natural measure of the uncertainty associated with the event X (Schneider and Griffies 1999).

In this context, an event is said to be *predictable* when the distribution of states changes to some extent after an ensemble of data is taken into account (DelSole 2004). Note that the unconditional and conditional PDF are typically referred to as *prior* ($p(x)$) and *posterior* ($q(x)$) distributions, respectively. In information theory, the predictability of an event can be simply measured as the reduction in the entropy

$$-\Delta H(X) = H_p(X) - H_q(X),$$

i.e. the decrease in the uncertainty of the distribution. Accordingly, Schneider and Griffies (1999) defined the *predictive power* as

$$\alpha = 1 - e^{\Delta H(X)},$$

which is equal to 0 when the uncertainty is not reduced and tends to 1 with increasing predictive information. Alternatively, Kleeman (2002) defined the *relative entropy* as the weighted increase in information,

$$R = \int q(x)[h_p(x) - h_q(x)]dx = \int q(x)log\frac{q(x)}{p(x)}dx,$$

which vanishes if and only if the prior and posterior distributions are equal.

The notion of climate predictability as the difference between a prior and a posterior distribution is somewhat differently interpreted when the PDF describing the relative density of states of the system is conditioned by a larger climate distribution (DelSole 2004). For example, the fundamental aim in climate change detection and forecasting is the description of the sensitivity of climate statistics to a modification in an external factor, i.e. the concentration of Greenhouse Gases (GHG; cf. problem of first and second kind in Section 1.1.3). These distributions can thus either be conditioned by the state of the system itself through observations initializating a prediction scheme (e.g. weather forecasts), or by the state of a boundary condition (e.g. ocean and land surface) or an external forcing (e.g. GHG; Section 1.1.4). Note that the choice of a prior distribution, as well as the way in which it is conditioned, is closely associated with, for example, the timescale of the forecast (Section 1.1.5).

Weather and climate predictions can be essentially formulated as a two-step transformation (Fortin et al. 2004). In a first step, referred to as assimilation, the distribution is conditioned by the whole set of available observations, generating a conditional PDF of initial states of the system. Atmosphere data assimilation is responsible for the large improvement of weather prediction schemes to date. In this way, coupled ocean-atmosphere data assimilation appears to be one of the most promising avenues for the improvement of the climate forecasting. The degree to which a future system state is predictable is thus intrinsically determined by our description of the system state in the moment of the prediction. In a second step, the distribution of future states is obtained after transforming the PDF of initial states according to a given statistical or dynamical climate model. The skill of the prediction is thus also determined by extrinsic factors, such as the chosen forecast scheme and its implementation; e.g. the level of complexity in the coupling of the different climate components, the propagation of errors in their interaction, the discretization in time and space, or our limited understanding and inaccurate parameterization of small-scale subgrid processes (Section 1.1.6).

Among all the climatic signals providing predictability, El Niño-Southern Oscillation (ENSO) is the most prominent modulator of atmospheric variability, and the major driver of climate teleconnections, and therefore the forecast of ENSO is a crucial factor for the predictability of climate worldwide (Section 1.2). The predictability of ENSO is generally lower in the boreal spring, and once this barrier is overcome, the subsequent phases of the event are clearly much easier to predict. The successful prediction of ENSO at longer lead times would ultimately increase the global predictability of the climate system.

1.1.2 Weather and climate predictability

The study of the predictability of weather and climate has evolved in parallel with the description and implementation of equations governing the motion of air. These equations (Bjerknes 1904), as well as the first hand written prediction scheme (Richardson 1922), were developed during the first decades of the 20th century. Advances in the theoretical description of the dynamics and general circulation of the atmosphere, and the development of new and faster computers in the fifties, encouraged the scientific community to generate the firsts weather predictions (Charney et al. 1950), and to develop the firsts numerical models of the general circulation (Phillips 1956). The Laplacian concept of determinism pushed forward a general optimistic belief among scientists from the Bergen and Chicago Schools, in the sense that the equations could be indefinitely integrated forward in time. Thus, the climate community believed that the accuracy of the forecasts would continuously increase in parallel with computer availability, improvements in the description and implementation of the equations, and the development of new, wider and

more reliable observation data sources.

This naive conception of unlimited capability in the prediction of the atmosphere rapidly faded away after a revolutionary discovery in the sixties: the atmosphere has an finite limit of predictability. Based on previous theoretical studies pointing that small errors should tend to amplify (Thompson 1957), Lorenz described in a set of papers a strict and unavoidable temporal upper bound in the prediction of weather that is inherent to the nature of the atmosphere itself (Lorenz 1963). He showed that the equations governing the evolution of the atmosphere are strongly dependent on the initial state, and thus the error of the prediction rapidly grows in time even only assuming an infinitesimally small initial perturbation. This property of weather forecasts, originally tested for a 28-variable model (Lorenz 1965), laid the foundations of the chaotic nature of the atmosphere (see butterfly effect), which states that weather is theoretically unpredictable beyond two weeks (Lorenz 1982). More interestingly, the degree to which the system is predictable is itself a function of the initial state, and therefore some initial states lead to better predictions than others (Palmer and Williams 2008).

Predictability at meteorological timescales is thus strongly limited by the chaotic nature of the atmosphere itself. Although this constraint defines a theoretical barrier for long-term weather predictions, other factors impose additional limits to the forecast of the atmosphere. To begin with, our understanding of the equations describing the dynamics of the atmosphere is limited. In turn, the implementation of these equations requires a discretization in space and time, preventing the capture of those smaller-scale physical processes that must be consequently parameterized (Cess et al. 1990). Thus, even if the differential equations governing the atmosphere are deterministic, their numerical discretization, and the subsequent parametrization of subgrid scale processes, are modeled by means of statistical methods that might not satisfactorily resolve their structure. This kind of solution also motivates the stochastic nature of uncertainties in current prediction schemes.

In addition, observations used for the initialization of the models are far from accurately describing the whole tridimensional troposphere. This imperfect knowledge of the system is expressed for example in the accuracy, representativeness, homogeneity and resolution of the observational data networks. Nevertheless, data sources have exponentially increased in number and become more reliable during the last decades, especially after the launch of meteorological satellites. In addition, assimilation techniques have been continuously developed, leading to clear and significant improvements in weather prediction schemes (Navon 2009). These achievements, as well as advances in parameterizations, model resolution, computer power and numerical procedures, have all contributed to the outstanding progress in weather forecasting skill during the recent decades.

Despite these major achievements, chaos inevitably imposes a severe upper threshold that restricts the domain of application of weather forecasting, and thus the general features of the atmosphere cannot be predicted more than two weeks in advance from an initial state and the set of equations governing the system. At most, under specific circumstances, some large-scale quasi-stationary anomalous circulation features can be forecasted on the timescale of one month (Stern and Miyakoda 1995). But, is it possible another kind of forecast at longer lead times? Since Lorenz masterful work, climatologists have shown that some processes and features of the atmosphere can be anticipated at longer leads. The use of dynamical models in climate forecasting thus defines a relatively new discipline that has grown relatively fast, essentially because most of the knowledge that is required for its implementation and modeling was generated as a need for the development of the firsts weather prediction schemes.

One way to extend the range beyond the predictability threshold enforced by atmospheric chaos is spatiotemporal averaging (Miyakoda et al. 1986). This technique is especially effective in

removing the more unpredictable processes occurring in the atmosphere. Thus, spatiotemporal averaging extracts the low-varying large-scale features from the less predictable high-frequency small-scale processes (Lorenz 1984). This technique is useful when the amplitude of anomalies filtered in the averages exceeds the unpredictable component of weather fluctuations (Epstein 1988). Uncertainties in these predictions are not homogeneous, so that the longer the lead time, the larger the temporal and spatial scales for the averaging. For example, weather forecasting can quite easily anticipate the daily mean temperature in a small area one day in advance through the description of changes in the atmospheric general circulation. Instead, climate forecasting can only predict, if so, the probability of higher- or lower-than-normal monthly or seasonal mean temperature anomalies for next summer in relatively large areas. This peculiar way of depicting the information contained in the prediction, so usually misunderstood by the general public, is characteristic of the stochastic nature of statistical techniques used in climate forecasting. Thus, models are commonly run several times by means of slightly perturbed initial states, and in this way the ensemble average is assumed to reduce the strong dependency to original conditions.

The interaction between the atmosphere and external boundaries or forcings is another basic element explaining how climate forecasting can overcome the deterministic limit imposed by the chaotic nature of weather. The major natural boundaries modulating the evolution of the atmosphere are the ocean and the land surface. Their slow-varying anomalies regulate the exchange of fluxes in the interface, so that the spatiotemporal averages of anomalies in the atmosphere become predictable at longer leads. This transfer of memory from the ocean and the land surface to the atmosphere is however limited to periods of large and persistent boundary anomalies. An external forcing can also modulate the evolution of climate at long time scales. For example, the increase in the concentration of atmospheric GHG is leading to a detectable anthropogenically induced global warming since the mid twentieth century (IPCC WGI 2007). All these factors are characterized by their relatively slow-varying interaction with the atmosphere (ocean and land) or the whole climate system (GHG), either as a transmission of energy and moisture to the atmosphere (ocean), a partitioning of incoming radiation into sensible and latent heat fluxes (soil wetness), or a planetary balance between incoming and outgoing radiation (GHG). Please, refer to Section 1.1.4 below for further details.

1.1.3 Limits to climate predictability

Lorenz (1975) distinguished between two basic cases in the study of the predictability of the climate system. The problem of first kind, or initial value problem, refers to the rapidly increasing propagation of initial errors. This problem is of especial interest in coupled ocean-atmosphere prediction schemes, because the state of the ocean surface is a major factor providing climate predictability, but available in-situ data sources are scarce. As an example, the prediction of ENSO several months in advance is strongly conditioned by the set of initial ocean-atmosphere conditions that are prescribed in the tropical Pacific. Two of the main factors preventing us from enlarging the predictability up to the theoretical limit are the degree of reliability of current climate models and the amount of available observational data. Thus, realistic initial ocean conditions are produced if observations are assimilated into the ocean under atmospheric forcing conditions. Nevertheless, when these initial conditions are prescribed in a coupled prediction scheme, the run is more likely to crash as a result of the sudden shift from an uncoupled to a coupled configuration (Chen and Cane 2008). The development of better ocean data assimilation schemes thus represents a new avenue, not widely explored yet, for the improvement of current prediction schemes.

The problem of second kind, or boundary value problem, is derived from uncertainties in

the prescription of the future state in the boundary conditions. The effects of this kind of problem are present, for example, in uncertainties arising from our lack of knowledge about the future evolution of surface ocean and continental land conditions or GHG emissions. Lateral boundary conditions are additionally required in those prediction schemes based on regional climate models. In most of these cases, the future state of these factors is completely unknown, and it is sometimes estimated by means of unsuitable techniques (e.g. persistence or climatology). Note, however, that this problem is in permanent evolution, since climate models are continuously incorporating additional components of the climate system into their configuration (e.g. ice, vegetation, aerosols; EC-EARTH project). Thus, some of these components can also become sensitive to the prescription of initial conditions and to the dynamical interaction with the other components of the modeling scheme.

The theoretical limit of deterministic predictability is smaller in the tropics than in the extratropics, essentially because the error growth rate is larger in the tropics, and the saturation value of errors in the extratropics (Shukla 1981). These theoretical considerations refer to natural atmospheric variability unconstrained by boundary forcings, and therefore they are typically valid only for weather forecasting (Lorenz 1984). When variability in the boundary components is however taken into account, climate predictability becomes clearly larger in the tropical belt, where atmospheric variability is strongly constrained by slow-varying boundary conditions in the ocean surface (Shukla and Kinter III 2006). In climate forecasting, the unpredictable component of natural variability is known as internal variability, in contrast to the boundary-forced component providing predictability, which is referred to as external variability. In accordance with this decomposition of variability, the potential predictability of a forecast model scheme is defined as the signal-to-noise ratio between the external and internal variability, and it is typically used to describe an upper bound for the predictability of a climate event (Phelps et al. 2004).

The predictability of the climate system is, for instance, higher when and El Niño (EN) or La Niña (LN) event is on the way. This kind of forcing anomalies define periods of general higher predictability, when the forecast schemes perform better than on average (i.e. windows of opportunity; Troccoli 2010). Nonetheless, the prediction of ENSO itself also exhibits a non-uniform skill, depending on other factors such as the period of the year in which the forecast is issued. Thus, the predictability of ENSO is generally lower in boreal spring, given the relatively small signal-to-noise ratio at this time of the year (Chen and Cane 2008). Indeed, the noise in the ocean-atmosphere coupled system is relatively constant throughout the year, but ENSO variability reaches its minimum amplitude during this season of the year (Quan et al. 2004). Once the spring barrier is overcome, the subsequent phases of the event are clearly much easier to predict, given the slow-varying and autoregressive characteristics of the phenomenon.

The potential predictability of the climate system depends on the coupled ocean-atmosphere prediction scheme that is used to evaluate it. When the effect of the initial value problem is studied, several simulations are typically initialized with slightly perturbed conditions (Boer 2000). The spread of the ensemble is thus estimated, and the dispersion among ensemble members gives an estimation of the flow-dependent predictability. Other authors have also evaluated the effect of the boundary value problem. At decadal timescales, for example, some of these studies compared pairs of simulations with and without active ocean dynamics, so that those regions with active ocean variability are assumed to be potential sources of predictability (Latif et al. 2009). All these methodologies provide upper thresholds of climate predictability, and they can therefore be used for the evaluation of the performance of current prediction schemes. For instance, some studies have estimated that ENSO is potentially predictable several seasons in advance, at lead times clearly beyond the spring barrier (Collins 2002). At decadal timescales, the potential predictability in the north Atlantic basin is larger than 5 years, being modulated by the southward transport of anomalously cool polar waters and the low-frequency changes in

the thermohaline circulation (Griffies and Bryan 1997).

1.1.4 Boundary conditions and external forcing

There is a large body of evidence from numerical experiments showing that the ocean and continental surface have a significant effect on the variability of the overlying atmosphere. This influence is exerted at relatively slow timescales (e.g. the large heat capacity explains the persistence of temperature anomalies in the ocean surface), compared to the characteristic timing of internal atmospheric variability, and therefore the delayed effect can last for much longer. This memory is expressed as a reddening of the spectrum of atmospheric variability (Marshall et al. 2001), and it thus enlarges the predictability of the climatic system under specific favorable circumstances; e.g. magnitude, season, location and persistence of the anomaly, kind of synoptic configuration in the overlying atmosphere or sensitivity of convergence and convection circulation processes to the forcing.

Mechanisms explaining the transmission of warm ocean anomalies to the atmosphere are summarized as follows (find for example further details in Shukla and Kinter III 2006). Sea Surface Temperature (SST) anomalies can modify the temperature and humidity conditions near the surface through changes in the ocean-atmosphere heat fluxes, which in turn alter the circulation in the lower levels of the troposphere. In this sense, convergence is especially active in the tropics, where the rotational component is relatively small. This low-level convergence drives moisture fluxes that ultimately enhance the convection through saturated adiabatic lapse rates. The release of latent heat due to condensation strengthens this mechanism, defining a positive feedback that modifies the circulation around the initial oceanic anomaly area. The release of energy from the ocean to the atmosphere occurs in about ten days, and the change in the low-level circulation is activated a month after the initial forcing. The effect of the ocean anomaly on the general circulation in the atmosphere depends however on many other factors, such as the seasonal position of the large-scale cells.

The mechanism explaining the transmission of continental land anomalies to the atmosphere is somewhat different, and its effect clearly more local and highly variable. These anomalies (mainly in soil moisture, but also in snow cover or vegetation) are less effective in modifying the general circulation, because the amount of moisture that can be released to the atmosphere is very limited. Soil moisture determines, for example, the fraction of energy that is released to the atmosphere as latent or sensible heat. An initial anomaly in the continental surface thus alters, if so, the low-level temperature, humidity and wind circulation through modifications in the radiation and exchange of heat. These changes determine the weather that is locally observed, which in turn enhance or weaken the initial land anomalies. Although some recent initiatives are under way (e.g. SMOS project), it is difficult to estimate the amount of available soil moisture in the continental surface, because station data is not representative of anomalies in adjacent areas.

We illustrate here the interaction between the atmosphere, the ocean and the continental surface in a set of climate simulations. A state-of-the-art atmospheric general circulation model is here coupled to a 50-m slab mixed-layer ocean model, except in some specific regions where SST conditions are prescribed (i.e. pacemaker design; Cash et al. 2008). In a first ensemble of simulations (control runs), climatological conditions are imposed around the tropical north Atlantic (here defined as [75W,0] $\times$ [5S,45N]). On the other hand, a warm spot is superimposed on these climatological conditions in a second set of simulations (prescribed runs). This ocean anomaly is here assumed to appear in autumn, the maximum in magnitude is reached in spring (see Figure 1.1b), and then it starts to decay until autumn, when it finally disappears. The spatial

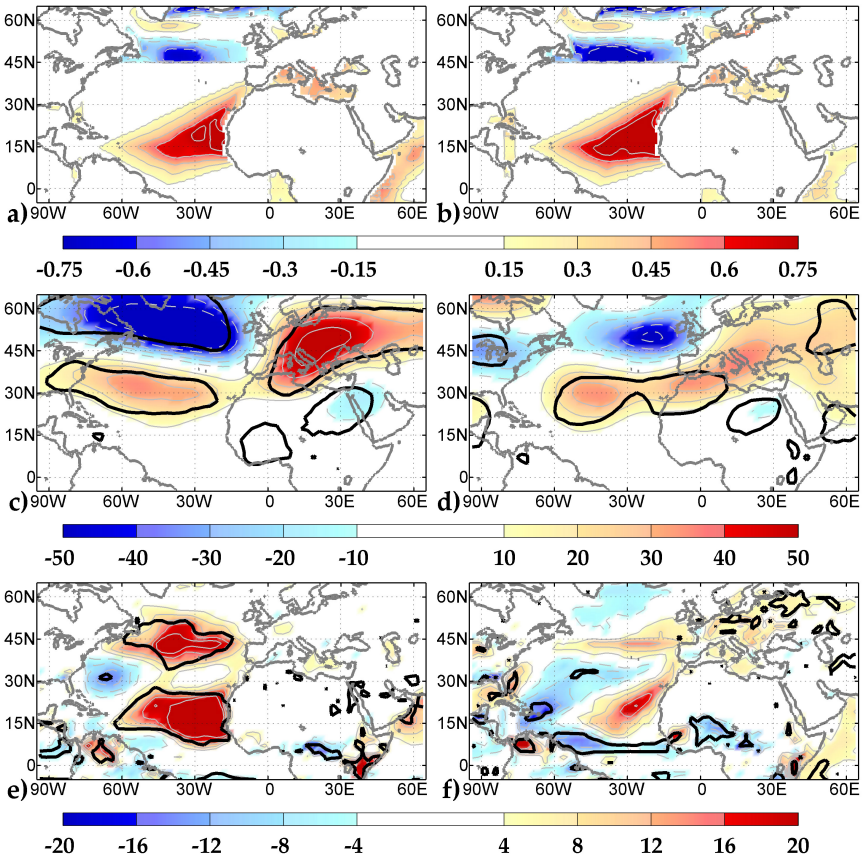

Figure 1.1: Ensemble mean anomalies (prescribed minus control runs) of SST (in K; panels a,b), 300hPa geopotential height (m; c,d), and evaporation (Kg/m^2; e,f). Anomalies were averaged for January-February-March (panels a,c,e) and March-April-May (b,d,f) before and during the maximum of the superimposed ocean warm spot anomaly in spring. Black thick contours in panels c-f indicate areas of significant anomalies ($p < 0.1$).

pattern of this anomaly corresponds to one of the main characteristic modes of variability in this area (see for example Ruiz-Barradas et al. 2000). Climatological SST conditions in the tropical Pacific were additionally prescribed in all the simulations, in order to suppress the modulating effect of ENSO. Note that each ensemble of simulations consists of 10 members initialized by slightly perturbed initial conditions. Further information of the design of the experiments can be found, for example, in Cash et al. (2009).

The tropical north Atlantic is characterized by its direct connection with the atmospheric variability occurring in the subtropics and extratropics. Indeed, during the rising phases of the warm spot anomaly in winter and spring (Figures 1.1a,b), the response of the atmosphere to the oceanic forcing resembles the East Atlantic mode (Barnston and Livezey 1987), i.e. a low pressure anomaly in the north Atlantic surrounded by an elongated area of high pressure anomalies extending from the subtropical Atlantic to central Europe (Figures 1.1c,d). This configuration strengthens the northeasterly trade winds in the northern tropical and subtropical area, which in turn enhance the release of energy through increased latent heat fluxes (Figures

1.1e,f). In addition, the persistent atmospheric dynamical stability also leads to an initial rise in evaporation in Europe (Figure 1.1f).

This situation of dynamical stability in Europe persists in summer (Figure 1.2a), with warmer-than-normal temperatures (Figure 1.2b) and decreased precipitation (Figure 1.2c). Note that the magnitude of the seasonal heat wave is here exacerbated by preexisting dynamical stability and latent heat fluxes (Figures 1.1d,f), which reduce the amount of humidity available in the soil (Figure 1.2d) and the low-troposphere (Figure 1.2f), and increase the fraction of energy that is released to the atmosphere as sensible heat (Figure 1.2e). All these factors indeed played an important role in the record-breaking 2003 heat wave in western Europe. Numerical simulations demonstrated that temperature anomalies in summer 2003 could have been reduced by around 40% in some regions if no preexisting deficit in soil wetness was observed (Fischer et al. 2007). The potential predictability of a negative soil moisture anomalies in spring is however essentially limited to lead times of up to a season, since the subsequent drop in summer precipitation is compensated by a much larger decrease in evaporation, which finally reduces the shortage in soil moisture.

The natural variability of the climate system is thus the final result of the interaction between its components, principally the atmosphere, the oceans, the land surface and the snow and ice covers. These and other constituents (e.g. vegetation, chemistry) define the internal dynamics of the system, but this intrinsic variability is also subject to the modulation of external factors. One of these external forcings is the amount of incoming solar radiation, which empowers the circulation in the ocean and the atmosphere as a mechanism of latitudinal heat exchange. The radiation balance in the climate system is defined by three main factors: (i) the incoming solar radiation, (ii) the fraction of it that is reflected, and (iii) the long-wave radiation emitted by the planet that is trapped by the atmosphere (IPCC WGI 2007). The amount of incoming radiation depends on factors such as the solar activity or the Milankovitch cycles. Some of this radiation is reflected for example by clouds, aerosols, several groups of atmospheric gases, the vegetation cover or the land surface. All these factors determine the temperature of the planet, and therefore the long-wave radiation that is then emitted. Atmospheric GHG concentrations (e.g. water vapor, carbon dioxide, atmospheric methane or nitrous oxide) thus determine the amount of long-wave radiation that escapes to the outer space.

All these factors have been the only ones to naturally fluctuate until the industrial revolution, when human societies started to emit GHG to the atmosphere. These emissions led to a detectable anthropogenically induced global warming signature since the mid twentieth century. The evolution of future emissions is however unknown (i.e. it is not a naturally-varying deterministic factor of the system), as it is the future trend in global temperatures. The Intergovernmental Panel on Climate Change (IPCC) thus defined several families of hypothetical GHG emission scenarios in order to overcome this lack of knowledge. Under these hypotheses, GHG-induced climate trends are predictable at very long lead times, since the aim of climate change forecasting is the description of the sensitivity of climate statistics to this external forcing. In accordance with these scenarios, climate model simulations provide an estimation of GHG-induced trends, which are superimposed on the naturally-oscillating internal variability of the climate system. Given that internal variability in essentially unpredictable at long lead times, information from these simulations is essentially extracted for relatively long periods (typically of at least 30 years), which are assumed to be nearly independent of decadal fluctuations.

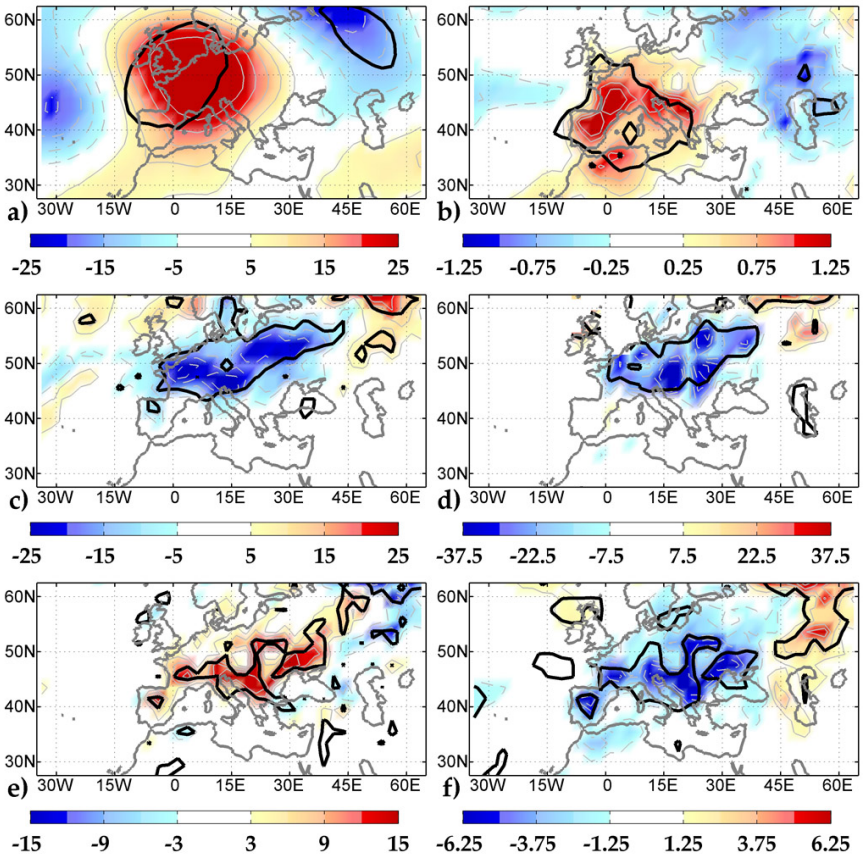

Figure 1.2: As in Figure 1.1, but for 300hPa geopotential height (in m; panel a), surface air temperature (K; b), precipitation (mm; c), soil moisture (Kg/m^2; d), sensible heat flux (W/m^2; e), and relative humidity (%; f). In this case, anomalies were averaged for June-July-August after the maximum of the superimposed ocean warm spot anomaly in spring. Black thick contours indicate areas of significant anomalies ($p < 0.1$).

1.1.5 Timescales of climate predictability

Seasonal predictability is both determined by the prescription of initial conditions and the subsequent propagation of errors (e.g. spring barrier in the prediction of ENSO), and by the effect of oceanic and continental surface anomalies regulating the exchange of fluxes in the interface (e.g. teleconnections of ENSO). Despite internal dynamics can also play a significant role at decadal and multidecadal timescales, the prediction of these slow-varying atmosphere anomalies is also subject to other major external influences, such as the circulation in the deep ocean or the concentration of GHG (Latif et al. 2009). Instead, only external factors play a significant role at centennial and longer timescales, which are mainly associated with radiative and planetary dynamics (Weber et al. 2004).

Despite the very different nature of problems arising in the prediction from daily to millennial timescales, some of the efforts of the climate community are now being redirected to the assimilation of the different modeling schemes used in weather and climate forecasting (WCRP 2005). Since weather fluctuations can also propagate on longer-term climate variability, this

approach is aimed at providing an unified framework for the description of those dynamical processes oscillating in a wider range of timescales. This framework, referred to as seamless prediction, implies seamlessness across space and time scales, scientific disciplines, institutions and geographical boundaries, and tries to integrate those aspects of the different prediction schemes used for the generation of specific forecasts for a particular range of timescales (Shukla 2009).

Although some extratropical large-scale atmospheric waves are even predictable for lead times of a month or so, seasonal predictability is basically extracted from slow-varying persistent boundary anomalies in the ocean, and essentially in the tropics. ENSO is by far the most prominent oceanic source of predictability at these timescales (details are not provided here, refer to Section 1.2 below). The contribution of the other two tropical basins is however considerably smaller and less understood. For example, the role of the tropical Indian ocean in the predictability of the coupled system is not always easy to determine, since part of its variability is connected with processes occurring in the tropical Pacific (e.g. relationship between ENSO and the Indian ocean dipole or the Indian monsoon; Krishnamurthy and Kirtman 2003; Kumar et al. 2006). Despite these problems, numerical experiments showed that ocean variability in this basin is indeed an important driver of climate teleconnections in many adjacent tropical areas (Goddard et al. 2001).

All the knowledge acquired (e.g. methodological approaches already implemented in seasonal forecasting) is currently being used to develop the necessary tools for the prediction of climate at decadal and longer-term timescales. The lack of suitable data records describing the potential areas providing long-term predictability is however the main problem arising at these timescales. In addition, observational data is less reliable for early historical time periods. Mechanisms providing predictability at these timescales are in general not yet understood, and results are essentially based on modeling studies that need to be validated against observations, and compared to other models. The major areas of potential predictability at decadal timescales are found in the ocean, particularly in the mid- and high-latitudes. As a rule of thumb, ocean-atmosphere interactions are mostly driven by the atmosphere at interannual timescales, but the direction of this modulation is reverted in the case of decadal and longer-term variability (Bjerknes 1964).

The north Atlantic region is by far the region of major potential predictability on decadal and multidecadal timescales, since the thermohaline circulation is predictable at least a decade in advance, with even larger lead times under specific favorable conditions (Collins and Sinha 2003). For example, the low-frequency fluctuations of the North Atlantic Oscillation have led to variations in the thermohaline circulation by about a decade during the last century (Latif et al. 2006). Note that changes in the strength of the thermohaline circulation are in turn associated with predictable climate variations in adjacent continental areas, such as in western Europe (Rahmstorf 2003). Other regions, such as the southern oceans, also exhibit a significant fraction of natural variability in the decadal and multidecadal band, but mechanisms explaining these potential sources of predictability are not clear yet. Note that predictability in the tropical and north Pacific is also significant, but clearly smaller, since an equivalent overturning circulation is not present there.

Two sources of decadal and longer-term variability will essentially define the future evolution of climate anomalies during the present century: (i) the naturally-varying internal variability of the climate system and (ii) anthropogenic GHG emissions. On the one hand, natural variability exhibits limited predictability, of the order of up to a very few decades at most, given the chaotic nature of the climate system and our limited knowledge about potential sources of predictability playing a role at these timescales. On the other hand, anthropogenic changes in long-term climate variables are predictable under the assumption of GHG emission scenarios. In this case, predictability is understood as the estimation of the sensitivity of climate statistics to an external

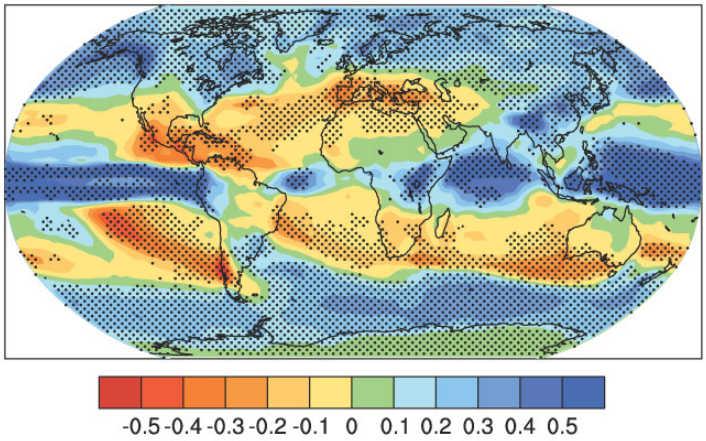

Figure 1.3: IPCC AR4 multimodel mean changes in precipitation (in $mm\ day^{-1}$). To indicate consistency in the sign of change, regions are stippled where at least 80% of models agree on the sign of the mean change. Changes are annual means for the SRES A1B scenario for the period 2080 to 2099 relative to 1980 to 1999. Adapted from IPCC WGI 2007.

forcing. Although these latter changes are expected to be detectable in most of the cases, they are also projected to be relatively small in the short term compared to natural low-frequency variability. Once emissions start to speed up, GHG emissions are projected to induce first order trends, especially for variables directly associated with temperature (e.g. sea level or ice melting). The relative contribution of these two superimposed sources of variability is thus projected to change in the coming decades. Indeed, some of the anthropogenic low-frequency changes will not be easily traced in the short term, e.g. the weakening of the meridional overturning circulation (Latif et al. 2009). An appropriate prescription of both initial and boundary conditions might thus improve the detection of simulated climate change signals.

One of the main consequences of anthropogenic climate change is the widening of the tropical belt, that is, the poleward shift of the descending branch of the Hadley cell in both hemispheres. This expansion is caused by an increase in the subtropical static stability, which pushes poleward the zone of baroclinic instability (Lu et al. 2007). In climatology, indirect tracers are used to characterize the limits of the tropical belt, such as the distance between subtropical jet streams, discontinuities in the tropopause height or measures of day-to-day and seasonal variability. According to these indirect tracers, recent studies have shown that the width of the tropical belt has increased by at least 2 degrees in the last 25 years (Hudson et al. 2006; Hu and Fu 2007). Surprisingly, state-of-the-art climate models are unable to reproduce the whole magnitude of this change, suggesting that the observed trend is partially affected by superimposed natural multidecadal fluctuations, or current climate models are unable to simulate the troposphere-stratosphere interactions that modulate these processes.

Either in one case or the other, these changes will have important effects on climate variability in the mid-latitudes. The widening of the tropical belt will lead to a poleward shift of the mid-latitude storm track (e.g. fewer cyclones in the Mediterranean area; Somot 2005), and a general decrease of precipitation near the subtropical areas (Figure 1.3). Particularly, climate change simulations point to a redistribution of precipitation anomalies in Europe, so that annual precipitation is very likely to increase (decrease) in northern Europe (Mediterranean basin).

Annual temperatures in Europe are likely to rise more than the global mean (Figure 1.4a), and

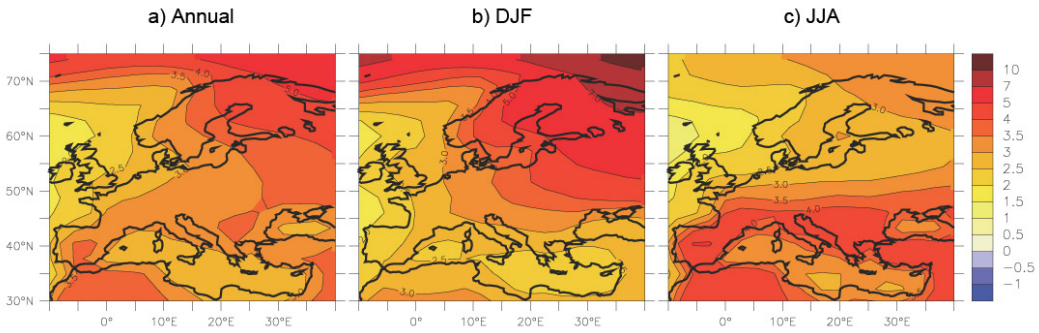

Figure 1.4: IPCC AR4 multimodel mean changes in annual (a), winter (b) and summer (c) mean temperature (in °C) over Europe. Changes correspond to the SRES A1B scenario for the period 2080 to 2099 relative to 1980 to 1999. Adapted from IPCC WGI 2007.

this increase is mainly explained by changes in the atmospheric circulation and thermodynamical processes (IPCC WGI 2007). Both factors indeed played an important role in the record-breaking 2003 heat wave in western Europe, an extremely unlikely event sharing similar characteristics with future summers (Schär et al. 2004), and causing up to 70000 excess deaths in only 3 months (Robine et al. 2008). According to future scenarios for central Europe and the Mediterranean region, the soil will dry in summer as a result of enhanced evaporation in spring (e.g. due to earlier snowmelt) and decreased precipitation, further exacerbating summer and heat wave temperatures (Douville et al. 2002). In northern Europe, it is not clear yet whether these enhancing mechanisms will play a role, since the increase in precipitation might compensate the enhanced evaporation in spring (Wang 2005a). All these factors define a scenario with more severe summer conditions in the Mediterranean area, compared to the relatively lower warming in northern Europe (Figure 1.4c). On the other hand, the winter snowline will be pushed northwards and eastwards, and therefore the increase in winter temperatures will be clearly larger in northeastern Europe as a result of albedo and snow emissivity feedbacks (Figure 1.4b; Déqué et al. 2007).

In addition to the general shift in annual mean temperatures, the different magnitude and spatial distribution of temperature rise in summer and winter will modify, in turn, the shape of the PDF. The standard deviation of the PDF is for example projected to increase (decrease) in western and southern (northeastern) Europe (Figure 1.5), depicting a larger (smaller) difference between future temperatures in the warm and cold seasons. Therefore, the effect of climate change will not only show up as a trend in annual mean temperatures, but also as a redistribution of the relative frequency of events within the PDF. Implications of these changes are enormous, ranging from food and water supply to the spread of diseases (IPCC WGII 2007). Particularly, the relationship between temperature and mortality, which is characterized by minimum incidence around a range of comfortable temperatures and larger rates in winter and midsummer (Christidis et al. 2010), might also change (spontaneously or not) as temperatures start to speed up (McMichael et al. 2006). Increasing temperatures and changing climate impacts will represent a major challenge for human societies during the 21st century, since it might require some degree of continuous adaptation from governments and individuals, as well as the design of new mitigation strategies (IPCC WGIII 2007).

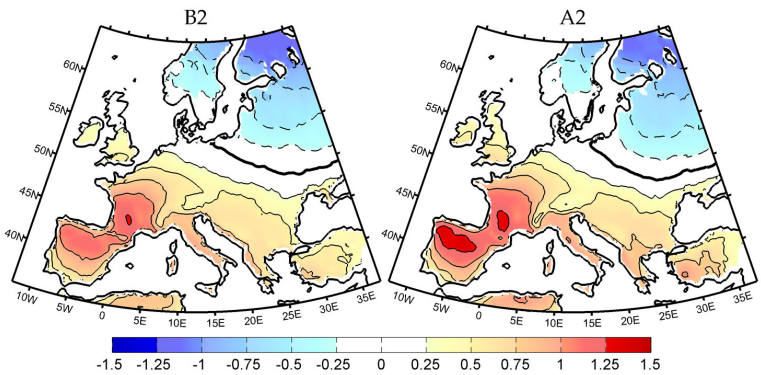

Figure 1.5: PRUDENCE multimodel mean changes in the standard deviation of temperature (°C) over Europe. Changes correspond to the SRES B2 and A2 scenarios for the period 2071 to 2100 relative to 1961 to 1990.

1.1.6 Climate modeling and forecasting

Two basic methodological approaches are used to produce climate predictions. On the one hand, statistical models infer the forecast from historical relationships between two or more variables. On the other hand, dynamical forecast schemes are based on the solution of the equations describing the discretization of physical components of the climate system (e.g. atmosphere, ocean, snow and ice covers, land surface). It is not clear yet which of these approaches provides more accurate predictions in general, since they depend on many circumstantial factors such as the base state of the physical components, the length of historical data, the region that is forecasted, the model resolution or the amount of computational resources. For example, in the case of ENSO, both approaches exhibit comparable predictive skills, which seem to have reached a plateau at moderate level (Chen and Cane 2008).

Statistical models are based on empirical relationships between predictor and predictand variables inferred from historical observations. The development of this approach was motivated by the increasing network of available meteorological data, and its main assumption (and perhaps weakness) is that the statistical relationships are nearly invariant in time. Future predictions are thus inferred from our past experience, and therefore a formal description of the physical mechanisms of these empirical relationships is neither provided nor required. Statistical models are however not transportable from one region or season to another, and therefore their application is generally constrained to specific circumstances. Assuming that no outstanding statistical procedure is discovered, the accuracy of statistical schemes can only be improved by increasing the amount of computational resources or the length of data records. Unfortunately, despite the remarkable increase in data availability during the last decades, these advances are relatively small compared to the increase in computational resources, and thus this kind of approach is strongly constrained by our limited knowledge of the historical evolution of climate. Most of the statistical prediction schemes include ocean-atmosphere ENSO variability as a relevant predictor variable, because this mode of variability is the most prominent modulator of atmospheric variability and the major driver of climate teleconnections (find further details in Section 1.2). As such, statistical predictions are clearly more skillful during and after EN and LN events.

Dynamical models describe the equations governing the dynamics of the physical components of the climate system. The complexity of this kind of models ranges from atmosphere-only general circulation models to more complex schemes coupling the atmosphere to the ocean,

the land surface or the ice cover. The most sophisticated schemes also include the interaction with other components of the climate system, e.g. vegetation, aerosols, atmospheric chemistry, carbon cycle or river flow (IPCC WGI 2007). As a rule of thumb, the larger the complexity of the models, the more reliable are the predictions. Nonetheless, modeling challenges also arise from this increasing complexity of the model schemes, such as the propagation of errors in the interaction between components, problems in their initialization, or our limited understanding and inaccurate parameterization of small-scale processes. Indeed, even assuming suitable initial conditions, dynamical models are also subject to systematic biases derived from our limited understanding of the natural processes. The simulation of the atmosphere is for example sensitive to the parameterization of the convective and marine stratus clouds, and the simulation of the ocean is limited by the parameterization of mixing processes above the thermocline, and particularly near the surface (Goddard et al. 2001). Advances in the description of these processes, as well as in computer availability (e.g. spatiotemporal resolution), would therefore lead to further improvements in the accuracy of dynamical modeling and forecasting.

1.2 El Niño-Southern Oscillation

The expression *El Niño*, which refers to the Christ child in Spanish, was initially used to describe the local warming in ocean temperatures near the Peruvian coasts during the celebration of the Christmas holidays. The worldwide importance of this phenomenon was however overlooked for almost a century. Meanwhile, Walker (1924) identified a seesaw in atmospheric pressure between Indonesia and the eastern Pacific which he referred to as the Southern Oscillation. It was not however until the sixties that climatologists described the intimate link between both phenomena, when they realized that the warm anomalies near south America are a local expression of a larger basin-wide phenomenon. Bjerknes (1969) thus proposed a positive feedback between ocean and atmosphere anomalies that explains the exceptional energetic nature of the mode. The description of the feedback laid the foundations of the coupling between the oceanic and atmospheric counterparts, known as ENSO since then. After a decade of modest advances, the broad interest for the ENSO phenomenon and its prediction finally took off after the 1982/83 EN peak, an intense event with strong worldwide teleconnections that could not be recognized until it was well developed.

1.2.1 Bjerknes feedback: growth and ending

Bjerknes, in his early works, described the coupled nature of ENSO as a fundamental mechanism for its energetic nature. Within the coupling between the ocean and the atmosphere, a positive feedback progressively amplifies incipient perturbations until the peak of EN or of its counterpart LN (An et al. 2005). Under normal conditions, easterly trade winds in the tropical Pacific drive warm surface ocean currents to the west of the basin, with subsurface cold ocean waters being pumped up in a region of active upwelling in the east. The thermocline, i.e. the surface separating the warm surface from cold waters at depth, thus deepens in the west and slopes up in the east. The zonal gradient in SST in turn enhances (suppresses) the convection and the upper-level divergence in the west (east). The strengthening of the Walker circulation thus completes the loop, with upper-troposphere westerly winds and surface easterly trade winds.

During the onset of an EN (LN) event, the coupled feedback between the ocean and the atmosphere is progressively weakened (enhanced) after an initial relaxation (reinforcement) of the easterly trade winds and/or a decrease (increase) of the zonal SST gradient. Due to the strongly coupled nature of the phenomenon, an initial perturbation in the ocean or the atmosphere is

transmitted to its counterpart, ultimately initializing a loop of anomalies between both components. Nevertheless, the onset of EN can be manifested through initial perturbations in different regions of the basin, which might vary from one event to another. Events can be thus classified in several types according to specific properties such as the onset time, the periodicity or the zonal SST structure (Kao and Yu 2009). Particularly, the location of initial SST anomalies and the subsequent direction of zonal propagation defined a multidecadal change in tropical Pacific variability around 1977.

Before this regime shift, EN events were predominantly characterized by the canonical type (Figures 1.6a,c,e,g), in which positive SST and subsurface anomalies have their origin near the coasts of south America, they propagate northwestwards along the equatorial surface up to the central Pacific, and then they move backwards to the east eventually reverting the initial pattern. In this case, the generation of the event is linked to anomalous westerlies in the eastern-to-central tropical Pacific (Kao and Yu 2009). On the other hand, EN events since the regime shift have been mostly associated with initial anomalies in the western equatorial Pacific (Figures 1.6b,d,f,h). These ocean anomalies propagate eastwards via equatorial downwelling Kelvin waves, and they are in turn connected with suppressed equatorial easterly winds traveling from the western to the central tropical Pacific (Vecchi and Harrison 2000).

Either in one case or the other, the mature phase of EN is typically characterized by the relationship between warm SST and negative Sea Level Pressure (SLP) anomalies in the eastern and central tropical Pacific, and opposite values in the western tropical Pacific (Figures 1.7a,c). In turn, this configuration defines an area of anomalous westerly surface winds along the equatorial Pacific (Figures 1.7b,d). The region of highest convection and precipitation also appears circumscribed within both pressure areas, but in this case it appears slightly more to the east in association with the area of warm anomalies, in part because deep convection over the warm water acts to transport air (Figures 1.7e,f). In the upper-troposphere, a couple of symmetric off-equatorial anticyclonic anomalies above the area of highest convection enhances the subtropical jet streams and the anomalous easterly circulation in the equator (Figures 1.7g,h; Horel and Wallace 1981; Karoly 1989).

The description of Bjerknes was however incomplete, as it requires a negative feedback stopping the loop and allowing the transition from the warm phase of ENSO to neutral or cold conditions, or vice versa. Several theories have been proposed during the last decades to justify this required out-of-phase mechanism. For instance, the delayed oscillator offers a description of some of the main physical processes involved in this transition, through continuous wind-induced ocean impulses throughout the evolution of an ENSO cycle. In this theory, westerly wind stress anomalies induce oceanic Kelvin and Rossby waves that are trapped within a narrow band of few degrees around the equator. Equatorial downwelling Kelvin (off-equatorial upwelling Rossby) waves propagate eastwards (westwards) along the tropical Pacific to deepen (shallow) the warm ocean layer, and thus the thermocline. The propagation speed of these waves depends on their latitudinal structure, so that they can cross the whole Pacific in around 3 (9) months (Troccoli 2010). Hence, the downwelling Kelvin waves initially deepen the thermocline in the eastern Pacific, while the relatively slower upwelling Rossby waves are reflected in the western boundary and shallows the thermocline in the eastern Pacific long time afterwards, reverting the initial warming.

Additional physical mechanisms complementing the delayed oscillator have been proposed. For example, the western Pacific oscillator (Weisberg and Wang 1997) does not overlook the role of the western Pacific. In this theory, convection initiates a couple of symmetric off-equatorial cyclones in the central Pacific, driving equatorial westerlies and deepening the thermocline in the eastern Pacific. These cyclones also shallow the thermocline and cool the ocean surface in the

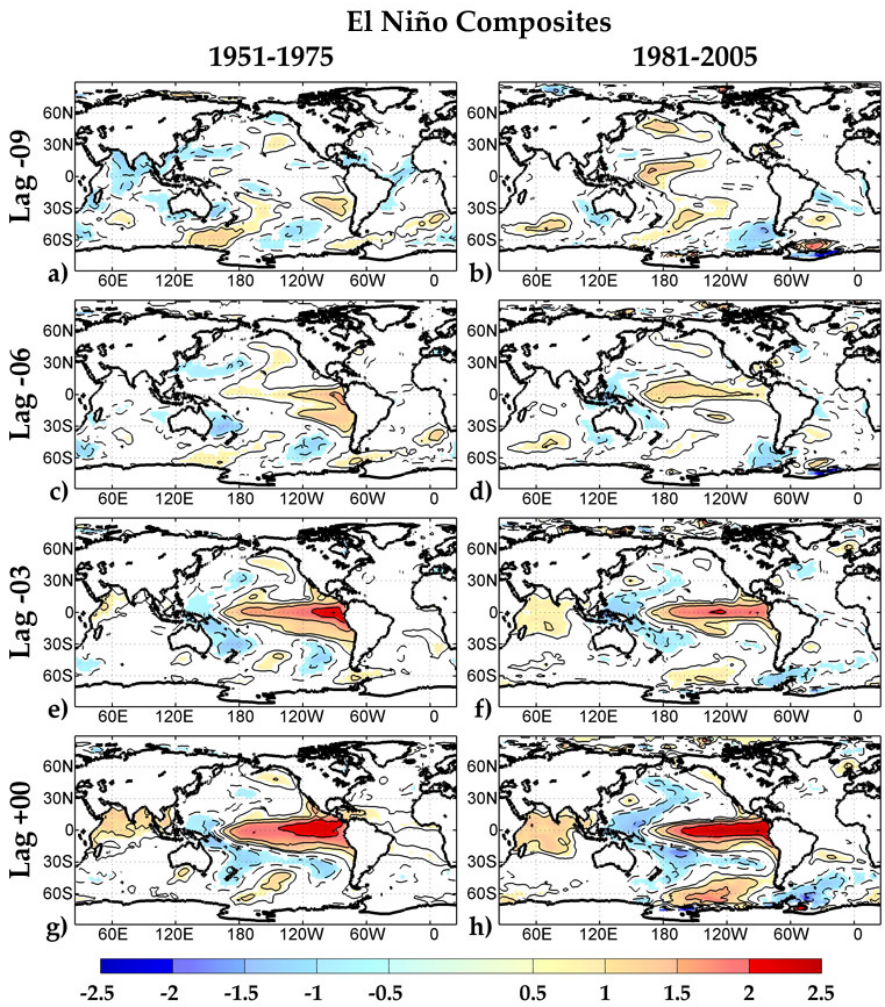

Figure 1.6: Standardized (unitless) interannual anomalies of SST, averaged for the 5 major EN peaks in 1951-1975 (a,c,e,g) and 1981-2005 (b,d,f,h). Anomalies are shown for monthly time lags -09 (a,b), -06 (c,d), -03 (e,f) and +00 (g,h) before the peaks. Colored areas indicate significant anomalies ($p < 0.05$).

western Pacific, increasing the surface stability there and leading to the formation of symmetric off-equatorial anticyclones. All these anomalies propagate eastwards, and thus pairs of cyclones and anticyclones follow one another. Another alternative theory is the recharge oscillator (Jin 1997), which assumes that the heat content in the tropical Pacific is progressively built up during the development of an EN event, and discharged or dispersed polewards after the peak. Note that the opposite mechanism is analogously proposed for LN events. An unified oscillator has even been proposed (Wang and Fiedler 2006), in which the relative contribution of each of these mechanisms varies in time. In short, this compromise between theories ultimately underlines the current limitations in our understanding of such a complex phenomenon.

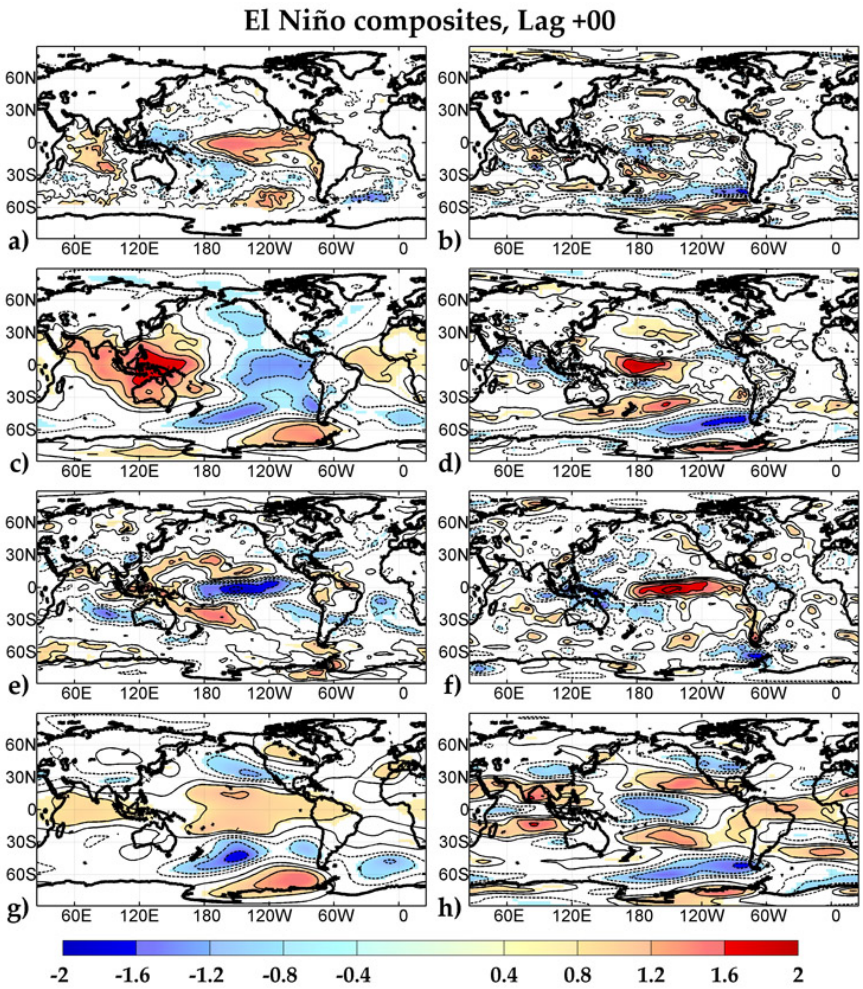

Figure 1.7: Standardized (unitless) interannual anomalies of SST (a), wind stress curl (b), SLP (c), surface zonal wind (d), outgoing longwave radiation (e), 500hPa vertical wind (f), 300hPa geopotential height (g), and 300hPa zonal wind (h); averaged for the 5 major EN peaks in 1986-2005. Colored areas indicate significant anomalies ($p < 0.05$).

1.2.2 Theoretical nature of ENSO

The mechanism proposed by Bjerknes, as well as the out-of-phase negative feedbacks stopping and reverting the loop, characterize the transition between ENSO states as a slow-varying change in the interaction between the ocean and the atmosphere. Against this conception of the evolution of ENSO, Wyrtki (1975) proposed an alternative theory, in which a sudden collapse of the easterly trade winds in the western Pacific could induce the formation of eastward Kelvin waves propagating the accumulated heat content in the warm pool and triggering an EN event in the eastern Pacific. In this theory, ENSO is considered as a highly damped oscillation sustained by stochastic forcing, and thus EN events are discrete, independent episodes interrupting longer periods of underlying neutral or cold ENSO conditions. This theory thus points to atmospheric noise as a trigger, explaining the historically irregular occurrence of EN

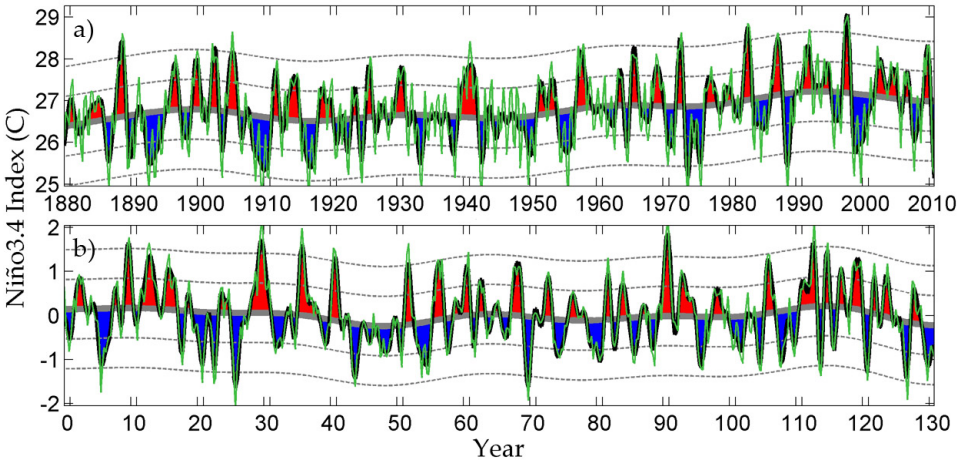

Figure 1.8: (a) Actual seasonal values (green) and interannual component (blue and red) of the Niño3.4 (N34) Index (in °C), which is defined as the regional average of SST in [170W,120W] × [5S,5N]. (b) As in panel a, but for a cyclo-stationary autoregressive process of order 2 modeling the SST variability of ENSO. Dashed lines indicate those events reaching the ±1 and ±2 standard deviation criterion.

events (Figure 1.8a). Note that multiple processes can exert this triggering role, but the main candidates appear to be the westerly wind bursts (Vecchi and Harrison 2000; Lengaigne et al. 2004) and the Madden-Julian oscillation (Zavala-Garay et al. 2005).

Although the contribution of atmospheric noise is unclear and not fully understood yet, it certainly plays a role in the evolution of ENSO. Time series often consist of two components, $X_t = S_t + N_t$, where S_t is a dynamically determined signal and N_t is a purely stochastic term. In the particular case of ENSO, the dynamically determined term is somehow affected by noise, and thus the oscillation is not always completely detectable. For example, Zwiers and von Storch (1990) described the SST time series of ENSO as a cyclo-stationary autoregressive process of order 2, i.e. $X_t = \alpha_{0,s} + \alpha_{1,s}X_{t-1} + \alpha_{2,s}X_{t-2} + \epsilon_{t,s}$, where subscript s indicates that coefficients $\alpha_{i,s}$ and the variance of the zero-mean white noise process $\epsilon_{t,s}$ are specified by invariant seasonal cycles (Figure 1.8b). Note that this kind of autoregressive process is characterized by the relative preponderance of maxima in a particular season, such as in ENSO. In this approximation, however, the dynamically determined term would decay to zero if the stochastic term was absent, i.e. $\epsilon = 0$, and therefore the oscillation is, at least, somehow affected by atmospheric noise (von Storch and Zwiers 1999).

Nonetheless, this mathematical model represents only a simplified approximation of ENSO (cf. Figures 1.8a and 1.8b). Climate variability in the tropical Pacific exhibits prominent frequencies of variability in a relatively large range of timescales, from 2 to 8 years only in the interannual band, and this fact seems to rule out the hypothesis of atmospheric noise as a necessary condition for the initialization of ENSO events. Indeed, warm events do not always follow episodes of sudden westerly winds, e.g. the 1982/83 EN event was not preceded by a collapse in trade winds. Moreover, atmospheric noise is not either a sufficient condition for the initialization of ENSO events. Vecchi et al. (2006) thus discussed the relative contribution of atmosphere-only and SST-induced zonal wind stress during westerly wind events a year before the large 1997/98 EN. Westerly wind bursts were indeed observed during the onset of this event, but the event was well captured in a set of retrospective forecasts of the ocean state after coupling only the (linear

plus non-linear) fraction of these westerly wind episodes related to large-scale tropical Pacific SST (Vecchi et al. 2006). This result showed that internal atmospheric variability unconstrained by low-frequency SST (i.e. atmosphere-only stochastic noise) had a minor modulation role, if any, in the evolution and subsequent amplitude of the event.

Large advances in the description of ENSO have been taken place since the eighties. The compilation of larger time series favored the view of ENSO as a self-sustaining interannual fluctuation, being chaotic yet deterministic. This alternative interpretation of the nature of the mode seems to be more plausible in the light of recent achievements in the prediction of ENSO events (Chen et al. 2004). Nevertheless, other aspects of the mode prevent scientists from being absolutely optimistic about this theory. For instance, differences in the magnitude and inhomogeneities in the precursory mechanisms still underline the irregular nature of the variability mode. Some authors have tried to reconcile both conceptions of the phenomenon, suggesting that EN is one phase of a weakly damped oscillation that is *sustained* (Philander and Fedorov 2003) or even *modulated* (Fedorov et al. 2003) by random noise. This compromise is found to be the best way to date to integrate both conceptions, explaining both the cyclic and irregular nature of ENSO, but, again, this compromise ultimately highlights our partial understanding of the phenomemon.

1.2.3 Predictability of ENSO

ENSO is the most energetic climate signal (except for the seasonal cycle), the major source of interannual variability, the main modulator of atmospheric variability and the most prominent driver of climate teleconnections. As such, the forecast of ENSO is a crucial factor for seasonal forecasting worldwide. The first successful attempts to forecast ENSO were done in the eighties, when the 1986/87 EN event could be anticipated 12 months in advance (Cane et al. 1986). This particular prediction was based on the Zebiak-Cane model, in which the characteristic spatial patterns result from the configuration of the mean wind, current and temperature fields (Zebiak and Cane 1987). Particularly, this model assumes that ENSO is a self-sustained ocean-atmosphere coupled mode in the tropical Pacific. The description of ENSO as a self-sustaining interannual fluctuation, being chaotic yet deterministic, implies that the prediction of ENSO is essentially constrained by the growth of initial errors. Under these circumstances, the potential prediction of ENSO is of the order of years. Alternatively, when ENSO is defined as a highly damped oscillation sustained by stochastic forcing, potential prediction of ENSO is also limited by random noise, and therefore the phenomenon becomes much less predictable.

ENSO is certainly predictable, but most of the climate models still exhibit large errors in the simulation of some of its main features, and as a result, current schemes are still far from the inherent limits to predictability. For instance, the mistreatment of surface fluxes partially explains why most of the coupled dynamical prediction schemes of ENSO exhibit large skill only in the central and eastern Pacific near the equator (Figure 1.9), an area where SST is mainly controlled by ocean dynamics (Chen and Cane 2008). Instead, thermodynamics plays a major role in the western equatorial Pacific. In addition, most of the climate models are unable to reproduce the transition between EN and LN, and vice versa (Ohba et al. 2010), and thus the (rather irregular) oscillating nature of ENSO.

Important achievements in operational forecasting have followed one another since the eighties (e.g. AchutaRao and Sperber 2006). Particularly, Chen et al. (2004) used a coupled ocean-atmosphere model that does not invoke stochastic forcing to successfully predict 6 of the 7 major EN events within 1857-2003 at lead times of up to two years. Prediction skill was however more modest at these or shorter lead times for less prominent events. Anyway, successful

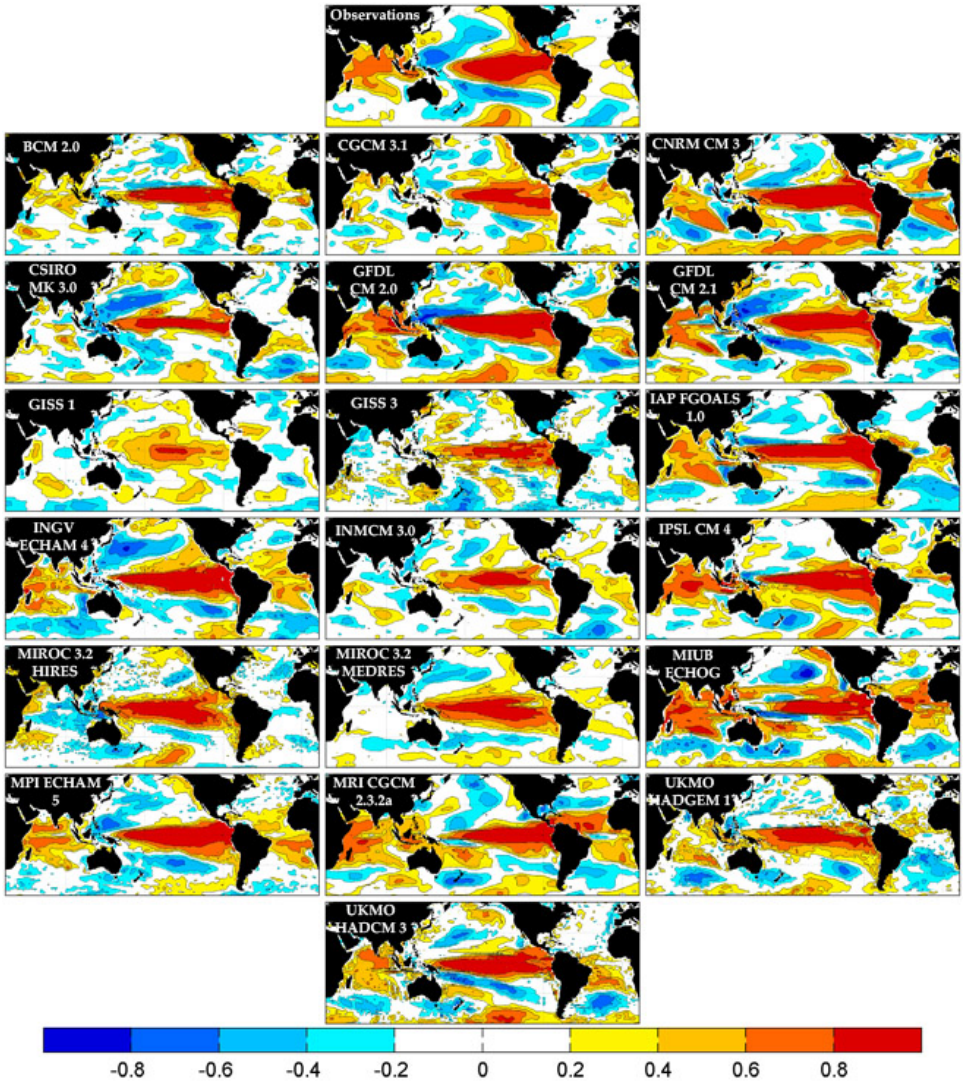

Figure 1.9: Year-to-year correlations between winter mean N34 Index and SST anomalies for observational data (top) and IPCC AR4 coupled ocean-atmosphere climate models (other panels).

forecasts at these lead times suggest a secondary role of stochastic processes in the initialization of ENSO events. Some authors have thus suggested that atmospheric noise is more likely to be an enhancer rather than a trigger for ENSO, so that the coupling between ocean and atmosphere mostly determines the phase and timing of ENSO, and noise modulates the subsequent evolution of anomalies around the peak (Chen and Cane 2008). According to this interpretation, the occurrence of EN and LN events can be potentially predictable at relatively long lead times, although the subsequent magnitude is subject to unpredictable irregularities generated by stochastic processes.

1.2.4 Precursors of ENSO

Some of the efforts in climate science have been directed to the search of precursors of EN events that could improve our understanding of the activation mechanism of the phenomenon, ultimately reverting to an improvement of the forecast in statistical or dynamical prediction schemes. Although some premonitory signals of ENSO are found in other basins (e.g. in the Indian ocean; Terray and Dominiak 2005; Izumo et al. 2010), the generation of an EN or LN event is traced through internal dynamics in the Pacific ocean, and essentially at tropical latitudes. Nonetheless, some precursors and independent variability modes in the subtropical and extratropical Pacific have also been identified.

Several premonitory signals have been described in the tropics, mainly in the western Pacific, and they essentially involve anomalies in SST, SLP, zonal winds, heat content, heat fluxes and equatorial waves (e.g. McPhaden and Yu 1999; Perigaud and Cassou 2000; Boulanger et al. 2001; Zhang and McPhaden 2008). Some authors pointed out that these processes are directly linked to ENSO itself (Weisberg and Wang 1997), although it is not completely clear which fraction of these features represents a particular leading phase of the oscillatory nature of ENSO. Kug et al. (2005) found positive equatorial heat content and western Pacific wind anomalies preceding winter N34 values by 10 months (find a detailed classification of westerly wind burst episodes in Vecchi and Harrison 2000). The exploration of optimal perturbations of SST (Penland and Sardeshmukh 1995; Chen et al. 1997), a technique that isolates the most rapidly growing perturbations in a system where dynamics is assumed to be linear (Farrell 1982), has also pointed to these and other precursors in the tropical Pacific (e.g. Thompson 1998; Moore and Kleeman 2001; Kug et al. 2010).

Other precursors of ENSO events have been found in the northern hemisphere, and they are essentially associated with the North Pacific Oscillation and its upper-troposphere equivalent, the west Pacific pattern. This mode of variability has been described as the second leading principal component of winter pressure in the central north Pacific, and it is basically characterized by an equivalent barotropic north-south dipole in the subtropics and to the north of the Aleutian Islands (i.e. around 60N; Linkin and Nigam 2008). Some studies have shown that a North Pacific Oscillation-like pressure configuration in the central subtropical Pacific (Vimont et al. 2003), as well as its subsequent SST footprint as a northern wing of a warm horseshoe (Anderson 2007; Figure 1.10a), precedes by one year the mature phase of EN events. Although the dipolar structure of the North Pacific Oscillation usually shows up in the analyses, other studies using different techniques only found the southern pole of this atmospheric feature (Anderson 2003, 2004; Figure 1.10c). Anyway, the SST anomaly then propagates to the central tropical Pacific, where it weakens the Walker circulation and favors the positive phase of ENSO. Although the transition between the precursor and the subsequent EN event is dynamically linked to the circulation induced by the southern pole of the North Pacific Oscillation, the ultimate origin of these precursors remains yet unknown. Other precursors in the northern hemisphere have been proposed at very long leads, such as an oceanic feature in the subtropical north Pacific 18 months before EN events (Chang et al. 2009), but the signal in this case seems rather weak ($r \simeq 0.25$) and supposedly non-significant.

In the southern hemisphere, the evolution of the ocean and the atmosphere before and during the mature phase of ENSO is intimately related to the Pacific-South American mode. This variability mode was described by Mo and Ghil (1987) and Karoly (1989) as an equivalent barotropic tripolar wavetrain pattern in the southern hemisphere extending eastwards and polewards from the western subtropical Pacific to the Drake Passage. Mo and Peagle (2001) indeed identified a Pacific-South American-like empirical orthogonal function that is excited during the mature phase of ENSO events (Figures 1.7c,g). Other similar patterns in the southern hemisphere have

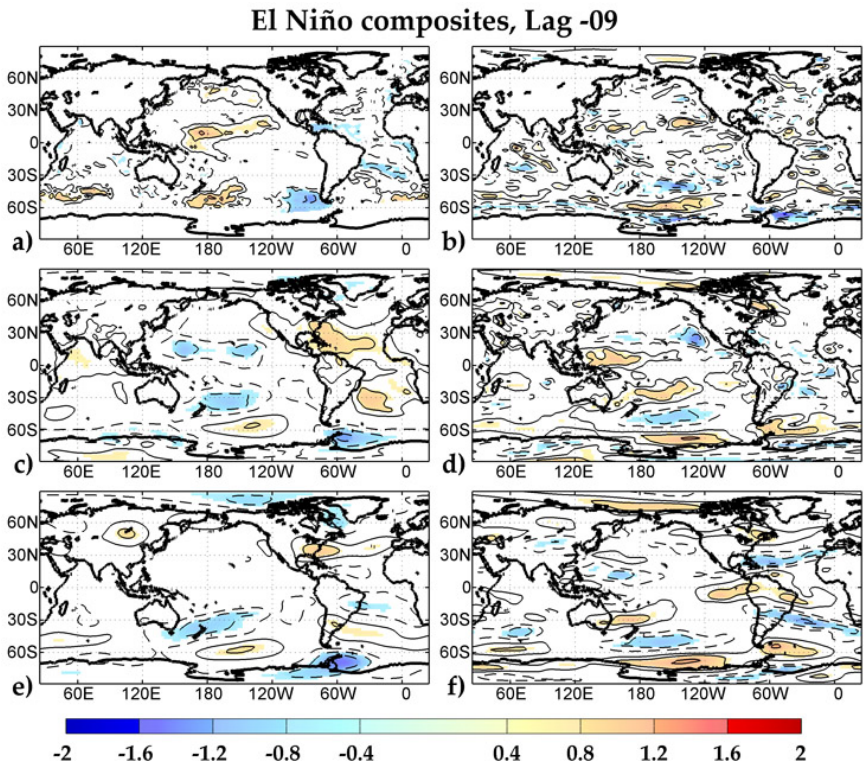

Figure 1.10: Standardized (unitless) interannual anomalies of SST (a), wind stress curl (b), SLP (c), surface zonal wind (d), 300hPa geopotential height (e), and 300hPa zonal wind (f); averaged for the 5 major EN peaks in 1986-2005. Anomalies are shown for monthly time lag -09 before the peaks. Colored areas indicate significant anomalies ($p < 0.05$).

been described as precursory signals leading to EN events by around 3 to 9 months (Wright 1993; Kidson and Renwick 2002; Jin and Kirtman 2009; e.g. Figure 1.10). In turn, Terray and Dominiak (2005) described the role of southern Indian ocean SST anomalies in the modulation of these Pacific-South American-like precursors after the 1977 regime shift, through changes in the southern Hadley circulation in the Pacific ocean.

The Antarctic Circumpolar Wave (ACW) has been proposed as a possible modulator of these precursory signals in the southern Pacific ocean, although the magnitude of its real contribution has generated scientific controversy (Hall and Visbeck 2002; White 2004), especially because the period with reliable data records is relatively short in the southern oceans. The ACW is a wavenumber 2 ocean-atmosphere wave circling Antarctica in around 8 to 10 years (Jacobs and Mitchell 1996). Initially thought to be mainly driven by the prevailing currents (White and Peterson 1996), subsequent modeling studies pointed out that its eastward propagation around the south pole depends upon coupling between the ocean and the atmosphere (White et al. 2004). Some authors found the origin of ACW SST anomalies in the western subtropical Pacific. These anomalies then spread south to the southern ocean, where they travel around Antarctica, finally propagating equatorward in the eastern Pacific, the eastern Atlantic, and the Indian ocean (Turner 2004). White and Annis (2004) showed that the magnitude and type of EN events before and after the regime shift in the seventies was influenced by the relative preponderance of these equatorward trajectories in the Indian and Pacific oceans (Figure 1.11). Thus, before (after)

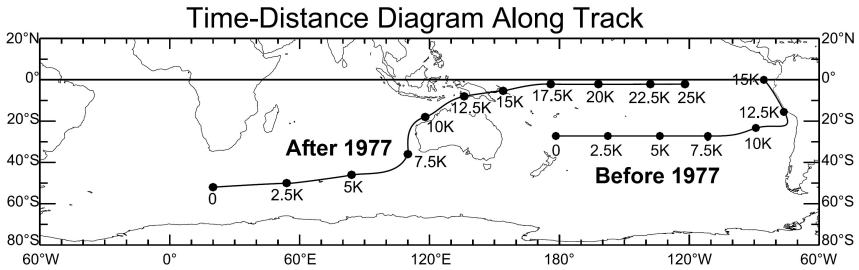

Figure 1.11: Dominant trajectories in the south Pacific and Indo-Pacific oceans along which covarying SST and SLP anomalies led to EN events during the epoch before and after the 1977 regime shift, respectively. Adapted from White and Annis (2004).

this regime shift, the propagation was especially active in the Pacific (Indian) ocean, leading to the formation of westward (eastward) EN events in the eastern (western) Pacific.

1.2.5 Teleconnections of ENSO

ENSO is the main driver of climate teleconnections. It can for example modify the latitudinal location and intensity of the midlatitude jet stream, and therefore it is able to generate well-defined wave trains propagating eastwards in both hemispheres (Hoskins et al. 1977). The prediction of ENSO is therefore a crucial element for seasonal forecasting worldwide, although other factors can interfere in these relationships, such as natural noise, non-linear feedbacks or planetary waves (Trenberth 1997). As a result of these interfering processes, teleconnections of ENSO events usually exhibit some degree of natural variability. In the case of the north Atlantic basin, for example, internal variability modes and atmospheric dynamics can explain alone some of these inter-event differences (Gouirand and Moron 2003).

Many authors have described the effects of an atmospheric bridge linking ocean anomalies in the tropical Pacific with lagged changes in other distant tropical and extratropical regions. Indeed, more than one fourth of the interannual SST variability in some distant regions is linearly explained by the Southern Oscillation Index alone (Klein et al. 1999). The original surface anomalies in the tropical ocean modify the evaporation rate, the tropospheric circulation and the cloudiness, and thus the net heat and momentum fluxes entering the remote regions (Rodó 2001). EN events are for example associated with the suppression of the Walker cell, and thus with an enhancement (weakening) of the Hadley circulation in the eastern Pacific (Atlantic and western Pacific) ocean (Wang 2005b). These circulation changes define a cross-shaped structure of upper-troposphere humidity anomalies around the central and eastern tropical Pacific (Klein et al. 1999). Note that this variable is typically used to highlight regions of large scale circulation anomalies and areas of active convergence and divergence (Soden 1998). As a result of this atmospheric bridge, northern extratropical anomalies in the surface ocean are normally more intense in late winter and early spring (Lau and Nath 2001).

Interannual variability in the Indian ocean is in close relationship with ENSO. During the mature phase of EN, the Walker circulation defines an area of anomalous subsidence in the western Pacific, with anomalous surface westerlies in the central Pacific and easterlies in the Indian ocean, which in turn drive warm surface waters from the western to the eastern Indian ocean. This mechanism defines the so-called Indian ocean dipole (Webster et al. 1999), a zonal dipolar configuration in equatorial surface temperatures. There is still a lot of controversy about

the dynamical relevance of this pattern. Some authors state that the Indian ocean dipole is an internal mode of variability in the Indian ocean (Saji et al. 1999), while others suggest that it is not independent from the Pacific ocean because much of its variability is simply explained by ENSO alone (Ihara et al. 2008).

The most remarkable impact of ENSO on the extratropical atmospheric circulation is associated with the Pacific-North American oscillation. This mode of variability has been described as the leading principal component of winter pressure in the central north Pacific, and it is basically characterized by a large equivalent barotropic pressure area near the Aleutian Islands (Linkin and Nigam 2008). The local impact of EN events resembles the negative phase of the Pacific-North American oscillation (Kumar and Hoerling 1997), and it is thus characterized by the strengthening and eastward displacement of the Aleutian low to the west coast of north America. On the other hand, the Aleutian low typically weakens in the case of LN events, but the displacement is smaller (Hoerling et al. 1997). This asymmetry is consistent with the non-linear response in the intensity of the Pacific-North American mode to ENSO forcing anomalies (Hannachi 2001).

Some of the remote ENSO-induced anomalies are however not instantaneous. The time lag between ENSO events and SST anomalies in remote tropical areas is about 1-2 seasons in the Indian ocean and 2-3 seasons in the Atlantic sector (Enfield and Mestas-Nuñez 1999). These lags correspond to the time period that is required for the activation of the anomalous atmospheric circulations cells, but also for the transmission of anomalies to the ocean (Latif and Grötzner 2000). In turn, these ocean anomalies can induce other atmospheric teleconnections through changes in heat and momentum fluxes in the ocean-atmosphere interface (Lau and Nath 2001).

1.2.6 Long-term changes in ENSO variability

Our knowledge of ENSO is limited essentially because we do not understand yet all the physical processes taking place in the Pacific ocean during the developing and decaying phases of EN and LN. Our understanding is however also intrinsically limited because we have only experienced a relatively small sequence of events. In the tropical Pacific, surface ocean records have been reconstructed back to the nineteenth century (e.g. Figure 1.8a), but the picture is less favorable in the southern extratropical Pacific, where data have been collected since the launch of meteorological satellites three decades ago. As an alternative approach, Wittenberg (2009) used a 2000-yr coupled atmosphere-ocean-land-ice run to describe some of the simulated low-frequency fluctuations in ENSO variability. This study showed that multidecadal periods with irregular magnitude, frequency and timing of events were followed for example by epochs of decreased variability, strong warm-skewed events or even regular sinusoidal oscillations. Note that this remarkable low-frequency variability in the properties of simulated events is in turn intrinsically constrained by our limited data records, which are commonly used to validate and tune the parameters defining the configuration of the model schemes.

Superimposed on this natural low-frequency variability, ENSO is also expected to change as a result of increasing GHG emissions in the atmosphere. During the last decades, a new type of EN event has been observed, the central-Pacific EN event, in which the largest SST anomalies are observed in the central Pacific and not in the eastern Pacific (Ashok et al. 2007). According to state-of-the-art climate model projections under GHG scenarios, this kind of EN event is expected to become more frequent during the present century (Yeh et al. 2009). This change in the relative frequency of EN types seems to be associated with a reduction of the Pacific Hadley and Walker circulations, which are expected to weaken the equatorial trade winds and to flatten the thermocline (Collins et al. 2010). Although these trends seem to point to a shift to EN-like

conditions, changes in the zonal SST gradient are not conclusive. Indeed, the thermocline is expected to rise up, and therefore changes in processes other than the Bjerknes feedback might be operating (Vecchi and Soden 2007). In conclusion, given the lack of consensus in state-of-the-art climate projections for the present century, it is still not clear how climate change will affect the physical mechanisms behind the generation of ENSO events.

Chapter 2

Motivation and objectives

Climate is certainly a determining factor of human activity. As such, it is present in the planning of the major social endeavors, but also in the daily life of individuals. The effects of climate are however the direct result of a myriad of superimposed processes occurring at very different timescales. For instance, impacts can either be the consequence of an unexpected tornado lasting for less than an hour, a season of low fishery activity caused by a teleconnection of ENSO, a multidecadal fluctuation with higher-than-normal precipitation due to a low-frequency oscillation in the thermohaline circulation, or a continuous but gradual and mostly imperceptible rise in temperatures as a result of climate change.

All these and many other factors define our notion of weather and climate in an unique realization. Therefore, there is an inevitable need for separating the set of processes defining the different components of the natural system, in order to comprehensively address the final causes of the natural variability that we experience. Even long time before the development of the necessary tools for the forecast of weather and climate, scientists have tried to explain and predict the natural factors that affect our welfare. Nevertheless, the fraction of climate that we experience is mostly governed by the naturally-oscillating variability of the atmosphere, which has a finite limit of predictability constraining our ability to anticipate its evolution. The limit of predictability discovered by Lorenz is thus transversally present at all timescales of interest in climate forecasting, from the prediction of monthly anomalies a season ahead to the study of changes in climate statistics after a slow-varying modification in an external parameter.

As previously mentioned, predictability is the study of the extent to which events can be known in advance. From a theoretical point of view, an event can be predicted when the distribution of climate states changes to some extent after an ensemble of data is taken into account. Although this rather theoretical notion is strikingly applicable to the whole range of timescales, it is differently implemented in each area of study. The present dissertation will particularly focus on two specific cases with high impact on the society. On the one hand, the change from a prior to a posterior distribution in seasonal forecasting is derived from ocean-atmosphere observations initializing a prediction scheme, and thus the gain in predictability is both determined by the atmosphere itself and the state of its boundaries. On the other hand, in climate change studies, both the prior and posterior distributions are determined by a larger climate distribution, and therefore the distribution of states of the climate system is conditioned by external climate-related parameters, such as the concentration of GHG.

At interannual timescales, the predictive skill is currently derived from the description of the state of the ocean and the continental land. This influence is exerted at relatively slow

timescales, and thus anomalies providing memory in the boundaries persist for much longer than atmospheric noise. ENSO is by far the most prominent modulator of atmospheric variability at interannual timescales, being the responsible for the most important impacts in continental areas near the Pacific ocean. For instance, among a myriad of many other effects, EN causes floods and landslides in the western coasts of central America, forest fires and air pollution in southeast Asia and Oceania, or droughts, crop failures and famine in southern Africa. It also drives large-scale climate teleconnections worldwide, and therefore the prediction of ENSO is a crucial factor worth taking into account for seasonal forecasting in many distant regions. Thus, the successful prediction of ENSO at long lead times would ultimately increase the global predictability of the climate system.

One of the main factors constraining our understanding of ENSO is the current limitations in the availability of long data records describing the natural variability in the whole Pacific ocean. ENSO seems to have exhibited large multidecadal fluctuations, with alternate periods of more or less regularity, intensity or symmetry. Although these oscillations have typically been inferred from long coupled ocean-atmosphere climate model simulations, the last change was strikingly captured by our monitoring systems. Thus, the regime shift in the seventies was a multidecadal change in ENSO variability of outstanding interest, since it defined a sudden change in the way that EN events are predominantly generated, propagate along the tropical Pacific, and transmit their anomalies to distant regions. As a result of the change in the onset of these events, scientists have tried to describe new precursors that could replace those premonitory signals that have become rather useless in the new dominant configuration. The description of new tracers enlarging the predictability of the phenomenon is one of the main goals of the first study included in this dissertation (Section 5.1). To this aim, the RossBell (RB) dipole is here defined. This oceanic feature is shown to anticipate all the recent eastward-propagating EN events by nine months, and to trace the transition between initial anomalies in Western tropical Pacific (WPAC) and the subsequent development of EN events.

Our knowledge of the theoretical nature of ENSO is also limited. No satisfying theory has been formulated to date reconciling both the cyclic and irregular behavior of the phenomenon. A comprehensive description of the extent to which ENSO is a highly damped oscillation sustained by stochastic forcing or a self-sustaining interannual fluctuation would have profound practical implications far beyond the theoretical sphere. Particularly, this characterization would shed light on the predictability of the phenomenon, determining whether the forecast of ENSO is largely limited by natural noise or *only* by the growth of initial errors. New insights about the theoretical nature of ENSO are also exposed in the first study, in which a Complex Empirical Orthogonal Function (CEOF) analysis is used to illustrate and discuss the extent to which the phenomenon is cyclic (Figure 2.1). Thus, this article shows that only a fraction of the variability occurring in the tropical and southern extratropical Pacific can be represented by a circular mode of variability, further confirming that none of the opposite theories completely adapts to the complex nature of ENSO.

The notion of predictability is somewhat different in climate change studies, where the underlying state of the system directly depends on the distribution of external climate-related parameters. In this case, the predictive skill that is gained in the forecast is understood as the sensitivity of the climate system to a change in external factors. It is well known that the increase in the concentration of atmospheric GHG has led to a detectable anthropogenically induced global warming since the mid twentieth century. Even this simple and generic statement referring to the long-term global mean value in a relatively simple variable has required alone years of investigation in climate modeling and forecasting. A myriad of other more subtle changes are however expected for the coming decades, both at the global and regional scale, and therefore they are requiring intensive efforts from the climate community. In the present disser-

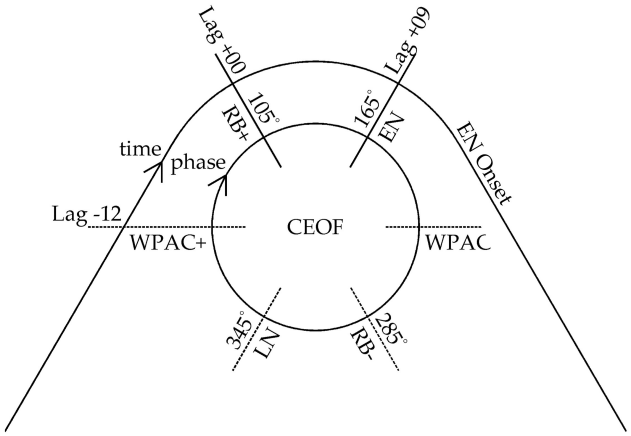

Figure 2.1: Schematic representation of the relationship between lead-lag composites showing the onset of EN events (non-circular curve) and the CEOF describing the cyclic component of ENSO (central circle). The distance between these lines represents a schematic measure of the similarity of results at a particular lag/phase. Adapted from the article in Section 5.1.

tation, the study of GHG-induced climate changes has been restricted to surface temperatures in Europe, since the old continent emerges as a critical responsive area to global warming.

Increasing attention to the effect of climate change on temperature extremes has been paid for the lasts years. The record-breaking 2003 heat wave in western Europe captured the attention of the climate community and policy makers, since it was an extremely unlikely event given the observed warming at that time. The posterior evaluation of its effects on the European society indeed offered a worrying glimpse of future conditions in summer. Recent updates of summer heat-related mortality thus rose the initial estimations up to 70000 excess deaths, of which 11000 occurred in June, 10000 in July, 15000 during the first week of August and 24000 during the second week (Robine et al. 2008). Nevertheless, the abnormal summer 2003 only exhibited slightly warmer temperatures in some local areas in France, Switzerland and northern Italy, compared to future climatologically normal summers simulated for the end of the century (Figure 2.2). As a result, scenario simulations seem to suggest that the unprecedented impact of this particular event could indeed represent a conservative glimpse of future conditions in summer.

As a rule of thumb, future heat waves are expected to become more intense, more frequent and longer lasting as a result of climate change (e.g. Meehl and Tebaldi 2004). Nevertheless, it is not clear whether this transformation in the statistics of temperature extremes is due to intrinsic changes in the generation of extremes themselves, or if they are instead a simple consequence of a more general change in the distribution of temperatures. For example, if the same threshold is used for the definition of extreme event in both the present and future climatology, then an increase in intensity, frequency and length of extremes automatically follows the widening and shift to warmer values of the temperature distribution (Figure 2.3a). Instead, if the threshold used for the characterization of extremes evolves in parallel with the changing climate, some of the statistics of these events might not change in the same way (Figure 2.3b).

Note that the way how information and conclusions from climate change studies are transmitted to the climate community and the general public are somehow influenced by the design

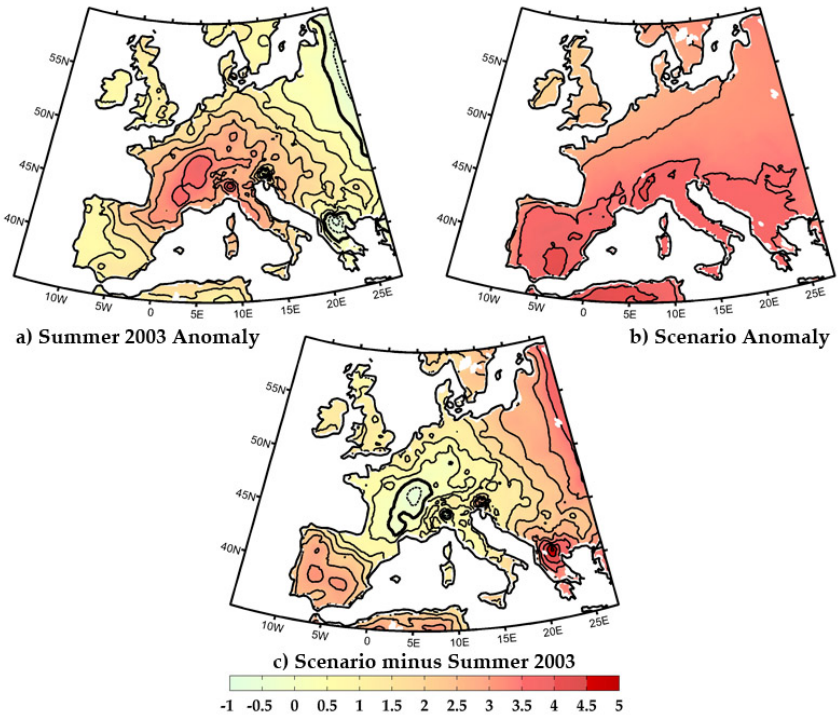

Figure 2.2: (a) Summer 2003 temperature anomalies relative to 1998/2002 (in °C). (b) Scenario changes in summer mean conditions for the period 2090/2100 relative to 1998/2008 (°C). (c) Comparison of summer 2003 anomalies with regard to scenario changes (°C).

of the methodological approach. The sensitivity of the shape of the temperature distribution to the evolution of the concentration of GHG thus appears to be an important issue, not yet comprehensively studied. The dependence of such a fundamental question on the criterion used to characterize the set of future extremes motivated the second study included in this dissertation (Section 5.2), in which an attribution experiment is performed in order to determine whether anomalies in extreme temperatures are significantly larger than changes in less extreme percentiles. In this way, the study determines whether future changes in the most damaging European heat waves are expected to mostly follow the summer mean warming. If so, the projected strengthening of future temperature extremes should be essentially considered as a simple consequence, or particular case, of a broader shift to warmer values of underlying summer base conditions.

The sensitivity of the temperature PDF to the concentration of GHG has profound implications for the welfare of human societies. This dependency is of particular interest for the most extreme temperature percentiles, which are those events with the largest relative impact on the society. By construction, however, these events are rare, and therefore their effect might be in some cases rather small in absolute terms. For example, Figure 2.4 shows that the average number of deaths in Europe is about 24 cases per million in a comfortable day (+17.5°C), 26 cases per million in a climatologically normal day (+12°C), and around 30 cases per million in an extremely cold day (-2.5°C). Despite the incidence is clearly larger in the distribution tails, and therefore these temperature percentiles have a larger contribution to overall incidence, the relative frequency of central temperatures is of an order of magnitude higher. As a result, in

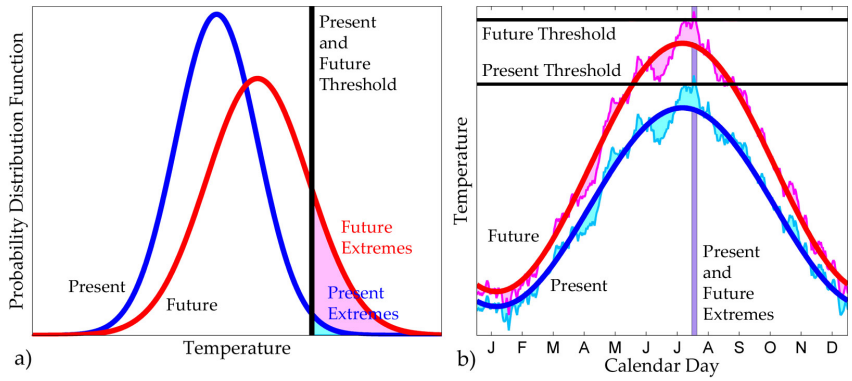

Figure 2.3: (a) Change in the frequency of extremes as a result of an increase of both the mean and standard deviation in the temperature distribution. In this case, extremes in the future climate are defined as those events reaching an extreme threshold in the present climatology. (b) Decomposition of present and future temperatures into seasonal base conditions (solid lines) and deseasonalized anomalies (dashed areas). Changes in the annual cycle correspond here to an increase of both the mean and standard deviation of the temperature distribution, as shown in panel a. In this case, however, the threshold defining the set of warm extreme temperatures warms in parallel with underlying summer base conditions. Adapted from the article in Section 5.2.

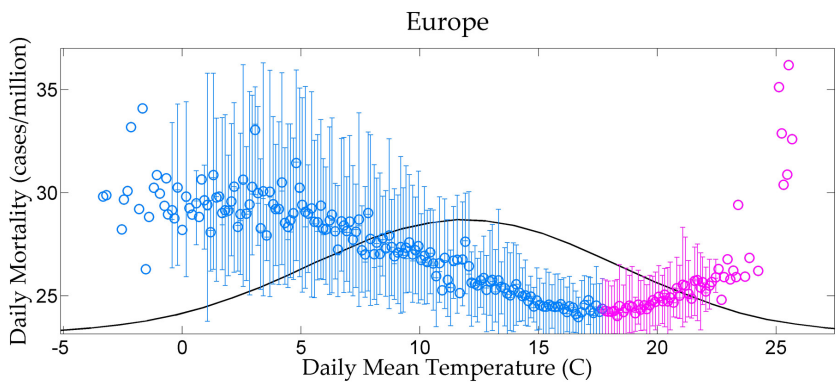

Figure 2.4: Relationship between daily mean temperature (°C) and mortality (cases/million) in Europe. Values within the warm and cold tails are represented in red and blue, respectively. Circles correspond to temperature and mortality averages of daily data within each equally spaced temperature interval, and vertical lines to 90% confidence intervals of daily mortality data within each temperature interval. The temperature PDF is shown in black.

the particular case of temperature-related mortality, the total number of deaths is determined by changes in the frequency of the whole range of temperature percentiles.

Hence, every single approach designed for the estimation of this kind of temperature-related impact needs to satisfactorily reproduce the change in the distribution of temperature throughout the whole range of percentiles. The feasibility of this challenge motivated the third study included in this dissertation (Section 5.3), in which a simplified methodology is presented in

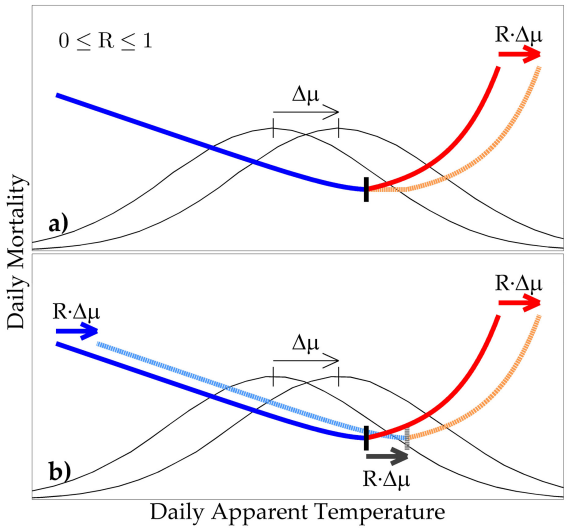

Figure 2.5: Scenarios of adaptation to temperature-related mortality, with only decreased susceptibility to warm temperatures (a), or also increased susceptibility to cold temperatures (b). Factor $0 \leq R \leq 1$ here represents the acclimatization pace of the society to an increase $\Delta\mu$ in annual mean temperatures, with $R = 0$ ($R = 1$) defining a scenario of no (immediate) acclimatization. Adapted from the article in Section 5.4.

order to reproduce the simulated changes in basic temperature statistics from a relatively small set of parameters describing the future evolution of the shape of the PDF. Thus, the change in frequency, length and intensity of warm, neutral and cold temperatures is here derived from the knowledge of changes in only three central statistics, the mean, standard deviation and skewness of the temperature PDF, for which current climate models are better suited. Note that this simplified methodology could be of great help for non-specialized institutions working on climate impacts that do not have direct access to climate change simulations.

This approach could be used, for example, to model the effect of rising temperatures on mortality. In a context of global warming, the future evolution of total mortality will be the result of an hypothetical balance between decreased deaths from cold temperatures and increased deaths from warm temperatures. Note that no scientific study have comprehensively modeled the future evolution of mortality in Europe due to both increased heat and reduced cold temperatures in a warming climate. Scientific studies have been traditionally restricted to changes in either the warm or the cold tail, performing an exhaustive analysis for a single country or deriving changes at the continental level from time series in no more than 15 cities or small regions. This lack of knowledge about future changes in both heat- and cold- related mortality in Europe motivated the last study included in this dissertation (Section 5.4). In this work, daily meteorological data and numbers of deaths were used to describe this link in nearly 200 European regions representing more than 400 million people, and these relationships were subsequently used to infer projections of mortality under GHG emission scenario simulations. The causes that explain the magnitude of changes in total mortality were studied in depth. Particularly, the relative contribution of the shift in annual mean temperatures and the change in the shape of the temperature PDF were separately addressed.

Anthropogenic GHG emissions are rising global temperatures much faster than any natural

source of variability at similar low-frequency timescales, and therefore it is not clear whether a technologically-developed society like Europe will be able to deal with the negative consequences of such a major change. Temperature trends have been small to date, at least compared to the change that is projected for the coming decades, and therefore the effect of warming temperatures is completely uncertain. It is thus highly speculative to make a guess about the extent to which the susceptibility to warm (cold) temperatures will decrease (increase) in a context of general temperature rise. Some scientific studies have theorized the future evolution of the relationship between temperature and mortality, but very few of them have quantified this change by inferring projections of mortality. Note, however, that none of them described the combined effect of a change in the susceptibility to warm and cold temperatures, and comprehensively analyzed the balance between both contributions at the continental level. All these factors were also considered in the last study included in this dissertation. Thus, projections of mortality were expressed as a function of different levels of acclimatization to future warm and cold temperatures (Figure 2.5), mimicking the scenarios of GHG emissions used in the IPCC reports.

Chapter 3

Discussion and conclusions

The predictability of two particular types of phenomena was studied in the present dissertation. On the one hand, a new extratropical precursor of recent eastward-propagating EN events was shown to trace the transition between an initial warming in the WPAC area and the subsequent development of warm anomalies in the central and eastern tropical Pacific. On the other hand, the sensitivity of the distribution of European temperatures to the increasing concentration of atmospheric GHG was studied by addressing the effect of these changes on the population.

The role of the RB dipole as a new premonitory signal for the onset of recent eastward-propagating EN events was established in the first article of this dissertation. This extratropical tracer was shown to be followed by EN events around 9 months later, but at the same time, to occur a year after the development of a warm oceanic area in the WPAC region. This initial anomaly appeared to generate an anomalous wavetrain extending eastward and poleward in the southern hemisphere, which is in turn associated with the generation of the oceanic RB feature. The article also showed that changes in the atmospheric circulation lead to warm SST anomalies in the central tropical Pacific, being later enhanced by suppressed equatorial easterlies. These processes appeared to be linked to an eastward shift in the convection, and thus a weakening of the Walker circulation, setting up the positive Bjerknes feedback that exponentially grows on top of the incipient warming and leads to the mature phase of EN.

Indeed, both the initial warming in WPAC and the RB dipole strikingly enlarge the predictability of EN to some extent. Figures 3.1a,b thus show the change in the distribution of SST in the N34 region after conditioning the PDF to the prior occurrence of major interannual peaks in the WPAC and RB areas, respectively. Due to the limitations in data records describing the state of the ocean surface in the southern extratropics, results were here derived from a long-term simulation of the GFDL CM2.1 coupled model (find further details of the model configuration in Gnanadesikan and Anderson 2009), and therefore the change in the distribution is representative of a large number of events. In addition, the combined effect of precursory ocean anomalies in both regions is shown in Figures 3.1c,d, in which the distribution is conditioned by both the occurrence of major interannual peaks in WPAC and the sign of interannual SST anomalies in the RB region, and vice versa.

Results show that the median of the distribution of the N34 Index is around +0.9°C (+1.5°C) warmer after taking into account the major interannual peaks in the WPAC (RB) region. The shift to warmer values is even larger, of up to +2.65°C, when the state of the ocean surface in both areas is combined. In this particular situation, the change in the distribution is so large

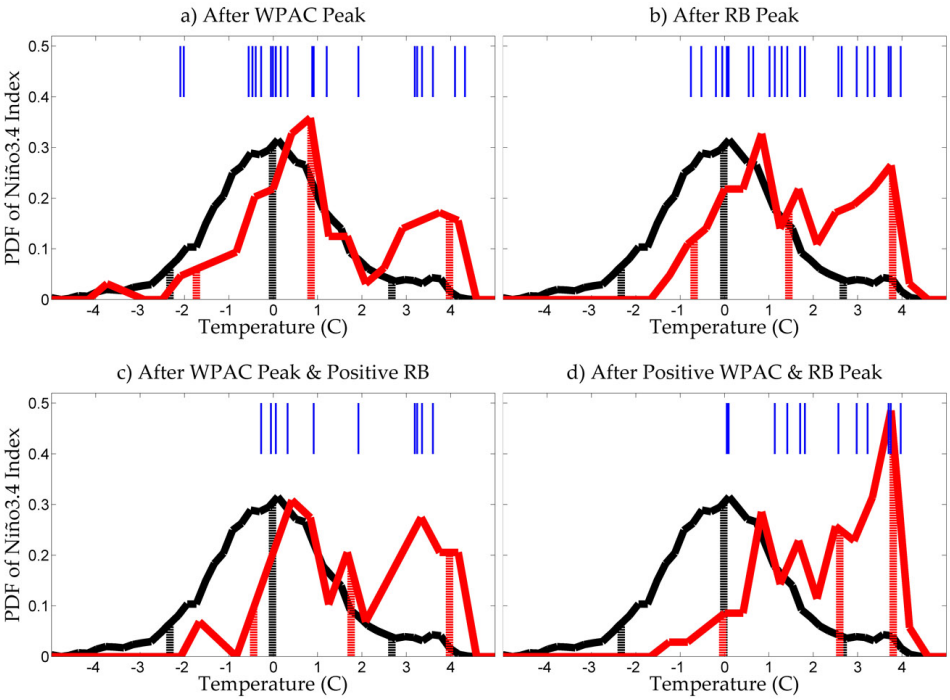

Figure 3.1: Predictability of EN derived from WPAC and/or RB in a 500-year simulation of the GFDL CM2.1 coupled model. The PDF of the simulated monthly time series of the N34 Index is here shown in black as a reference distribution (see all panels). Note that the whole period mean of the N34 Index was previously subtracted. The PDF was then conditioned by the 22 major interannual positive SST peaks in the WPAC (a,c) and RB (b,d) regions. Thus, the red distribution in panel a displays this same PDF, but restricted to only those months occurring +18, +19, +20, +21, +22, +23 or +24 months after a WPAC peak. As a particular case, N34 values at lag +21 are indicated in blue. Note that dashed vertical black and red lines correspond to percentiles 5, 50 and 95 of the corresponding distribution. (b) As in panel a, but for lags +09 (vertical blue lines) and +06 to +12 (red PDF) after RB peaks. (c) As in panel a, but results are only shown for those WPAC peaks that are followed by interannual positive SST values in the RB region $21 - 9 = 12$ months later. (d) As in panel b, but results are only shown for those RB peaks that are preceded by interannual positive SST values in the WPAC region $21 - 9 = 12$ months earlier.

that only 5% of monthly data in the conditional distribution are cooler than the median of the original PDF. The distribution indeed becomes rather bimodal, where modes correspond to weak-to-moderate and very strong EN events. These changes thus highlight the potentiality of both regions in the forecast of EN, since they effectively enlarge the predictability of the phenomenon by shifting the distribution to clearly warmer values.

Interestingly, RB maxima lead to EN events by around nine months on average, and therefore they typically occur in late winter or early spring. Although the RB dipole stands on the edge of the spring barrier in the prediction of ENSO, earlier stages of the interaction between the WPAC feature and the RB dipole could be used to improve our ability to predict the phenomenon at longer leads. Alternatively, the low-frequency variability in the WPAC region is shown to anticipate the RB dipole by around a year, and therefore this statistical relationship could be

used to predict the southern extratropical precursor of EN. Note, however, that this two-step prediction approach still requires further research.

ENSO can be considered as a basin-wide phenomenon, since most of its leading and lagged teleconnections interfere with some of the dominant modes of variability in the southern and northern extratropical Pacific. Nonetheless, the search of new precursory signals of EN has been traditionally restricted to the tropical Pacific, especially near the equator, where ENSO anomalies exhibit larger intensity. This is the case, for example, of optimal perturbations of SST, due to the computationally intensive nature of the calculations. As a result of the relative lack of premonitory tracers at extratropical latitudes, most of the precursors used in state-of-the-art statistical prediction schemes of EN are located in the tropical Pacific, such as an increase of ocean temperatures and heat content in the western Pacific, or a weakening of easterly trade winds in the central equatorial Pacific.

The first article thus described a new premonitory signal of EN events in the southern extratropics. This precursor might solve the general lack of leading tracers at these latitudes, which is partially caused by our rather limited data records describing the state of the ocean surface. Time series in the southern extratropics have only been available since the launch of meteorological satellites in the eighties, clearly limiting our understanding of the dynamical processes occurring in the ocean at these latitudes. Up to our knowledge, the RB dipole represents the first oceanic index leading to EN that is defined in the southern extratropics, and as such, it is an outstanding feature worth taking into account. This precursory signal is indeed a feature relatively less dependent on ocean-atmosphere coupled dynamics in the tropical domain, suggesting that it could help to complement the prediction skill of current statistical forecast schemes of EN.

There are other reasons that justify the use of this precursory signal in operational prediction schemes. First of all, the RB dipole is an oceanic feature, and therefore its time series is not largely affected by the characteristic and less persistent timescales of atmospheric noise. For instance, the statistical relationship between EN and the atmospheric tripole was also evaluated, and appeared to be lower compared to that of the ocean counterpart. Secondly, the RB dipole is a privileged external observer of processes occurring at tropical latitudes during the generation of EN events. As such, it is neither embedded in nor interacting with those mechanisms directly explaining the growth of EN anomalies in the central and eastern tropical Pacific, and therefore it offers a new perspective of the phenomenon. Finally, it is a nearly 90° out-of-phase feature in the transition between processes in the western and eastern tropical Pacific. It thus represents an orthogonal and equidistant phase in the evolution between warm anomalies in WPAC and EN events.

The RB dipole is a leading feature of the mature phase of EN, but it cannot be considered to lead the whole ENSO phenomenon. It is thus anticipated by other earlier stages of ENSO within the overall envelope of the variability mode. The RB dipole is therefore a feature that masterfully traces this large-scale transition, of which we had only a partial and disconnected idea until now. Examples of features belonging to this transition are, for example, the elongated warm SST anomaly occurring in the northern tropical and subtropical Pacific a year before EN events, or the weak upper-troposphere anomaly wavetrain extending poleward and eastward in the southern Pacific during the developing stages of EN. The RB dipole strikingly links all these features in a coherent sequence, and further extends this evolution backwards in time up to almost two years. It also highlights the role of WPAC as the initial origin of all these features during the onset of recent eastward-propagating EN events.

This ordered sequence of patterns first occurring in the western tropical sector, later in the central tropical and southern extratropical Pacific, and finally in the eastern tropical area, is

clearly exemplified by the CEOF analysis applied to ocean surface data in the southern high-latitudes, further confirming the role of this region as a privileged observer of tropical dynamics. This methodology was compared to lead-lag peak composites of the RB dipole, and key differences were identified. Due to the rotating nature of the CEOF, which optimizes the extraction of features involving a large amount of variance in a cyclic component, this analysis masked the incipient warm spot in the WPAC region and showed up instead the much wider LN-like warm horseshoe. Nevertheless, this cyclic mode only represents a simplification of the real transition, which is far more complex. Indeed, different processes interact during the generation of the initial WPAC warm spot in each particular individual event. As an example, a moderate LN event was observed the year before the RB maximum in spring 1997, but not before the RB event in spring 1991.

The changing relationship between the RB dipole and preceding LN-like conditions in WPAC further exemplifies the complex nature of the phenomenon. Thus, ENSO sometimes seems to be a regular periodic mode, in line with a self-sustaining interannual oscillation, but in other cases EN episodes do not appear to be necessarily preceded by LN events or LN-like conditions, suggesting a more sporadic and punctual behavior. The asymmetry between EN and LN is for example illustrated by the recent multidecadal time series of the N34 Index (e.g. see 1977/2006 in Figure 1.8a), which was characterized by few LN events and relatively long periods without LN-like conditions. Interestingly, the rotating approximation shown by the CEOF analysis exhibited maximum resemblance during the positive phases of anomalies in the WPAC, RB and N34 regions, and failed to reproduce the observed interactions between the corresponding cold stages. The physical mechanisms responsible for this asymmetry will require further research, but they might be linked to nonlinearities in ocean dynamical heating and SST-forced tropical deep convection in WPAC, which seem to define the way in which tropical variability affects the extratropics. Note, however, that this asymmetry might correspond to a particular multidecadal phase in low-frequency variability of ENSO.

The transition between WPAC and EN inferred from the RB composites is in principle only valid for recent eastward propagating events, which have become the dominant type of EN during the time period with available satellite observations. It is however difficult to determine whether other oceanic tracers in the southern high-latitude Pacific were really observed in previous time periods when other types of EN predominated. Another issue that still remains open to further investigation is the origin of the initial perturbation in the WPAC region. It might be for instance related to high-frequency stochastic forcing and the subsequent collapse of the trade winds in the WPAC region. Alternatively, these initial perturbations might be modulated by low-frequency coupled ocean-atmosphere processes occurring in the southern latitudes. Thus, White and Annis (2004) showed that the magnitude and type of EN events before and after the regime shift in the seventies was influenced by the relative preponderance of equatorward trajectories in the Indian and Pacific oceans. Thus, before (after) this regime shift, the propagation was especially active in the Pacific (Indian) ocean, leading to the formation of westward (eastward) EN events in the eastern (western) Pacific.

The origin of initial perturbations in WPAC is thus a major line of research requiring further investigation. In this sense, the description of the role of preceding LN events might shed light on the theoretical nature of the phenomenon. In addition, it is essential to further characterize the dynamical mechanisms explaining the interaction between the tropics and the extratropics during the onset of EN events. In this line, the analysis of climate simulations from those models reproducing these processes might help to characterize the teleconnections. Particularly, pacemaker experiments with suppressed or enhanced variability in key regions, such as WPAC or RB, could be conducted to describe these interactions. Also, the description of scenario changes in the relationship between WPAC, RB and EN in a context of global warming might provide

additional clues for understanding the future evolution of ENSO variability. Last but not least, the inclusion of WPAC and/or RB in operational statistical prediction schemes might ultimately lead to an improvement in the predictability of the phenomenon.

The predictability of changes in the distribution of European temperatures in a context of global warming was studied in the other articles included in this dissertation. Particularly, article two focused on changes in future intense, widespread and persistent heat waves with large impact over central Europe. Note that this triple criterion defines the set of the most damaging events. As a rule of thumb, moderate long episodes have larger impact on human health than intense short events, and therefore the persistence criterion was also included. This study showed that the increasing intensity of these damaging summer heat waves is mostly due to higher summer base temperatures, and not to specific changes in temperature extremes. This result was shown to be compatible with previous studies stating that future heat waves will be more intense, more frequent and longer lasting as a result of climate change.

Although specific changes in deseasonalized heat wave anomalies are projected to be relatively small, this study demonstrated that 36% to 47% of future July and August days at the end of the 21st century in central Europe are projected to be extreme according to the present-day climatology and notion of extreme event. This increase in the frequency of extreme days is however the result of general changes in the whole distribution, and not due to intrinsic modifications in the PDF tail. In particular, the simple shift of annual mean temperatures to warmer values, as well as the widening of the temperature PDF in response to an increase of the summer-winter temperature range, essentially explain the significant rise in the intensity and number of warm tail days.

In our alternative methodology, instead, the characterization of the threshold defining the notion of extreme event evolves in parallel with the changing climate. As a result, scenario anomalies in extreme events do not depend on changes in non-extreme percentiles of the PDF, as it is customary in the literature. Temperature values were additionally decomposed into two terms, one depicting the mean seasonal cycle, and the other the deseasonalized anomalies. According to this decomposition, the deseasonalized term remains approximately constant after a simple shift of annual mean temperatures and a change in the amplitude of the summer-winter range (i.e. change of the distribution mean and standard deviation, respectively). Therefore, our results showed that changes in heat wave statistics are essentially attributable to a change in summer base conditions, or equivalently, to a change that mostly affects the shape of the PDF within the range of non-extreme percentiles.

From a practical point of view, these results have important implications for the predictability of the most damaging heat waves in Europe, their associated impacts and the design of future response policies. Indeed, future changes in the intensity of extreme heat waves can be solely based on anomalies in summer base temperatures, for which current climate models are better suited. Therefore, in a first-order approximation, adaptation policies can be essentially based on projected summer base conditions, rather than on changes in the PDF tail. This conclusion was further explored in the third article included in this dissertation, in which a simplified methodology was shown to reproduce the simulated changes in basic temperature statistics from a relatively small set of parameters describing the future evolution of the shape of the PDF.

Article three thus showed that scenario changes in extreme and non-extreme warm and cold temperatures can be successfully reproduced to a first-order approximation after applying the simulated changes in the PDF mean, standard deviation and skewness coefficient to the distribution in the control simulation. Nonetheless, in accordance with article two, changes in the mean and standard deviation were enough to reproduce the anomalies in the warm tail, and thus the change in the skewness coefficient only appeared to play a crucial role in the case of

the cold tail. Note that intensity changes throughout the whole distribution were shown to be non-trivial and clearly not uniform for the whole range of percentiles, as they are the result of the interaction of changes in multiple processes at several spatiotemporal scales. As a result, the set of extreme events in both tails is expected to warm to a larger extent than the central values of the distribution. Despite this non-uniform behavior, our simple double (triple) transformation appeared to strikingly deal with the major projected changes within the warm (cold) tail.

Since anomalies in the mean, standard deviation and skewness are essentially determined by changes in the portion of variability that corresponds to non-extreme percentiles, this article showed that general changes in the whole distribution, rather than specific changes in both extreme tails, can explain most of the magnitude of projected scenario changes in the structure of the tail itself. In a context where climate predictability is defined as the sensitivity of statistics to the concentration of GHG, this result has direct implications for the predictability of extreme events, since anomalies can be easily characterized and reproduced by means of a simplified methodology. From a practical point of view, changes in these three central statistics, for which current climate models are again better suited, can be used in those impact studies in which the temperature distribution mostly determines the degree of affectation. Note that this result is coherent with article two, which showed that changes in extreme heat waves are essentially similar to those in summer base conditions. The range of application of these findings is remarkable. This methodology could be applied, for example, to the study of the effect of temperature rise on infectious diseases, warm and cold-derived illnesses, or any other aspect concerning socioeconomic organization that might be linked to the frequency, length and/or intensity of temperature events (e.g. energy demand).

As an example, this methodology was applied to the relationship between temperature and mortality. Thus, the last article included in this dissertation described the connection between daily temperatures and relative counts of deaths for an unprecedented set of regions in Europe, and used these associations to derive projections of mortality according to temperature simulations under GHG emission scenarios. This work indeed showed that the simplified methodology described in article three effectively reproduces the projections of mortality directly inferred from the temperature simulations. As a result, the amount of information provided by the three central statistics of the distribution, namely the mean, standard deviation and skewness, was shown to be large enough to derive virtually equal projections of mortality. As shown in article three, however, the role of the skewness coefficient appeared to be crucial, and thus the projections of mortality were shown to be underestimated when changes in this parameter are not taken into account.

This work also showed that we might face a minimum in total mortality in three decades. Nevertheless, this initial reduction will be followed by a continuous and monotonic rise in the overall incidence, which is expected to compensate the initial decrease during the second half of the present century. In a context of temperature rise, the difference between annual mean and comfort temperatures, which ranges from +3.5°C to +12.5°C throughout Europe, could wrongly suggest a shift to more comfortable conditions in terms of mortality in the mid- and long-term, i.e. an increase in the relative occurrence of comfortable days. Indeed, a minimum in mortality would be reached at the end of the present century if only annual mean temperatures were shifting (i.e. invariant shape of the PDF: M-A1B scenario).

The real picture is however much more complicated, since two other factors are opposed to this behavior, explaining the non-trivial projected evolution in the incidence of total mortality. The large difference between projections of mortality in scenarios M-A1B and MS-A1B or MSW-A1B indicates that changes in the shape of the PDF will indeed play a crucial role. Thus, when changes in the standard deviation and the skewness coefficient are taken into account, total

mortality keep decreasing at a nearly constant rate until 2025-2055, but then the number of deaths starts to monotonically rise. Note that the future evolution of both coefficients was shown to lead to an increase in the frequency of warm extremes and a decrease in the number of comfortable days. The contribution of new (extreme and non-extreme) warm days will be even larger given the asymmetric shape of the transfer functions, which defines a scenario in which mortality rise due to the set of new warm days will compensate in the mid-term the decline in deaths as a result of the lower occurrence of cold days.

Our analyses also point to a change in the seasonality of mortality, with maximum monthly incidence shifting from winter to summer, although cold-related mortality will still represent the major factor contributing to the total number of deaths, compared to the relatively lower incidence of heat effects. As a matter of fact, although these changes are actually foreseeable, they have important implications for human welfare, health surveillance systems and adaptation plans. Thus, causes for increased risk of death are completely different under heat and cold conditions, and therefore the general shift to warmer conditions might change the profile of those individuals that are mostly susceptible to non-comfortable temperatures. In addition, it might require a substantial reconsideration of decision-making policies, as a result of the new challenges that the society will have to tackle.

The role of humidity was also taken into account for the estimation of the relationship between climate and mortality, since it affects the body's ability to cool itself by evaporation and perspiration during hot days. The decrease in atmospheric (and soil) moisture explains a fraction of the simulated warming, but this additional rise in temperatures does not equally contribute to an equivalent increase in heat stress. Although the decrease in relative humidity slows down the increase in the projections of mortality, it does not affect the main qualitative conclusions of article four, since the fraction of days in which humidity enhances the effect of heat is relatively small. A similar reasoning is also valid for the role of extreme warm or cold days. Although their relative contribution is large (i.e. the largest daily incidence occurs in both tails), these events are relatively infrequent, and therefore their contribution in absolute terms is rather small. This result is for example illustrated in the comparison between mortality projections inferred from linear and non-linear extrapolations (see Supplementary Figure 11 in article four).

Article two showed that specific changes in deseasonalized heat wave anomalies are projected to be relatively small, and thus, changes in extremes are essentially explained by a change in summer base conditions. As a rule of thumb, if the European society was not able to acclimate to the new warmer environmental conditions, and therefore the susceptibility to a given temperature was not reduced (i.e. unchanged transfer functions, see $R = 0$ in Figure 2.5b), then the increase in the magnitude, frequency and persistence of warm extremes would lead to a general rise in heat-related mortality. Instead, if the society was able to adapt immediately to the new environmental conditions (i.e. maximum shift in the transfer functions, see $R = 1$ in Figure 2.5b), the effect of summer base temperatures would remain nearly constant. In such a case, given that changes in deseasonalized heat wave anomalies are small and they can be essentially explained by a change in summer base conditions, heat-related deaths would increase, if so, to a much lower extent.

Projections of mortality exposed in article four assumed that climate is the only factor contributing to total mortality that is allowed to evolve in time. This methodology was preferred because, only in this way, we can isolate and evaluate the contribution of temperature rise. The relationship between climate and mortality was additionally allowed to change. The degree of adaptation was here represented by the acclimatization pace of the society to new warmer environmental conditions, a factor determining the horizontal shift of the transfer functions along

the temperature axis. Note that this approach is actually unprecedented and might be key for policy making, since it provides additional clues for understanding the future incidence of thermal stress. Some speculations about the range of plausible values of the coefficient of acclimatization were formulated (e.g. decreasing small values), but its evaluation was far beyond the scope of this work.

Among all the other factors contributing to mortality, two of them might push the sign of mortality trends either in one direction or the other. On the one hand, changes in profile demographics are expected to rise the relative incidence of temperatures. Indeed, Europe has today the oldest population worldwide, with a median age of nearly 40 years, and it is expected to reach 47 years in 2050. Around 60% of deaths in France were aged 75 or over during the 2003 heat wave, stressing the especial susceptibility of this age group to non-comfortable temperatures. On the other hand, changes in physiological mechanisms and behavioral habits, as well as improvements in health care and early warning systems, are certainly going to slow down or even revert the increasing mortality trends that are projected for the second half of the present century. All these factors might be evaluated through alternative scenarios in which the transfer functions are also shifted or transformed along the y-axis, but again, the description of the effect of these individual factors is far beyond the scope of this dissertation.

All these factors, namely (i) global warming, (ii) changing relationship between temperature and mortality due to temperature rise, (iii) evolving profile demographics, and (iv) improvement in health care systems and planned adaptation strategies, among many others, might have contributed to a larger or smaller extent to observed mortality trends during the twentieth century. The evaluation of the role of these factors in the past is however an extremely difficult task, not yet comprehensively addressed in the literature, and prevents us from speculating about the future trends. Article four has only described the contribution of the first factor, suggesting that the observed warming has induced a moderate decrease in continental mortality since 1950. The evaluation of factor two is far more difficult, but its effects might be assumed to be relatively small (except for summer 2003), because observed temperature trends have been relatively small compared to future projections. Hence, in those countries or regions where multidecadal records of daily mortality are available, direct comparison between simulated and observed mortality trends might give us a qualitative insight into the combined contribution of the other (non-climate related) factors. This attribution experiment could be used, in turn, to evaluate the effectiveness of improvements in health care systems, in order to design future adaptation and mitigation strategies for the second half of the century, once mean temperatures start to speed up.

The relationship between temperature and mortality was here estimated by means of a least squares fitting of equally spaced interval mean temperature and mortality data. The fitting was not applied to the original daily data because the estimation of the relationship near the tails was an essential part of the methodology, requiring thus an equally spaced sampling. Note that cubic smoothing splines, multiple regression models or weighted least-square fittings are typically used in classical impact studies (see references in article four). In all cases, the amount of data represented by each interval of temperatures is different in each segment of the distribution, and so is the range of associated uncertainty. Particularly, the role of uncertainty was here ignored, since mortality projections were essentially computed for relatively long periods of 30 years. Nonetheless, the fraction of uncertainty that is entirely associated with the fitting is an issue that needs to be addressed in forthcoming studies, so that it can be evaluated by comparing it to other sources of uncertainty (e.g. global and regional climate modeling or GHG scenarios).

This dissertation has characterized the notion of climate predictability in two particular cases. On the one hand, climate events such as EN are predictable when the distribution of states

changes to some extent after initializating a prediction scheme by means of ocean-atmosphere observations (see article one). On the other hand, the notion of predictability in climate change studies describes the sensitivity of climate statistics to the concentration of atmospheric GHG (see articles two and three). In line with this second notion of climate predictability, article four described the sensitivity of mortality projections to a non-natural external factor, namely the acclimatization pace of the society to warmer environmental conditions. Thus, in some way, this novel approach further generalizes the notion of predictability to non-climate, albeit climate-related, events. The present work might therefore enlarge our understanding of an issue of polyhedric complexity such as climate change, by providing an innovative point of view of one of its major impacts.

Chapter 4

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Chapter 5

Publications

Report of the director

The director of the doctoral thesis, doctor Xavier Rodó López, certifies the contribution of the candidate to the articles included in this dissertation, as well as their impact factor, as specified below.

Article 1.

Ballester J, Rodríguez-Arias MA, Rodó X.

A new extratropical tracer describing the role of the western Pacific in the onset of El Niño: Implications for ENSO understanding and forecasting.

Journal of Climate 24, 1425-1437, doi: 10.1175 / 2010JCLI3619.1 (2011).

Impact factor 3.363 in 2009. Ranking number 9/63 in the category *Meteorology & Atmospheric Sciences* (first quartile).

The contribution of the applicant was very important to this study. He actively discussed the experimental design, participated in the data mining and subsequent analysis process and elaborated on resulting conclusions. No other coauthor has used or plans to use any of these results for a PhD dissertation.

Article 2.

Ballester J, Rodó X, Giorgi F.

Future changes in Central Europe heat waves expected to mostly follow summer mean warming.

Climate Dynamics 35, 1191-1205, doi: 10.1007 / s00382-009-0641-5 (2010).

Impact factor 3.917 in 2009. Ranking number 5/63 in the category *Meteorology & Atmospheric Sciences* (first quartile).

The PhD applicant's contribution to this article has been absolutely principal to this study. He designed the research to perform, run the simulations and made all the necessary diagnostics to reach the conclusions obtained. Other author's contributions were much less substantial to

this study. No other coauthor has used or plans to use any of these results for a PhD dissertation.

Article 3.

Ballester J, Giorgi F, Rodó X.

Changes in European temperature extremes can be predicted from changes in PDF central statistics.

Climatic Change - Letters 98, 277-284, doi: 10.1007 / s10584-009-9758-0 (2010).

Impact factor 3.635 in 2009. Ranking number 8/63 in the category *Meteorology & Atmospheric Sciences* (first quartile).

As a continuation of the previous study, this publication in Climatic Change Letters was mainly designed by the applicant, who had a fundamental contribution in both performing the simulations and the ulterior diagnostic analyses. No other coauthor has used or plans to use any of these results for a PhD dissertation.

Article 4.

Ballester J, Robine JM, Herrmann FR, Rodó X.

Long-term projections and acclimatization scenarios of temperature-related mortality in Europe.

Nature Communications 2, 358, doi: 10.1038 / ncomms1360 (2011).

Impact factor for this magazine is due in 2012.

Joan Ballester's interests on the work on climate extremes led him to try to apply results from his former studies to the modeling of extremes impacts. In this study, he had a fundamental contribution in the design of the study, the processing of data and simulations and in the conceptualization of the important results obtained. No other coauthor has used or plans to use any of these results for a PhD dissertation.

The director of the thesis,

Xavier Rodó López

5.1 A new extratropical tracer describing the role of the western Pacific in the onset of El Niño: Implications for ENSO understanding and forecasting

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Journal of Climate 24, 1425-1437, doi: 10.1175 / 2010JCLI3619.1 (2011)

Abstract

A complex empirical orthogonal function analysis was applied to sea surface temperature data in the southern high-latitude Pacific to identify and isolate primary processes related to the onset of El Niño (EN) events. Results were compared to those of a lead-lag composite analysis of a new tracer of EN events in the southern high-latitude Pacific, the Ross-Bellingshausen (RB) dipole. Both techniques successfully isolate the main low-frequency features in the interaction among the tropical and southern extratropical Pacific during the onset of recent eastward-propagating EN events. Particularly, positive RB peaks were followed by EN events around 9 months later, on average. In turn, RB maxima were anticipated by local warm anomalies in the western tropical Pacific a year in advance, which enhance local convection and upper-troposphere divergence and generate an anomalous wave train extending eastward and poleward in the southern extratropics. In addition, circulation changes lead to a warm SST region in the central tropical Pacific, which is then strengthened by suppressed equatorial easterlies. Convection thus starts to move to the central Pacific and so the Walker circulation weakens, activating the positive Bjerknes feedback that ultimately leads to the development of an EN event. These results highlight the enormous potential of the interaction between the tropics and this high-latitude region in the Southern Hemisphere to increase El Niño-Southern Oscillation understanding and to improve the long-lead prediction skill of EN phenomenon.

5.2 Future changes in Central Europe heat waves expected to mostly follow summer mean warming

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Climate Dynamics 35, 1191-1205, doi: 10.1007 / s00382-009-0641-5 (2010)

Abstract

Daily output from the PRUDENCE ensemble of regional climate simulations for the end of the twentieth and twenty-first centuries over Europe is used to show that the increasing intensity of the most damaging summer heat waves over Central Europe is mostly due to higher base summer temperatures. In this context, base temperature is defined as the mean of the seasonal cycle component for those calendar days when regional heat waves occur and is close, albeit not identical, to the mean temperature for July-August. Although $36 - 47\%$ of future Central Europe July and August days at the end of the twenty-first century are projected to be extreme according to the present day climatology, specific changes in deseasonalized heat wave anomalies are projected to be relatively small. Instead, changes in summer base temperatures appear much larger, clearly identifiable and of the same order of magnitude as changes in the whole magnitude of heat waves. Our results bear important consequences for the predictability of central European heat wave intensity under global warming conditions.

5.3 Changes in European temperature extremes can be predicted from changes in PDF central statistics

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Climatic Change - Letters 98, 277-284, doi: 10.1007 / s10584-009-9758-0 (2010)

Abstract

Although uncertainties are still large, many potentially dangerous effects have already been identified concerning the impacts of global warming on human societies. For example, the record-breaking 2003 summer heat wave in Europe has given a glimpse of possible future European climate conditions. Here we use an ensemble of regional climate simulations for the end of the twentieth and twenty-first centuries over Europe to show that frequency, length and intensity changes in warm and cold temperature extremes can be derived to a close approximation from the knowledge of changes in three central statistics, the mean, standard deviation and skewness of the Probability Distribution Function, for which current climate models are better suited. In particular, the effect of the skewness parameter appears to be crucial, especially in the case of cold extremes, since it mostly explains the relative warming of these events compared to the whole distribution. An application of this finding is that the future impacts of extreme heat waves and cold spells on non-climatological variables (e.g., mortality) can be estimated to a first-order approximation from observed time series of daily temperature transformed in order to account for simulated changes in these three statistics.

5.4 Long-term projections and acclimatization scenarios of temperature-related mortality in Europe

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Nature Communications 2, 358, doi: 10.1038 / ncomms1360 (2011)

Abstract

The steady increase in greenhouse gas concentrations is inducing a detectable rise in global temperatures. The sensitivity of human societies to warming temperatures is, however, a transcendental question not comprehensively addressed to date. Here we show the link between temperature, humidity and daily numbers of deaths in nearly 200 European regions, which are subsequently used to infer transient projections of mortality under state-of-the-art high-resolution greenhouse gas scenario simulations. Our analyses point to a change in the seasonality of mortality, with maximum monthly incidence progressively shifting from winter to summer. The results also show that the rise in heat-related mortality will start to completely compensate the reduction of deaths from cold during the second half of the century, amounting to an average drop in human lifespan of up 3-4 months in 2070-2100. Nevertheless, projections suggest that human lifespan might indeed increase if a substantial degree of adaptation to warm temperatures takes place.

Chapter 6

Appendixes

6.1 Present-day climatology and projected changes of warm and cold days in the CNRM-CM3 global climate model

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Climate Dynamics 32, 35-54, doi: 10.1007 / s00382-008-0371-0 (2009)

Abstract

The impact of global warming on the warmest and coldest days of the annual cycle is explored according to an A2 scenario simulated by the CNRM-CM3 climate model in the framework of the IPCC AR4 intercomparison. Given the multi-model spread in IPCC projections, a validation strategy is proposed using the NCEP/NCAR reanalysis. Validation of the late twentieth century model climatology shows that warm and cold model events are slightly too long and infrequent. Although interannual trends in the warm (cold) day occurrence were positive (negative) only for six (three) of the nine considered sub-continental regions, simulated model trends are always positive (negative). This different behaviour suggests that simulated non-anthropogenic decadal variability is small relative to anthropogenic trends. Large-scale synoptic processes associated with European regional warm and cold peaks are also described and validated. Regional cold peaks are better reproduced than warm peaks, whose intensity accuracy is limited by other physical variables. Positive (negative) winter anomalies of sea and land surface temperature lead to summers with severe (weak) temperatures. These inter-annual anomalies are generated by a persistent pressure dipole over Europe. Regarding climate change, warm (cold) events will become more (less) frequent and longer (shorter). The number of warm days will largely rise and the number of cold days will dramatically decrease. The intensity of warm days will be particularly pronounced over Europe, given the projected summer drying in this region. However, according to the limited skill of the CNRM model, these results must be considered

with caution.