THE LEGACY OF TOGA: SEASONAL CLIMATE PREDICTION

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Abstract

TOGA was a ten year programme to advance understanding of the important interactions between the atmosphere and ocean which influence climate over interannual timescales. During its lifetime, a major observational system for the tropical Pacific ocean was established, with measurements of surface air temperature, surface wind, sea surface temperature, upper ocean thermal structure, ocean currents, and sea level being taken on a regular basis, and deployments of instruments to measure surface pressure, rainfall, and salinity under test. Armed with these data, oceanographers and meteorologists can not only analyse the present state of the ocean and atmosphere, but even predict climate change for several months ahead.

Before TOGA, ideas of tropical atmosphere-ocean interaction were hazy. Ten years on, still no complete theory exists, but considerable progress has been made in modelling and in understanding some aspects of that interaction. In particular the largest climate signal on interannual timescales, ENSO, (El Nino Southern Oscillation) with its origins in the tropical Pacific is better, though still inadequately, understood. There are both interannual and decadal signals in the Atlantic and Indian oceans which are related to climate variability, but the full role of these oceans in climate variability has yet to be clarified. TOGA also established only a minimum observing system in the Atlantic and Indian oceans.

Experimental climate forecasts for typically a year or more in advance are made regularly by several groups. Both physically based and statistical techniques are used, but the emphasis in this paper is on the former. Models range from intermediate complexity in which both atmospheric and oceanic components employ simplified physics, through hybrid models in which the ocean is a GCM but the atmosphere not, to comprehensive models where both components are of GCM complexity. At present the more complex models do not have predictive skill significantly better than that of intermediate models.

1 Introduction

To appreciate the legacy of TOGA, it is necessary to go back in time to see what things were like pre TOGA. The biggest ENSO on record occurred in 1982/83. One estimate of the damage caused by this ENSO is put at 9 billion dollars. Yet it was totally unforecast and even worse, largely unrecognised until it was near its peak intensity. Although sea surface temperature anomalies were analysed on a monthly basis they were not produced as quickly as they are now when analyses can be carried out for the tropical Pacific, about the remotest place on earth within hours to days and made available over the WWW. There is therefore little delay in knowing the analysis of the ocean SST or of the surface wind field.

May 82 analysis showed an SST anomally of over 1K and almost 2K. If we saw such an SST pattern now we would be alert to incipient ENSO conditions. By Oct 82 the anonally was greater than 2K over a large part of the Pacific, but it was not until the November issue of the Monthly Summary that an announcment was made that ENSO conditions were present when the then-analysed temperatures were over 5K. The problem in realising that EL Nino was active was partly that the El Chichon volcanic eruption had seriously blinded the infrared radiometric satellite radiances, partly that the SST anomalies were sufficiently large that they were deemed to be unreliable and rejected, and partly that ocean indicators did not conform to the canonical ENSO evolution of Rasmusson and Carpenter (1982).

The devastation to South American countries associated with the rise in SST was enormous. The coastal region of Peru, usually arid, received more than 3 m of rain, in places 300 times the usual amount. Bridges, roads, homes, crops, livestock, people were all washed away, some rivers carrying 1000 times their normal volume. Yet at the same time, southern Peru and Bolivia suffered severe drought. The Peruvian fishery, one of the richest in the world, was nearly eliminated as a result of changes in the ocean. But it was not just the coastal region of South America which suffered. Rainfall in the Nordeste region of Brazil was reduced to half its normal amount, while southern Brazil experienced severe flooding. (Moura, personal communication) Thousands of miles away, Australia had its worst drought since settlers arrived, with immense dust storms and devastating bush fires. The list could continue: Indonesia, China, Africa, India .. all suffered. See also Quiroz 1983 Clearly there was an obligation never to be caught out again. But how can that be done? Could this devastation have been foreseen? The need to observe the tropical oceans, to understand and model ocean-atmosphere interaction and to determine if any part of it was predictable was pressing.

The fact that the arrival of ENSO was so poorly appreciated, together with the development of some theoretical ideas of potentially important processses involved in ENSO lead to the formation of an international experiment called TOGA (Tropical Ocean Global Atmosphere). Planning for TOGA began in the early 1980's and the experiment started on 1st Jan 1985. It required massive multidiscipline interaction, bringing together oceanographers, meteorologists, theoreticians, modellers, experimentalists and engineers all devoted to the objective of understanding and predicting climate variability on timescales of a few months to a few years.

TOGA stands for Tropical ocean Global atmosphere. Why is the emphasis on the tropical ocean but the global atmosphere? Ten years ago complex models of the atmosphere showed a strong response to equatorial SST anomalies but only a weak response to extratropical SST anomalies (an anomaly is defined as the difference between the value of the SST and the multi-year average for that time of year). This result suggested that interactions between the ocean and atmosphere are strongest in the tropics. So there seemed little point in making ocean measurements outside of the tropics: it would be a monumental task to assemble enough resources to instrument just the tropical oceans. Some present-day atmospheric models show some sensitivity to extratropical SSTs but it is usually considerably less than for tropical anomalies, and there is no convincing evidence that the extratropical SST anomalies are themselves predictable. Although, ten years ago, interactions between the tropical and extratropical atmosphere were unclear, it was a reasonable hunch that they would exist and that in time they would be understood. That has largely been confirmed though even now the exact nature and extent of the interactions are still poorly understood. So dealing with the global atmosphere made sense.

There was a further reason for concentrating on the tropics. Because the Coriolis parameter, measuring the local vertical component of the earth's rotation, is zero at the equator, special dynamics apply giving rise to equatorially trapped waves, known as Kelvin and Rossby waves. The density properties of the ocean make the scales of interest relatively latitudinally confined. These waves are very large scale in the zonal direction with wavelengths of thousands of kilometers, but are trapped within a few hundred kilometres of the equator in the meridional (north-south) direction. The speed of propagation varies with their vertical structure, which in turn depends on the density stratification of the ocean. The Kelvin wave travels along the equator from west to east with speeds of between 1.3m/s and 3m/s. When it reaches an eastern boundary, it splits into two. Half the energy goes into the northern hemisphere and half into the southern. The waves then propagate poleward along the coastal boundary. Rossby waves (sometimes called planetary waves), travel westward. For short wavelengths their group velocity can be eastwards but we will not consider this aspect here. When they reach a western boundary most of their energy goes into coastal Kelvin waves which travel equatorward. (Kelvin waves always have the coast on their right in the northern hemisphere, on their left in the southern hemisphere.) They are excited primarily by the wind. Fig 5, will later show the passage of these waves in response to a wind forcing in the central equatorial Pacific. The waves of interest have a small signature in the height of the ocean surface, (and so can be detected by satellite altimetery) but their main importance is that they move the isotherms beneath the ocean surface up and down and can thus influence ocean temperature in certain regions.

The two key variables for atmosphere ocean interaction are SST and surface wind and while they are routinely measured by ships the coverage pre TOGA was poor and the data were not made available in a timely fashion. The ocean surface temperature gradients could be detected by satellite using infrared sensors. However, these did (and still do) not work in cloudy areas such as the western equatorial Pacific, which is an important region for air sea interaction, and in the early 80's such measurements were not sufficiently accurate. The only other method of observing the ocean was by ship. Research vessels are capable of making very high quality measurements, frequently to higher accuracy than really needed for TOGA purposes, but they can sample only a small part of the ocean, and can not provide the large-scale coverage needed to monitor the tropical oceans. Another strategy was needed.

Merchant ships travel certain routes quite regularly. They take measurements of surface wind and SST which are recorded in their log books. Surface wind and SST are necessary, but insufficient, measurements. The merchant ships had to be persuaded to take more measurements and a mechanism was needed to make that data available quickly. One relatively cheap system which could be mounted on a merchant ship and used while the ship was underway without the need to slow down or hold station was the XBT (eXpendable BathyThermograph). This device, fired from the side of the ship, consisted of a temperature probe connected to a long thin wire along which temperature information was relayed back to the ship. It could measure temperature from the surface to depths of around 500m. (For processes with timescales of a year or two, only the upper ocean is involved so the depth range of the XBT was perfectly adequate.) Provided sufficient spatial and temporal coverage could be obtained, the XBT could be a satisfactory instrument. The first task of TOGA was to build an XBT array. Since climate forecasts were the real aim of the TOGA programme, the data had to be available in near real time. New technology came at around the same time allowing fixed moorings to survive in up to 5000m of water and measure temperature at various depths from the surface to 500m. In addition they measure the surface wind and the atmospheric humidity. Tests have been made of their ability to carry automatic rain guages and salinity sensors. The buoys relate their data back via satellite in near real time: for the winds typically three times a day, for the subsurface daily averages once per day. The observing array (called the TOGA TAO array) has been built up over the years. There are currently about 60 moorings deployed in the equatorial strip. Fig 1 shows the current observing system for the tropical Pacific.

De facto, ENSO was the largest climate signal on interannual timescales so most of the effort, both modelling and observational, went to the Pacific. I will therefore describe ENSO first in section 2, followed by modelling and prediction work in sections 3 and 4. The importance of the other oceans and of tropical-extratropical interactions will also be discussed in section 4. The reader who wants to explore many of the issues in greater depth than can be covered here is referred to Glantz et al (1991) for a review of ENSO observations, to McCreary and Anderson (1991) and Neelin et al (1994) for reviews of modelling and theoretical work relevant to ENSO, to Gill (1982) for a thorough treatment of atmospheric and oceanic dynamics, to Philander (1990) for an account of equatorial dynamics and an enjoyable history of El Nino, and to Latif et al (1993) and Palmer and Anderson (1994) for reviews of seasonal forecasting.

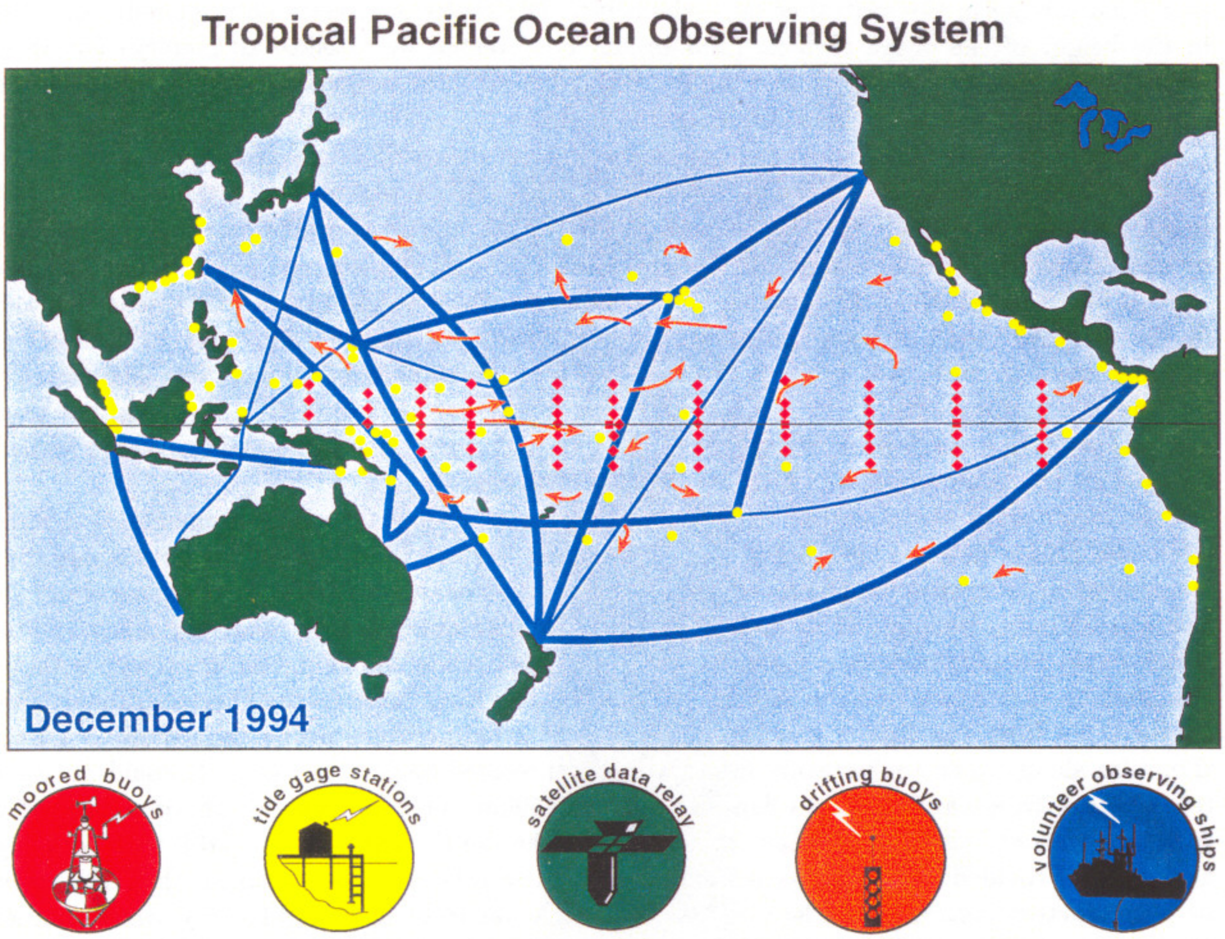


Fig 1 The TAO observing array, consisting of bottom anchored moorings in 5000m of water, (diamonds). These moorings can measure the temperature of the top 500 m of the ocean and relay the data via satellite back to shore in near real time. In addition they measure the surface wind and the atmospheric humidity. Tests have been made of their ability to carry automatic rain guages and salinity sensors. The lines on the figure indicate the XBT network. Merchant ships ply these routes and measure temperature of the upper few hundred meters of ocean using XBTs. Much of this data too is relayed to shore via satellite, but if ships do not have the instrumentation to do this, the data are recorded on board and collected at the end of a voyage. During TOGA most ships have been converted to return data via satellite. Also shown by arropws are some drifters. These buoys, developed for TOGA are both light weight for ease of deployment and relatively cheap making it possible to deploy a fair number of them (typically two hundred). They are deployed throughout the tropical Pacific, and increasingly in the other oceans.They are drogued at a depth of 10m. By measuring their position regularly via satellite, their velocity can be determined. The buoys have been designed to follow the water so their velocity gives a measure of the surface currents. In addition they measure surface temperature and can carry atmospheric surface pressure sensors. The comprehensive nature of this observing system was only achieved towards the end of TOGA. It took 10 years to establish. From McPhaden 1994

A combination of XBT lines, drifters and tide guages has been implimented for the Atlantic and Indian oceans, also, but it is much less well developed than for the Pacific (not shown).

2 A DESCRIPTION OF ENSO

As early as the beginning of this century, it was realised that there was a large scale readjustment of mass in the atmosphere between the west and east Pacific. When atmospheric pressure is anomalously high in the Indonesian region it is anomalously low in the tropical east Pacific. This was called the Southern Oscillation (SO) by Gilbert Walker, who was motivated to find predictors for the Indian monsoon rainfall. He failed but he found the SO instead. (Posterity should not judge Walker harshly for failing to find predictors of the monsoon. The role of the monsoon in interannual climate variability is still poorly understood.) A measure of the SO is now usually defined as the pressure difference between Tahiti and Darwin. Fig 2 (solid line) shows the surface pressure anomaly (i.e. the difference between the measured value and the multiyear average for that time of year) at Darwin in northern Australia from 1882. The pressure record is noisy so it has been smoothed by repeatedly applying a 5 month running average. There is clear evidence of variability on timescales of a few years between occurrences of high pressure and similarly for low pressure anomalies.

Independently, it was realised that there was an irregular behaviour in SST and currents off the coast of Ecuador and Peru, with attendant disruption of the biology. This is sometimes called El Nino. The history of El Nino is a long one going back to the 1890's but it was not until Bjerknes (1966) that it was realised that both the El Nino and the Southern Oscillation were manifestations of the same coupled atmosphere ocean interaction. (The use of the name El Nino (the Child) has become a little confused. It was originally a name used by fishermen of Paita in Peru to denote a warm current from the Gulf of Guayaquil which arrived around Christmas time. Later, it was associated with the warming resulting from interannual variability which had a tendency to occur in Spring, and now El Nino is used to indicate the basin wide changes in the ocean, which can last for up to a year. I prefer and will use the acronym ENSO to emphasise that there is a strong connection between the El Nino (EN) as measured by SST and the SO.) Fig 1 also shows SST anomalies in the central east equatorial Pacific (dashed line). Clearly this is strongly tied to Darwin pressure on interannual timescales. What sets the time scale of a few years between events will be considered in section 3.

Fig 3 shows a cartoon of normal conditions in the equatorial Pacific, and of the warm phase of ENSO (i.e. when the tropical central east-Pacific is warm). Under normal conditions, SST is highest in the west and atmospheric convection (and rainfall) is greatest over the warmest waters. The surface wind is from east to west. During a warm event, the warm waters extend eastward and the maximum in SST may in fact move close to the dateline. Convection moves eastward with the warm water giving enhanced precipitation in the central Pacific, but reduced precipitation in the Indonesian region. The westward winds reduce over and to the west of the SST anomaly, and in strong events, may even completely reverse and become eastward.

Although ENSO is centered on the tropical Pacific it influences climate as far afield as India and the Sahel and on occasion possibly even Europe. Fig 4a, from Ropelewski and Halpert (1987) shows some of the precipitation anomalies associated with the warm phase of ENSO, while panel b (from Halpert and Ropelewski 1992) shows regions in which surface temperature is correlated with the warm phase of ENSO. Within the tropics the signals are almost reversed between warm and cold events (not shown, but see Halpert and Ropelewski). Fig 4 shows some connection between tropical Pacific SSTs and north American climate. There is also a connection to European climate which shows up more clearly in the cold phase of ENSO. Over Europe, the signal represents only a small fraction of the interannual variance, however.

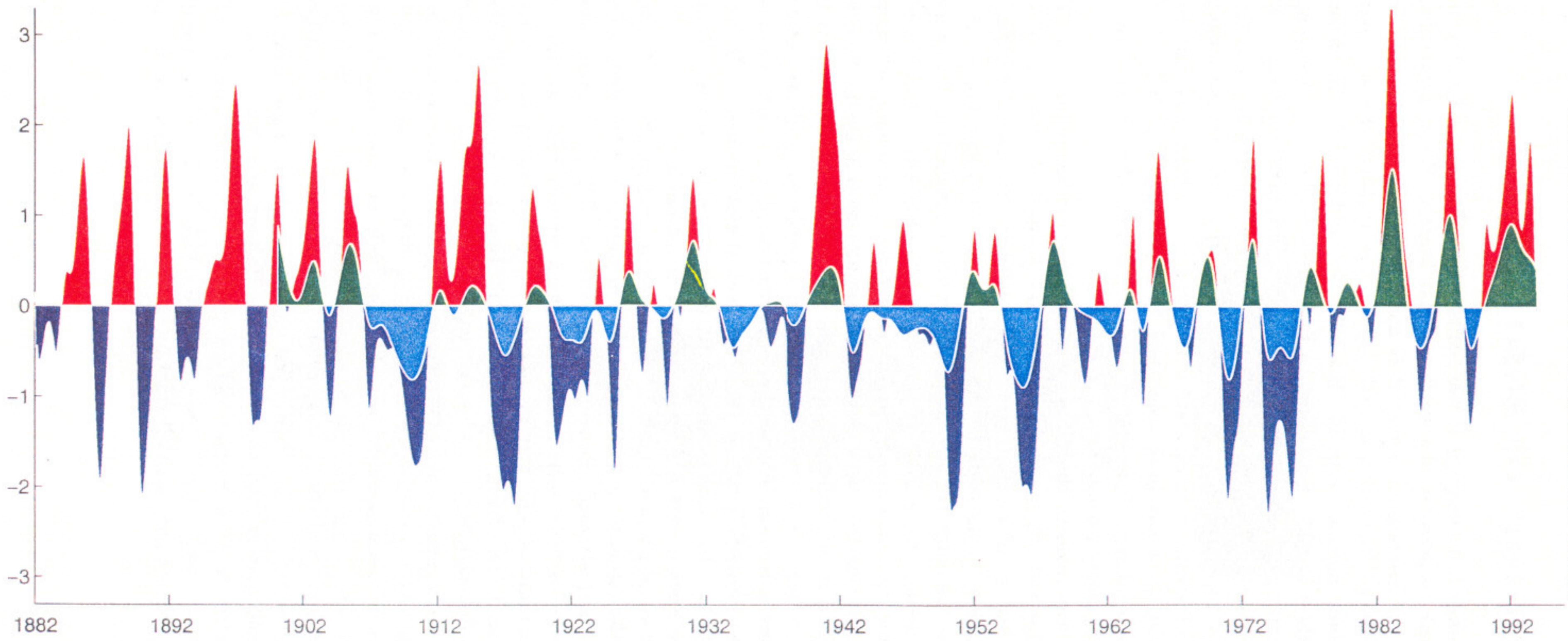


Figure 2 Time history of the surface pressure at Darwin, Northern Australia from 1882, and of SST anomalies in the central eastern Pacific from 1900 showing that the warm events in SST and high pressure at Darwin are strongly correlated on interannual timescales. The time between one warm event and another is variable. It is usually 3 to 4 years but on occasion can be as long as 8 years.

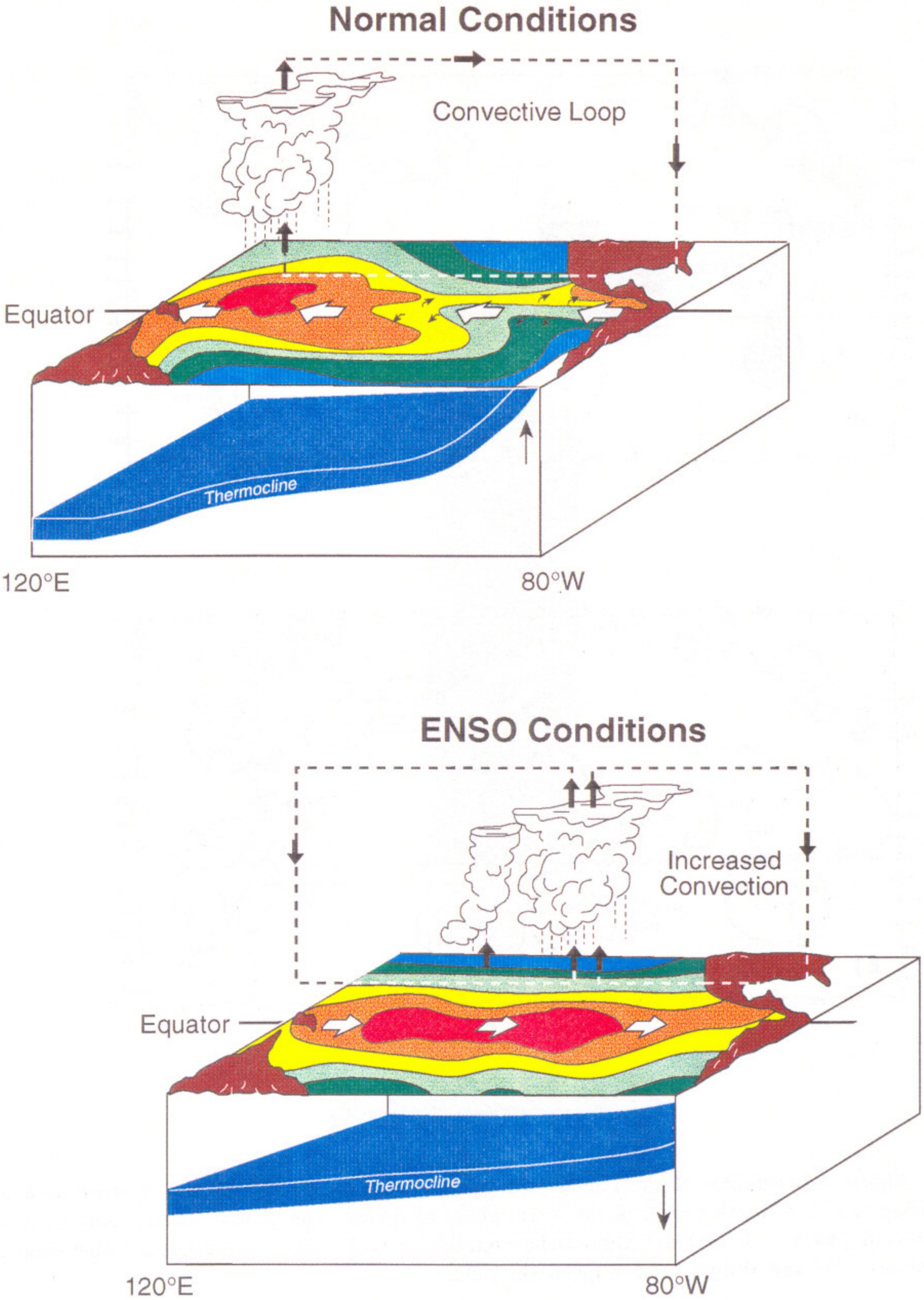


Fig 3 Schematic of normal and warm conditions in the Pacific. Convection is displaced from the Indonesian region of the west equatorial Pacific to the central equatorial Pacific. The circulation in the vertical plane along the equator is known as the Walker circulation. The usually westward winds relax and may even become eastward leading to a deepening of the thermocline in the east and a raising of the thermocline in the west. See fig 5 for a description of thermocline, a region of high vertical temperature gradient. Fig provided by M McPhaden, Director TAO project Office, NOAA/PMEL

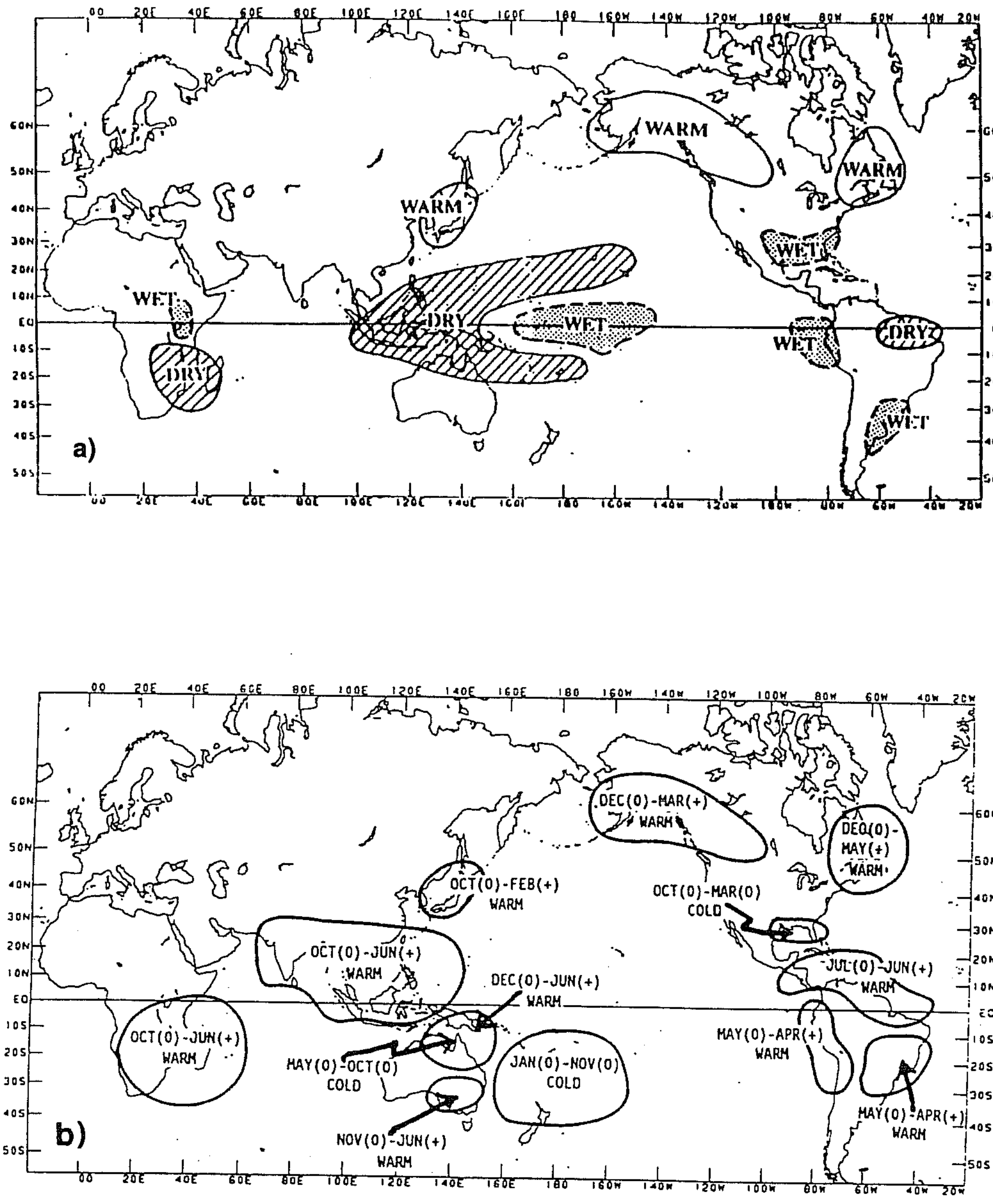


Figure 4 Schematic of the regions which experience a) rainfall b) temperature anomalies during the Nov-March period of the warm phase of ENSO. The climate signals associated with the cold phase of ENSO are almost the opposite of those shown. Based on Ropelewski and Halpert 1987 and Halpert and Ropelewski 1992.

3 MODELLING ENSO

3.1 Some theoretical ideas

A plausible connection between variations in SST in the equatorial Pacific and the Southern Oscillation was made by Bjerknes (1966) who argued that the East-West (often called the zonal) temperature gradient along the equator was related to winds along the equator. When eastern Pacific SST is cold, there is a strong thermal gradient along the equator. The warm water in the west is associated with rising air motion in the atmosphere and the cold water in the east with descending air. Since the Coriolis parameter in the equatorial zone is small, the low-level flow will be largely down the pressure gradient (rather than across it as at mid-latitudes) giving rise to a zonal circulation that Bjerknes called the Walker cell (see fig 2). The SST in the eastern Pacific is cold when the thermocline is close to the surface. The thermocline along the equator slopes from a depth of 150m in the west to near the surface in the east as shown in fig 5, the slope being roughly proportional to the wind stress. The stronger the westward winds the more the thermocline slopes up to the east, and vice versa. Weakening the zonal thermal gradient leads to a reduced atmospheric pressure difference between the east and west equatorial Pacific and a weakening of the winds. If the winds relax, less cold water is brought to the surface and the temperature rises reducing the winds further: i.e. there is a positive feedback mechanism. The feedback also works in reverse- if the westward winds strengthen, more cold water is brought to the surface and the temperature in the east cools leading to stronger westward winds.

Bjerknes' notion is clearly incomplete since it gives no indication of a cycle between warm and cold events, but it still lies at the heart of the ENSO mechanism as it represents the mechanism whereby small perturbations can grow. A missing ingredient is that the ocean can also adjust remotely to changes in the winds. The adjustment is accomplished by the Kelvin and Rossby waves mentioned in section 1, which can propagate long distances in the ocean without crippling attenuation. Typical speeds for the free waves are from 3 to 0.2 m/s, corresponding to crossing times of the Pacific basin of between 50 and 1000 days. A plausible mechanism for obtaining the 3–5 year time scale of ENSO could then involve ocean wave processes. This idea was embodied in the earliest numerical models of ENSO in which the time for waves to transit round the basin set the time-scale of ENSO (see McCreary and Anderson 1984, or McCreary and Anderson 1991).

The role of the Kelvin and Rossby waves is illustrated in fig 6 which shows the depth of the thermocline in a theoretical rectangular ocean basin. An eastward wind is applied to an ocean at rest with a flat thermocline. For clarity, the wind is located in the middle of the ocean basin which is 10,000 km wide, and extends from the equator north to 4500km. (One degree of latitude corresponds roughly to 110 km). The spatial extent of the applied wind stress is shown by the rectangular box on the first panel, and the wind profiles in the zonal and meridional directions are indicated by the curved lines. The initial response is for the thermocline along the equator to tilt. In response to an eastward wind the tilting is downwards in the east, and upwards in the west. But the tilting does not remain localised to the forcing region. Even after one month the 'downwelling' signal has travelled substantially to the east via an equatorial Kelvin wave (shaded region). After three months this wave has collided with the eastern boundary and started to reflect energy back into the interior, as a Rossby wave. This is interesting enough but events to the west of the forcing are also of relevance. The wind excites a Rossby wave which is just visible after 1 month (panel a) but clearly so by 3 months (panel b unshaded contours). This wave has its maximum off the equator and propagates westward. After 13 months it has collided with the western boundary. The energy is then channelled into a Kelvin wave and reflected back along the equator.

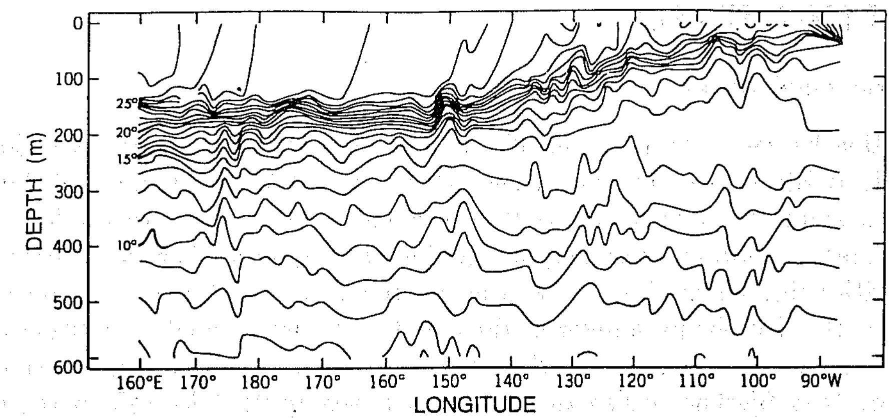


Figure 5 Ocean temperature as a function of depth along the equatorial Pacific. From Colin et al 1971. The region of strong vertical temperature gradient is known as the thermocline. It rises from a depth of about 150m in the west Pacific to near the surface in the east. The slope is a consequence of the surface westward wind. From Pacanowski and Philander 1981, based on Colin et al 1971.

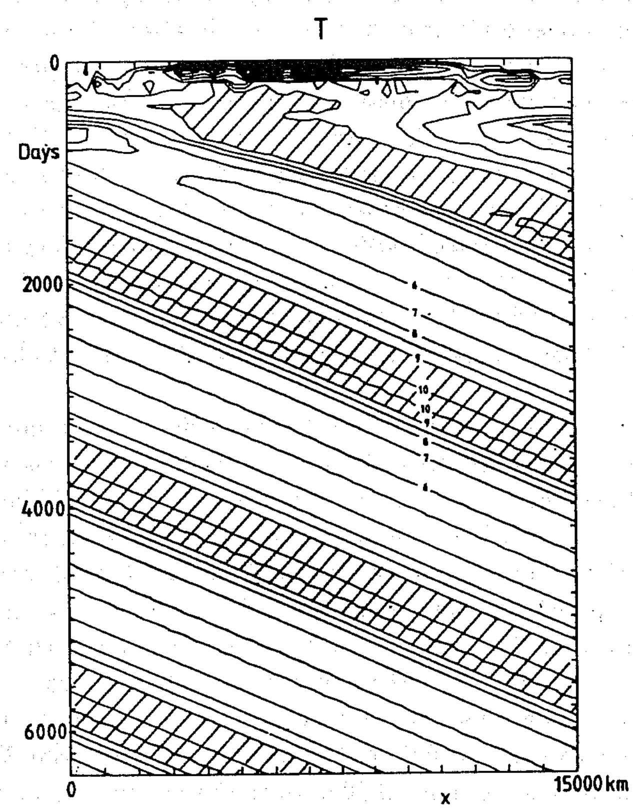


Fig 7 Plot of sea surface temperature in a coupled model as a function of time. As a result of instabilities in the coupled system, a finite amplitude disturbance developes and drifts slowly eastward. The model ocean has no boundaries, so the mechanism here is different from the delay oscillator mechanism which requires energy to be reflected back from a western boundary. From Anderson and McCreary 1995. Hirst 1987 has examined the origin of this and other local instabilities in a theoretical frmework.

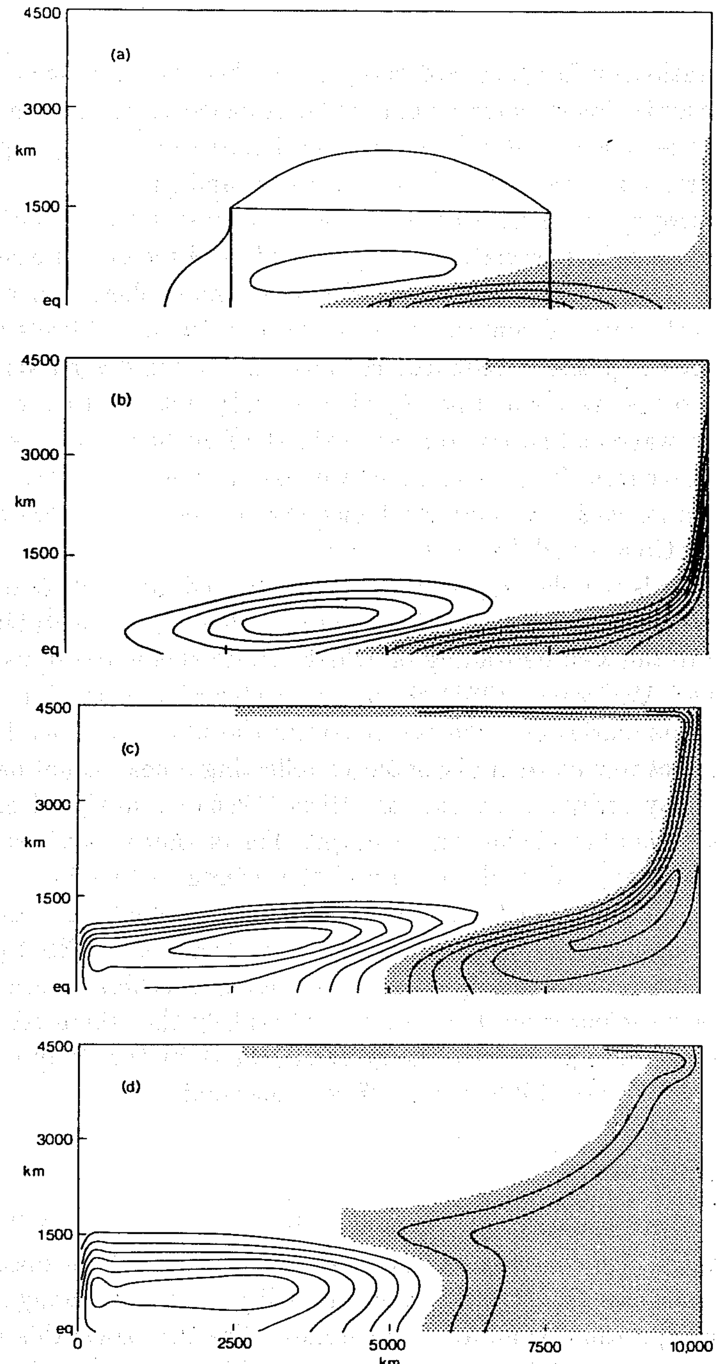


Figure 6 The depth of the thermocline in a numerical model after the application of a steady wind stress. The wind is confined to the central equatorial region and is zero outside the box region shown in panel a. It is maximum on the equator at 5000km and drops off poleward and with longtitude. The wind is in the zonal direction. Only the northern hemisphere is shown, but the wind and the ocean response are symmetric across the equator.

The thermocline is shown after (a) 1 month , (b) 3 months (c) 13 months (d) 60 months. Panel (a) shows the thermocline deepening to the east of the forced region, by Kelvin waves. By (b) these waves have propagated along the eastern boundary and by (c) Rossby waves have started to carry the signal away from the eastern boundary into the ocean interior. To the west of the forced region Rossby waves have also been excited, but they are of opposite sign to the Kelvin wave and eastern boundary excited Rossby waves. The Rossby waves excited directly by the forcing travel to the western boundary (b and c) where they reflect as Kelvin waves. From McCreary and Anderson 1984.

If the wind is stationary in space and time, as in this example, then after 13 months the solution on the equator is close to equilibrium. It takes longer to achieve an equilibrium state off the equator because the planetary Rossby waves travel more slowly. The slope to the thermocline in the region of the forcing is evident in both panels (c) and (d).

If now the wave propagation idea is combined with the instability idea of Bjerknes then a 'delay oscillator' scenario can be postulated. In the wind forced region, an eastward wind anomaly tilts the thermocline down to the east, giving rise to warming there and weaker cooling to the west. This decreases the thermal contrast between east and west and leads to a strengthening of the eastward anomaly i.e. positive feedback. However an eastward wind also leads to shallowing of the thermocline to the west and this signal eventually returns to the forcing region via a Rossby and a Kelvin wave and starts to shut down the feed back and even reverse it, starting the cool phase. This scenario is appealing and various conceptual models have been developed which illustrate it rather well, the simplest being that of Suarez and Schopf (1987) but other forms are given in McCreary and Anderson (1991).

The delay oscillator is not the only way by which coupled atmosphere ocean oscillations can occur. There can be local instabilities of the coupled system. These instabilities may propagate either to the east or to the west depending on which aspect of equatorial waves is dominant. Fig 7, from Anderson and McCreary (1985) shows the eastward propagation of an SST anomally in a ocean with no boundaries i.e. the ocean covers the planet just as does the atmosphere. This instability does not rely on ocean boundaries reflecting a heat signal back from the western boundary as in the delay oscillator mechanism. Hirst (1986) has analysed several of these 'local' instabilities, of which the above is but one example. For further analysis see the review articles of McCreary and Anderson, and Neelin et al and the references therein.

It is quite likely that the predictability of instabilities of this type is different to that associated with the slow propagation of waves, so implying that some ENSO events are harder to predict than others. At a guess one might expect the local instability mechanism to be less predictable than the delay oscillator mechanism. It is also likely that the predictability will depend on the phase of ENSO, as suggested by Davey et al (1995) for the delay oscillator mechanism, and by Penland and Magorian (1993) for a different mechanism.

3.2 Model validation

The best validation of atmospheric models is done by prescribing the SST field and simulating interannual variations, which can then be validated against atmospheric observations. In general the models simulate the large scale tropical flow anomalies fairly well, though there are significant errors in terms of precipitation, wind and heat fluxes. (See Brankovic this volume)

A similar validation is possible for ocean models. The forcing required to drive an ocean model is not SST but the surface wind (actually the wind stress) and the heat and fresh water fluxes. None of these quantities is known to sufficient accuracy, especially the latter two. Validating ocean models is therefore much harder than validating atmospheric models as errors in the modelled SST field can result as much from errors in the forcing fields as from errors in the ocean model, and partitioning the error between forcing and model is hard. In fact the single greatest impediment to developing accurate ocean models has been the lack of accurate forcing fields.

TOGA has given high importance to getting the data needed to improve forcing fields. We know the wind field better now but even so still not to sufficient accuracy. This is partly a scientific issue, partly a practical one. The practical aspect is one of getting enough measurements of high accuracy from this remote part of the world. The scientific issue relates to how momentum is transmitted to the ocean. The first effect of wind blowing over water is to excite surface waves of wavelengths ranging from cms to several tens of meters. If these waves tramsmitted their momentum locally into ocean currents then the forcing would be local and direct. Unfortunately, swell waves (wavelengths of tens of meters) can travel long distances and influence the transfer of momentum in non local ways. This has not really been a major issue to date as errors in measuring the surface wind field have dominated, but might become more of an issue in the future as knowledge of the wind field improves. The wind stress is usually analysed assuming the ocean is at rest. In fact, in parts of the tropics current speeds may be non negligible compared to the wind speed suggesting this correction should be taken into account at some stage.

Obtaining reliable measurements of the heat flux is difficult as it is a residual of rather large quantities. The net heat flux, i.e. the difference between the incoming solar short wave radiation and the outgoing long wave radiation, the evaporative flux and the sensible heat flux is small, in places such as the west equatorial Pacific, as small as $10W/m^2$ whereas some of the components might be $200 - 300W/m^2$. A second difficulty is that the formulae used to calculate the evaporative and sensible fluxes are validated by ships operating on small scales (a few square metres), whereas what modellers need are values appropriate to an area of 10^{10} square metres, so there is a huge scaling issue.

Errors in wind stress can give rise to errors in currents and errors in thermocline slope which in turn give rise to errors in the thermal field and SST. So not only is it difficult to separate model and forcing error, it can be difficult to separate wind and heat flux errors. Errors in the wind will lead to errors in the currents, but the model will still reach an equilibrium none the less: the current error will not just keep growing. But for heat flux error there is no equilibrium: if it is too high the ocean will just continue to heat up. Any practical use of heat flux forcing therefore has to include a negative feedback. While this is required to control error there is, fortunately, some justification for it. Different models address this in different ways: one approach is given in Balmaseda et al (1994).

Ocean models forced over the last three decades with the 'observed' wind, together with some estimate of the heat flux and a negative feedback on temperature give suprisingly realistic results, given the many uncertainties in the forcing fields. Significant errors still remain, however. These differ to some extent from model to model suggesting that at least part of the error results from model deficiencies but a significant part is undoubtedly a result of forcing error. The reanalysis project at ECMWF should help to provide more reliable forcing fields, but there is no substitute for missing data. The reality is that the observing system was inadequate until the 90s so large uncertainties remain in the modelling of the ocean state.

4 SEASONAL AND LONGER CLIMATE FORECASTS

4.1 ENSO forecasts

The wave-oscillator idea embodied in fig 6 suggests that ENSO events should be exactly repeating and therefore infinitely predictable, but this is clearly not so. As fig 1 shows there is a considerable spread in the time between events. The differences could arise from inherent non-linearity in the oscillator reminiscent of chaotic behaviour as expressed by Lorenz (1963); from variability in the atmosphere such as the intraseasonal oscillation; or from interaction between ENSO and other processes causing climate variability such as the annual cycle, the quasi-biennial oscillation, changes in snow cover over the Eurasian land mass, or variability outside the tropical region.

In spite of these diverse and uncertain aspects, one might conjecture that if one could somehow determine the state of the coupled system in general, but especially the state of the ocean, it might be possible to predict subsequent events. How can one obtain an analysis of the ocean? One might think that the problem is similar to that for the atmosphere. You go and measure as much as you can and then assimilate the data into a model using 4D data assimilation. Unfortunately that is not possible. There are almost no subsurface data to assimilate for the pre Toga period. Towards the end of TOGA there is quite a lot of useful data though still not enough to determine the ocean state outside of the band ± 8 degrees of the equator.

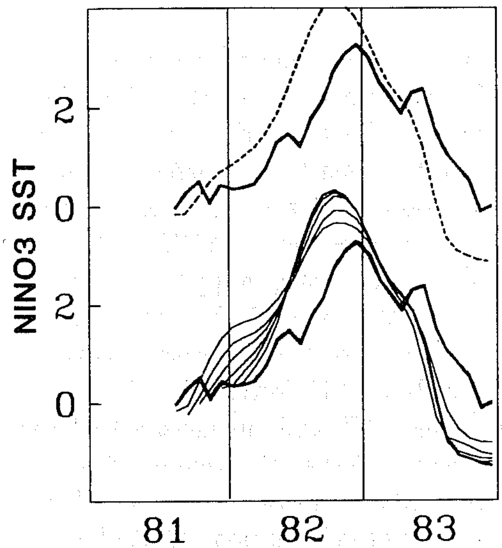
So how can we obtain an ocean analysis from which to launch a hindcast for periods in the 70s or 80s? (This is a necessary activity in order to validate model skill). Fortunately the tropical ocean is a forced deterministic system, which in this context means: given the past history of the forcing, say the last few years, the ocean comes into adjustment with that forcing. Memory of the ocean initial conditions is lost; only memery of the forcing remains. This is because the ocean is not governed by internal instabilities, as is the case in the atmosphere. Unfortunately this puts great importance on the quality of the forcing, which we do not know adequately well. The hindcasts/forecasts must be degraded as a result. We will return to this point later.

Cane et al. (1986) were the first to attempt climate forecasts. In each case an ensemble of six forecasts was made, each starting from a different consecutive month. These forecasts can sometimes differ considerably showing a strong sensitivity to initial conditions, so it has been found advantageous to average forecasts over the initial conditions to give a consensus forecast. The hindcasts for the 82/3 ENSO were spectacularly good, especially the ensemble average, which is shown in fig 8. [This result is even more impressive given that most models have had difficulties in hindcasting this event and simulating it has been an equally difficult task.] Since then other models have been developed and forecasts from several models are now made on a regular basis and published in the Climate Bulletin, and the Experimental Long-lead Forecast Bulletins produced by the National Meteorological Center of the United States.

There are only a relatively few ENSO events for which we have adequate data to initialise models. Assessing the skill of a model is therefore rather difficult and problematical since it is trained and tested on only a few realisations and frequently verified against only one parameter, viz SST in the central eastern equatorial Pacific. Fig 9 from Balmaseda et al (1994) shows the correlation between observed and predicted SST for the years from 1969 to 1991. Ideally one would like to use a much longer test period but regretably wind fields over the Pacific ocean extend back only to 1961, and even then are probably very inaccurate. The skill is seasonaly variable, with forecasts for the Northern Hemisphere spring period tending to be least skillful. This may be a consequence of the stability of the mean state changing seasonally (Webster and Yang 1992) but other explanations are possible. Balmaseda et al (1995) argue that the seasonality in skill is not evident in the decade of the 1980's while in the 1970's the low skill in spring is because the signal is very small then, consistent with Xue et al (1994). Whether the contrast between the 70's and the 80's reflects some significant climate change is unclear, but there is other evidence suggesting a rapid regime change in climate at the end of the 70s.

The model skill appears to be greatest in predicting major events and almost zero in predicting the smaller month to month variability in SST. This is particulary evident if forecasts are made of only major events during the above period, when the prediction skill correlation may be as high as 0.8 after 15 months.(See Latif et al 1993) Nonetheless, fig 9 shows that on average the correlation is above 0.5 out to about 9 months. So there may be some useful skill in such models.

Both the Cane et al and Balmaseda et al models are anomaly models in that only anomalies of SST are used by the atmospheric model to calculate wind anomalies. These wind anomalies are



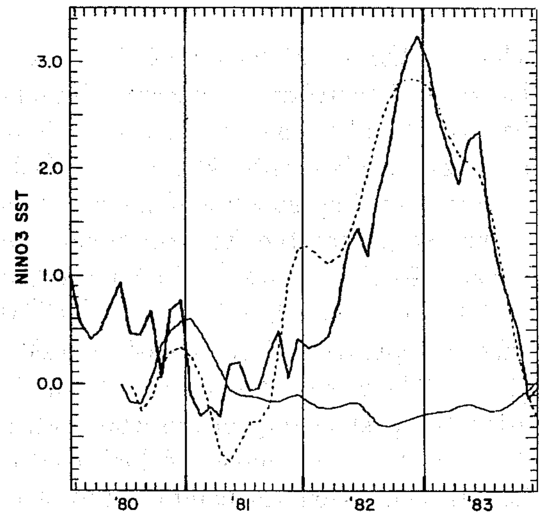


Fig 8 Left panel: Plot of hindcast SST for the 1982/3 ENSO. Lower curve shows 6 individual hindcasts initiated at 1-month intervals. The dashed curve in the upper panel shows the ensemble average. The heavy curve is the observed temperature anomally in the region Nino 3. Right panel: Plot of two hindcasts initialised only one month apart, showing dramatic differences in their evolution. The coupled system is apparently extremely sensitive at this time. From Cane et al 1986.

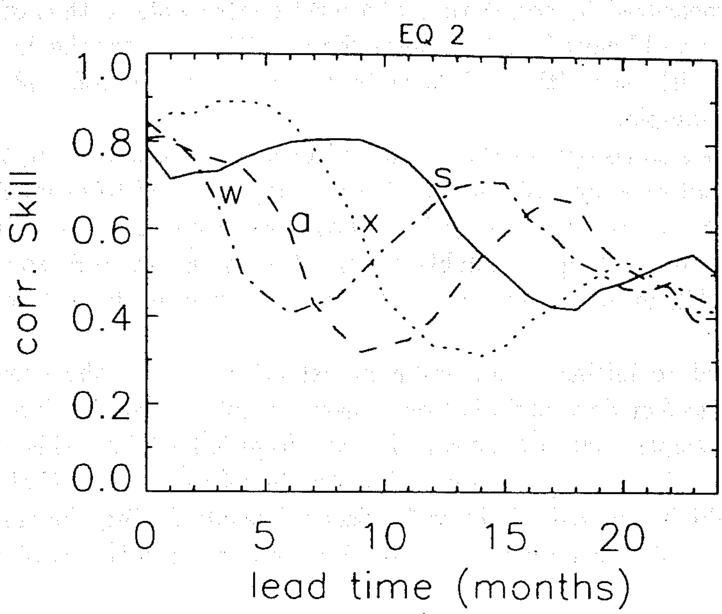


Fig 9 Correlation skill between model predicted temperature for the region Nino3 and observed SST as a function of lead time of the forecast. The four curves are for forecasts initiated at different times of year (Spring (s), summer (x), autumn (a), winter (w).). The seasonality in the predictive skill could indicate a sensitivity of the coupled system to time of year, with the spring being least predictable. All curves loose skill as they pass through spring. For a fuller discussion see Latif et al 1993. From Balmaseda et al 1994a

added to the climatological wind to force the ocean model, so the ocean model responds to full forcing and therefore can correctly simulate advective and other nonlinear processes. However, because the atmospheric component is only an anomally model, climate drift is not a major problem. Models with physics simplified along these lines have proved to be especially useful both in providing understanding of how ENSO works and in making forecasts. The expectation is, however, that the fully comprehensive coupled general circulation models (CGCM's) will surpass the skill shown by the intermediate models, but that has not clearly happened to date.

The experience in using coupled GCMs in forecast mode is rather limited. To date there is only one operationally produced forecast using a coupled GCM, that of Ji Kumar and Leetmaa (1994). The ocean state from which the hindcast is launched is derived by initializing the model using the past history of the wind but, in addition, assimilating SST and subsurface thermal data into the ocean model. A high level of dynamical consistency between the analysis and the forecast components is maintained to ensure that the analyzed ocean initial states do not undergo a sudden change at the beginning of the forecast, leading to loss of information. Considerable effort was expended to tune the physical parametrization schemes in order to obtain sufficiently strong winds to maintain the appropriate ocean subsurface structure during the coupled model forecasts. Even so, the model is not fully coupled as some intervention to the atmospheric momentum and solar heat fluxes is made in order to reduce climate drift. Results to date are limited to a smallish set of hindcasts and forecasts for periods during the years 1984-93 in which there were three realisations of El Nino, but appear encouraging. Fig. 10 shows that for July initial conditions, skill scores, for December, January February predictions, of over 0.9 prevail over most of the central and eastern Pacific which are better than the skill scores obtained with a persistence forecast.

Other attempts have been made to assimilate data. Rosatti et al (1995) have performed hindcasts from analyses obtained with and without data assimilation. The latter are worse than the hindcasts made using intermediate models, but the former is a significant improvement, bringing the skill as measured by correlation to a level comparable to that of the intermediate models discussed above and below (Fig 11). Since there is little data for the 70s, it is not possible to compare hindcasts with and without data assimilation for this period and so the results are perforce for a limited sample.

Kleeman et al have also compared the impact of data assimilation, using their intermediate model. The data are not directly assimilated. Rather they are used to produce an analysis and then the mean temperature over the range 0 to 500m used to adjust the depth of the model's single active level. Results are only available for the 80s for the same reason as for Rosatti et al., but a significant skill improvement is noted relative to the no data assimilation case (Fig 12).

The technique used to initialise the ocean model is to spin-up the ocean with observed winds, regardless of whether data assimilation is used or not. When the forecast is started an atmospheric model is coupled onto the ocean. The winds produced from this model are unlikely to match those used in the spin-up, so some shock is inevitable. Chen et al (1995) sought to minimise this by combining model winds with observed winds during the spin-up phase. The resulting hindcasts were significanmtly improved for the 80s, though less obviously so for the 70s as seen in fig 13.

A somewhat different strategy has been adopted by Yun et al. The FSU winds are undoubtedly noisy, partly resulting from inadequate sampling. Some of the small scale features are probably unphysical. Could these winds be filtered to advantage? It appears that they can. The strategy is to derive a statistical relation between observed SST and observed winds by decomposing into EOFs. Only the lowest few EOFs of SST are retained and from these the

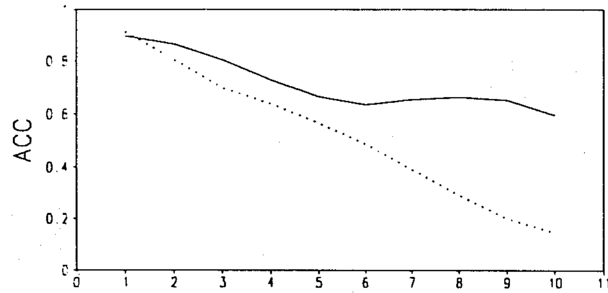


Fig 10 Forecast skill for the Nino-3 region as a function of hindcast lead time (solid curve). The skill of persistence as a predictor is shown by the dotted curve. The skill is based on forecasts for the period 1984 to 1993. From Ji et al 1994.

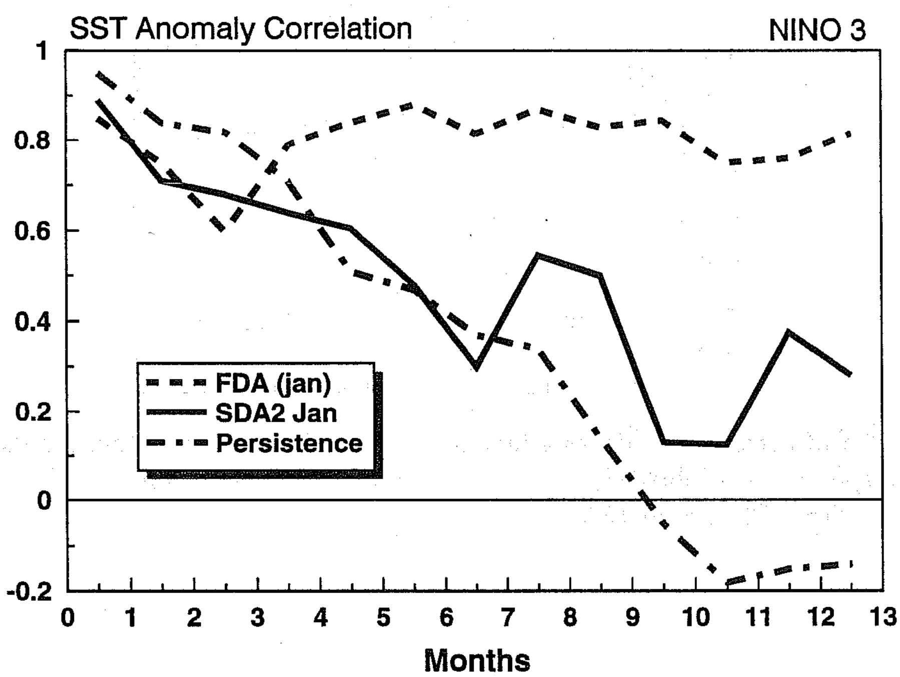


Fig 11 Plot of correlation for the nino3 region as a function of lead time for two experiments (FDA) -Full data assimilation and SDA2, an experiment in which data are not assimilated although the ocean surface temperature is nudged towards observed SST. From Rosati et al 1995.

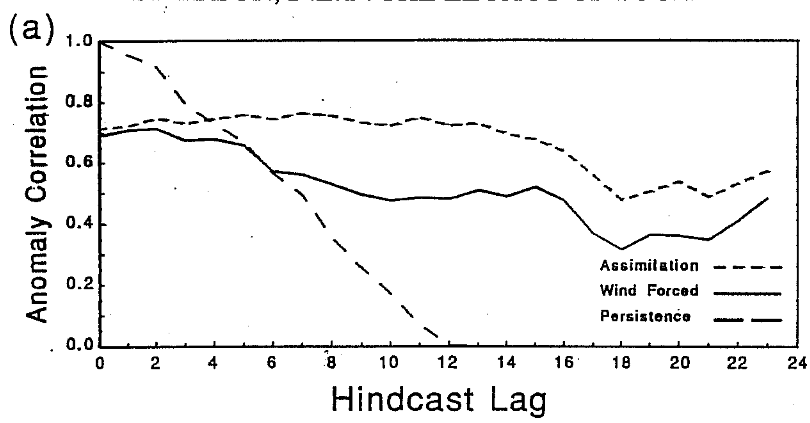


Fig 12 Plot of correlation skill as a function of lead time for two experiments, with and without data assimilation. Persistence is also shown. The model is a one active layer ocean coupled to an equilibrium atmosphere. Data are not directly assimilated into such a simple model. Rather a multi-level ocean analysis is made using all data. Anomalies of upper ocean heat content are then used to correct the layer ocean depth using a 4D var system. From Kleeman et al 1995

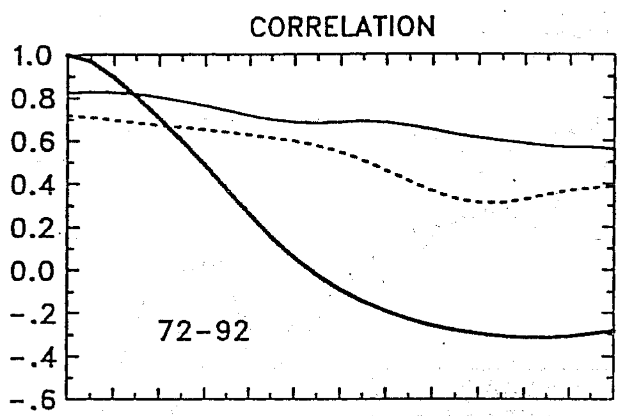


Fig 13 Plot of correlation skill as a function of lead time for two experiments, the standard wind forced spin-up case (dashed) and the revised spin-up using a combination of observed and model winds. From Chen et al 1995.

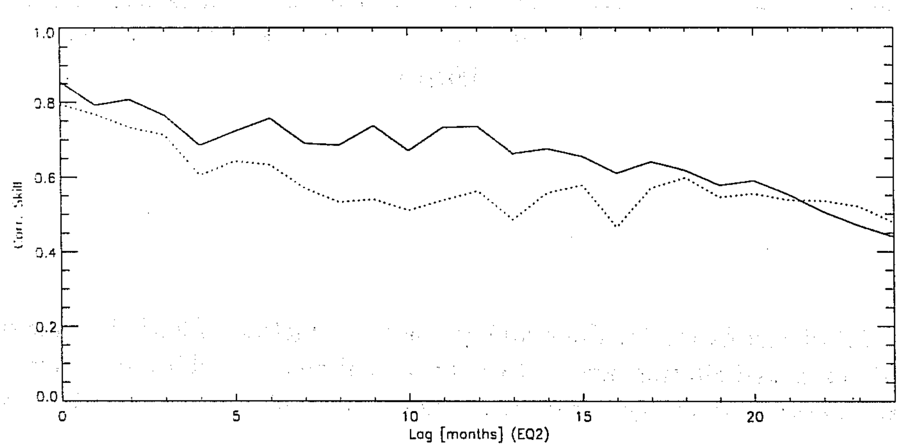


Fig 14 Plot of correlation skill as a function of lead time for two experiments, the standard wind-forced spin-up (dotted curve) and a revised spin-up in which the winds are filtered using a statistically derived relationship between observed SST and observed winds (solid curve). From Fan Yun, personal communication.

associated wind patterns. Off the equator this statistical relation is less good than close to it, where the atmospheric response is less chaotic. So the wind stress used to force the model is a combination of the observed and the SST filtered version, being primarily the filtered version near the equator but the observed at higher latitudes. Fig 14 shows the correlation skill for the 70s and 80s using an ocean model initialised in this way. The coupling with the atmosphere is made seasonally dependent in these integrations.

A recent review of 5 different models used for ENSO prediction is given by Barnston (1994): two of the models are totally empirical, while three are physically based. He finds that all the models have comparable skill scores, but there is a significant difference in their behaviour for any given prediction. The correlation between models is rather low. The interpretation of this result is unclear. If it is because the models have really little coherence then it is a pessimistic result. On the other hand if it is because there are sytematic differences between models - for example one does well for the onset phase, while another does better for the cooling phase then this information should be useable in any super-emsemble spanning different models. The jury is still out, however.

Although the skill scores of many models look impressive, none of the models did well at predicting the 90s. In many cases the ability to even simulate the SST changes in the 90s is poor. What about predictions for next year? A series of predictions is available in the Experimental Long-lead Forecast Bulletin produced by the Climate Prediction Centre at NCEP. These may be summarised as 2 models are predicting very warm conditions in 1996, and several are predicting normal. Several models have been revising their predictions upwards over the last few months, giving an overall prediction of above average SSTs in Nino 3 in 1996.

But what of predictions for the extratropics? Can we make climate forecasts for Europe? The answer is quite uncertain. Empirical and model studies (e.g. Livezey, 1990) indicate that some useful skill for forecasts out to a season exists during large ENSO events. Coupled model hindcasts of extratropical climate anomalies have been made using a two-tier approach by Bengtsson et al. (1993). Here tropical Pacific SSTs are first hindcast using a coupled model consisting of an ocean GCM coupled to an atmospheric component derived from a statistical relationship between surface wind and SST. The SSTs so produced are then used to force an atmospheric GCM which can represent accurately the tropical extratropical interactions, and so hindcasts of extratropical variables can be made. The results of these hindcasts of precipitation for six major ENSO events of the last twenty years are encouraging, indicating that there are grounds to believe useful forecasts could be made for at least parts of the extratropics. The use of coupled models for seasonal prediction, particularly for the extratropics, demands that techniques be developed in order that the predictable component of the forecast signal can be separated from the essentially unpredictable component i.e. weather. The degree of predictability will depend on the initial state itself, but at present not enough comprehensive model studies have been performed to allow a very clear quantitative assessment of mid-latitude predictability.

4.2 The role of the other tropical oceans

One may wonder from the material covered in this paper if there is any other coherent climate variability besides ENSO. Certainly most effort in TOGA has been expended on measuring the Pacific, as it makes sense to tackle the biggest signal first. However, there does seem to be variability on decadal timescales which may be independent of ENSO, and there also seems to be interannual variability which may not be directly linked to ENSO although it may yet be that ENSO is lurking there somehow. There is not yet a clear enough view of the various components of the climate system and how they fit together. What understanding there is, is acquired from a mixture of empiricism and focussed modelling.

It is frequently the case that statistical connections have been noted between the climate in one part of the world and another before these connections have been succesfully modelled. These statistical linkages are the result of empirical studies and frequently lead to empirical prediction schemes. Considerable progress has been made in determining some empirical linkages during TOGA. It is anticipated that ultimately physically-based numerical models will surpass the skill of empirical schemes but for the moment empiricism can give a flavour of what might be possible. In some cases the empirical results have been confirmed by numerical studies.

Atlantic Ocean SST anomalies are associated with rainfall fluctuations over north east Brazil. (ENSO also influences rainfall in this region). The rain in this region is associated with the most southerly extent of the Atlantic InterTropical Convergence Zone from January to May. A dipole pattern of SST anomalies (Moura and Shukla 1981) centred on 15N and 15S is most relevant to the position of the ITCZ. Anomalies in the position of the ITCZ cause anomalous rainfall. When SST in the northern hemisphere is colder than normal, and/or the SSTs in the southern hemisphere are above normal, the ITCZ is displaced further south giving above average rainfall over north-east Brazil, and vice versa. The origins of the north-south oscillation are unclear, but it has a low frequency (decadal) variability as well as a higher frequency variability as found by Servain (1991). SST anomalies in the tropical Atlantic are smaller than in the Pacific, but they may on occasion be related to events in the Pacific. For example, the warm periods in the Pacific in 1983 and 1987 were followed a few months later by warming in the equatorial Atlantic. The tropical Atlantic may also generate its own mode of variability, analogous to ENSO but weaker (Zebiak 1993). Even in the case of the 1983/4 warming, while model studies confirm an ENSO origin, the amplitude of the signal in the Atlantic was larger than could be accounted for by ENSO alone, suggesting some amplification by processes internal to the Atlantic. (Delecluse et al 1994).

Empirical studies of rainfall variation in the African Sahel (Folland et al., 1991) suggest that SST variations in the Indian and Atlantic Oceans may be important predictors in addition to tropical Pacific SSTs. This empirical relationship has been supported with carefully constructed numerical experiments. For example, the association of Atlantic SSTs with Sahel rainfall that was tentatively suggested by Palmer (1986) has been well supported by extensive sets of modelling experiments in which SST is allowed to vary in one ocean at a time or in combinations, with a climatological distribution being used elsewhere. Also, a set of integrations of the UKMO general circulation model (GCM) has demonstrated skill in modelling interannual (and interdecadal) fluctuations in tropical North African rainfall, given observed SSTs (Rowell et al., 1992). The ability of the model to simulate the relatively wet conditions that prevailed in the 1950s, and the drought years in the 1980s is encouraging. There does not appear to be major sensitivity to initial conditions for this area, but the extent to which this depends on the individual model has still to be clarified. Later versions of the UKMO model have simulated Sahel rainfall variability less well than in Rowell et al 1992.

The statistical relationship between Australian climate and the sea-surface temperature (SST) distribution in the surrounding oceans has been observed for many years. The influence of the Southern Oscillation on Australian rainfall has been well documented, but recently the impact of anomalies in the global SST distribution on Australian climate has become apparent. For example, Nicholls (1989) has found that Australian winter rainfall is correlated with the gradient of SST between the central Indian Ocean and Indonesia, while Drosdowsky (1993) has shown that there is a lagged statistical relationship between Indian Ocean SSTs and Australian rainfall, which could provide a basis for long-range forecasting of Australian rainfall.

5 Conclusions

The observing and forecasting system so successfully developed by TOGA can not be assured. However, as climate predictions become progressively more skillful, useful and used, the hope is high that the ocean will be routinely observed. For the present, a new research programme is under development, CLIVAR, to carry forward the work started by TOGA.

One of the major successes of the TOGA programme has been the establishment of a near real-time observation network for the Pacific ocean. Up-to-date information is available on the World Wide Web or by direct ftp transfer.

Some of the data from the various observing platforms are assimilated in ocean models to provide analyses of the tropical oceans, but no model yet assimilates all the data. All the pieces needed to develop an integrated assimilation forecast system for TOGA using state-of-the-art CGCM's are being worked on, however, including assimilation of altimetric data. What remains is to pull the research together into one or more unified systems.

Forecasts are currently carried out using a number of coupled models, ranging from models of intermediate complexity to rather comprehensive general circulation models. The skill of the models as measured by correlation between hindcast/forecast NINO3 SST and observed looks encouraging with several models attaining skill levels of above 0.6 at leads of 1 year. This level can be increased by improving the ocean initial state. Different techniques exist for doing that: data assimilation, using 3DVAR or 4DVAR systems, by blending observed and model generated winds to reduce the initialisation schock, or by filtering the observed winds through the observed SST fields. No doubt other strategies will be developed in future. Whether anomaly correlation is an adequate test of model skill is debatable. What is predictable is still an open question. If it is only the large scale low frequency signal, then perhaps the intermediate models will be as successful as the more comprehensive models. At the time of writing it is not clear that one class of models out-performs any other. And a salutory lesson has been learned hopefully that nature is not always as cooperative as during the 80s. Predictions of the anomalies for the 90s were not very good.

It is only by developing full state-of-the-art systems that the potential for prediction can be fully exploited. An experimantal programme to develope a full end-to-end system at ECMWF is in progress.

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