Annu. Rev. Fluid Mech. 1994. 26: 617–59

DYNAMICS OF COUPLED OCEAN-ATMOSPHERE MODELS: The Tropical Problem

J. David Neelin

Department of Atmospheric Sciences, University of California, Los Angeles, California 90024

Mojib Latif

Max-Planck-Institut für Meteorologie, D-20146 Hamburg 13, Federal Republic of Germany

Fei-Fei Jin

Department of Meteorology, University of Hawaii at Manoa, Honolulu, Hawaii 96822

KEY WORDS: Ocean-atmosphere interaction, climate variability, El Niño/Southern Oscillation

INTRODUCTION

Large-scale ocean-atmosphere interaction plays a crucial role in natural climate variability on a broad range of time scales and in anthropogenic climate change. The development of coupled ocean-atmosphere models is thus widely regarded as essential for simulating, understanding, and predicting the global climate system. Although these efforts typically benefit from years of previous work with atmospheric and oceanic models, coupling the two components represents a major step because of the new interactions introduced into the system. These can produce new phenomena, not found in either medium alone, the mechanisms for which present exciting theoretical problems. The removal of artificial negative feedbacks

0066–4189/94/0115–0617$05.00

produced by fixed boundary conditions in the uncoupled case also provides a stringent test of physical processes represented in both component models.

Pioneering work on coupling oceanic and atmospheric general circulation models (GCMs) began during the late 1960s and the 1970s (Manabe & Bryan 1969, Bryan et al 1975, Manabe et al 1975, Manabe et al 1979, Washington et al 1980). The difficulties encountered in obtaining accurate climate simulations with these models were sufficient that use of such coupled GCMs (CGCMs) did not gain momentum until the late 1980s and early 1990s. While the anthropogenic warming problem drove the development of global models (e.g. Gates et al 1985, Schlesinger et al 1985, Sperber et al 1987, Bryan et al 1988, Manabe & Stouffer 1988, Washington & Meehl 1989, Stouffer et al 1989, Manabe et al 1990, Cubasch et al 1992, Manabe et al 1992), evidence that ocean-atmosphere interaction is responsible for the El Niño/Southern Oscillation (ENSO) phenomenon provided a driving force in the development of models aimed at the tropical regions, both CGCMs and less complex models.

In this article, we consider the dynamics of coupled models relating to internal variability of the climate system that arises through ocean-atmosphere interaction. We focus on the tropical problem because it has been more thoroughly studied than the extratropical problem, and the crucial role of coupling has been clearly demonstrated. The field has developed to a stage that can be well summarized, and where short-range climate prediction is becoming a reality. A briefer section provides an indication of developments for the problem of coupled extratropical variability, which is in its infancy.

Despite the importance of coupled models to the study of anthropogenic global warming, we do not address this question beyond providing an indication of some of the difficulties these models face. It is the subject of many articles (e.g. Mitchell 1989, Houghton et al 1990, Gates et al 1992 and references therein) and merits a separate review. For other general references on coupled models, we note a review of global CGCMs (Meehl 1990a), a textbook on the tropical problem (Philander 1990), edited volumes on climate modeling (Trenberth 1993, Schlesinger 1990), and selected conference proceedings (Nihoul 1985, 1990; Charnock & Philander 1989).

COUPLED OCEAN-ATMOSPHERE MODELS

A hierarchy of complexity exists in climate models, the most complex being the atmospheric, oceanic, and coupled general circulation models (AGCMs, OGCMs, and CGCMs; for these and other acronyms, see Table 1). GCMs are generally based on the primitive equations (a filtered version of the Navier-Stokes equations; e.g. Washington & Parkinson 1986), with detailed parameterizations of sub-grid-scale processes (e.g. turbulent mixing, and for AGCMs radiative transfer and moist convection). These attempt to simulate an approximation to both the climatology and natural variability. A variety of models based on further approximations are used for particular applications; often these are formulated as anomaly models about a specified climatology. Coupling considerations tend to be similar—we outline the procedures as applied to GCMs. The class of models often used in global warming studies in which the ocean acts only as a heat capacitor—and has no active dynamics—is not discussed.

For climate time scales, a division of the coupled system at the ocean-atmosphere interface is not easy to defend. Incoming solar (shortwave) radiation is primarily absorbed at the ocean surface and energy is lost through evaporation, infrared (longwave) radiation, and sensible heat fluxes to the atmosphere, which in turn re-emits longwave radiation to space. The one-dimensional equilibrium of these processes (and the strength of the negative feedback to perturbations from this equilibrium) provides a first approximation to the climate, modified of course by three-dimensional transports and feedbacks in both media. Interrupting this exchange at the ocean surface is questionable on time scales longer than a few months (shorter for some phenomena). Historically, however, this division permitted atmospheric and oceanic modelers to concentrate purely on problems in their respective media, as necessitated by the complexity of these subsystems. Since the parameterization of sub-grid-scale processes is one of the most crucial aspects of climate modeling, this separate development may be partially justified by arguing that the difference in density and effective heat capacity is sufficient that individual parameterizations of fast sub-grid processes may be developed initially in uncoupled models. The limitations of this approach will no doubt be re-examined when coupled models reach a more mature stage. Surface heat flux boundary conditions for uncoupled ocean models are particularly problematic (e.g. Bretherton 1982, Seager et al 1988) since the negative feedback on sea surface temperature (SST) involves the atmospheric response.

Table 1 Acronyms used in the text

ENSO	El Niño/Southern Oscillation
GCM	General Circulation Model
AGCM/CGCM/OGCM	Atmospheric/Coupled/Ocean GCM
HCM	Hybrid Coupled Model
ICM	Intermediate Coupled Model
SSO regime	Standing-SST Oscillatory regime
SST	Sea Surface Temperature

In a typical coupling scheme for an ocean-atmosphere model, the ocean model passes SST to the atmosphere, while the atmosphere passes back heat flux components, freshwater flux, and horizontal momentum fluxes (*surface stress*—oceanographic usage refers only to stress tensor components associated with vertical fluxes of horizontal momentum). Land temperature is necessarily computed interactively, with parameterizations ranging from the zero heat-capacity approximation to more complex land-surface models (e.g. Dickinson 1983). The numerical coupling interval (over which interfacial variables are averaged before being passed) is chosen for computational convenience or to satisfy assumptions of physical parameterizations. Although heat fluxes are calculated using the atmospheric boundary-layer parameterizations based on SST from the previous interval, the important dependence of heat flux on SST is retained as long as the heat flux coupling interval is sufficiently small.

The atmospheric response to SST is rapidly redistributed vertically, especially in convective regions, and is nonlocal horizontally on time scales longer than dynamical adjustment times—on the order of a few days to a month. For most purposes, the atmosphere can be assumed to be in statistical equilibrium with given SST (and land/ice/snow) boundary conditions on time scales longer than a season. The ocean responds on a wide range of time scales, from days (for some features of the mixed layer) to millenia (for the deep-ocean thermal adjustment). It is thus common to characterize the ocean as having the *memory* of the system. For global coupled models where the deep ocean is integrated to equilibrium, asynchronous coupling techniques are sometimes used (e.g. Manabe et al 1979).

Climate drift—i.e. departure of the model climatology from the observed (and from the climate simulated by the component models in uncoupled tests)—is a common problem in coupled models. It often appears as a slow adjustment away from initial conditions towards an internal equilibrium, hence the term "drift;" it may also refer to cases of faster adjustment and to the error at equilibrium. Although numerics contribute, climate drift arises primarily from the cumulative effects of errors in the sub-grid parameterizations; as such the process of correcting it based on careful physical arguments can be slow and painstaking. In cases where the sources of drift are well separated from mechanisms governing the geophysical phenomena of interest, it has been argued (e.g. Manabe & Stouffer 1988, Sausen et al 1988) that correcting the drift by a *flux correction* may permit progress even with an imperfect model. Roughly speaking, the model's equilibrium climatology of all or some of the interfacial variables is subtracted and replaced with observed values that are passed between subsystems; effectively the model is only used to compute anomalies from climatology. The success of flux-correction techniques depends on the problem.

Coupled models designed for the tropical problem do not treat the deep-ocean thermohaline circulation which maintains cold waters at depth. Typically the ocean basin is simply interrupted at some latitude, using a sponge layer (with temperature and salinity strongly constrained toward climatological values) to avoid effects of the artificial boundaries propagating into the region of interest via wall-trapped Kelvin waves. Observed climatological SST is specified in the ocean regions which are not actively modeled. Other models simulate the upper ocean only, with motionless deep waters (e.g. Gent & Cane 1989). Additional design specifications for the tropical problem include the use of sufficiently high-resolution ocean components to resolve equatorial wave dynamics with characteristic meridional scales of order 2° latitude. For the global problem, the ocean models typically are used with coarser resolution because of the necessity of very long integrations for equilibration.

THE TROPICAL PROBLEM

Background

Ocean-atmosphere interaction is particularly amenable to study in the tropics because at large scales each medium is strongly controlled by the boundary conditions imposed by the other. The upper ocean circulation is largely determined by the past history of the wind stress with little internal variability; likewise the major features of the tropical atmospheric circulation are determined by the SST, with internal variability largely confined to time scales less than 1–2 months. This contrasts to the midlatitude situation where internal variability of both atmosphere and ocean is large.

THE BJERKNES HYPOTHESIS Because the ENSO phenomenon is the largest signal in interannual climate variability, it has dominated the literature; here we bring in other aspects of the tropical problem where possible. The reigning paradigm for ENSO dynamics is that it arises through ocean-atmosphere interaction in the tropical Pacific (although its influence extends globally and interactions with other parts of the climate system are by no means excluded), as first hypothesized by Bjerknes (1969). The essence of Bjerknes' postulate still stands as the basis of present day work—that ENSO arises as a self-sustained cycle in which anomalies of SST in the Pacific cause the trade winds to strengthen or slacken, and that this in turn drives the changes in ocean circulation that produce anomalous SST. Within this paradigm, one may still distinguish a variety of mechanisms that potentially contribute to the maintainance and time scale of the cycle; these have provided challenges for both theory and simulation.

MODEL HIERARCHY Beginning at about the same time as Bjerknes' hypothesis was formulated, the foundations for modeling the tropical coupled system were laid through the study of the individual physical components. The dynamics of the equatorial ocean response to wind stress were examined in shallow-water models representing the upper ocean (e.g. Moore 1968; Cane & Sarachik 1977, 1981; McCreary 1976), modified shallow-water models (e.g. Cane 1979, Schopf & Cane 1983), and ocean general circulation models (OGCMs; e.g. Philander & Pacanowski 1980, Philander 1981). And in the atmosphere, it was demonstrated semi-empirically that simple atmospheric models with steady, damped shallow-water dynamics could provide a reasonable approximation to the low-level tropical atmospheric response to SST anomalies (e.g. Matsuno 1966, Gill 1980, Gill & Rasmusson 1983). There is still disagreement as to the best formulation of these simple atmospheric models (Zebiak 1986, Lindzen & Nigam 1987, Neelin & Held 1987, Neelin 1989a, Allen & Davey 1993) but their simulation of anomalous wind-stress feedbacks to the ocean from given SST is given credence by AGCM simulations (e.g. Lau 1985, Palmer & Mansfield 1986, Mechoso et al 1987, Shukla & Fennessey 1988, and references therein).

As a result of the development of complementary models of varying degrees of complexity, the tropical coupled problem has benefited from a full hierarchy of models. The basis for a more quantitative understanding of coupled ocean-atmosphere interaction was initially provided by coupled models constructed from variations on modified shallow-water ocean and simple atmospheric models: both in simple linear versions (Lau 1981a; Philander et al 1984; Gill 1985; Hirst 1986, 1988; Wakata & Sarachik 1991; Neelin 1991) and in nonlinear versions (e.g. Cane & Zebiak 1985, Anderson & McCreary 1985, Zebiak & Cane 1987, Battisti 1988, Battisti & Hirst 1989, Schopf & Suarez 1988, Yamagata & Masumoto 1989, Graham & White 1990). The simplest linear shallow-water models, together with some useful models that condense the dynamics even further, are loosely referred to as *simple models*, while the more complex and carefully parameterized of the modified shallow-water models are often referred to as *intermediate coupled models* (ICMs). The next step up the model hierarchy, in order of increasing complexity, is the *hybrid coupled models* or *HCMs*. These consist of an ocean GCM coupled to a simpler atmospheric model (e.g. Neelin 1989b, 1990, Latif & Villwock 1990, Barnett et al 1993), the justification being that the ocean contains both the memory and limiting nonlinearity of the system—the atmosphere is thus treated as the fast component of a stiff system. The most complex models are the coupled GCMs in which both components include relatively complete sub-grid parameterization packages (e.g. Philander et al 1989, 1992; Lau et al 1992; Sperber & Hameed 1991; Gordon 1989; Meehl 1990b; Nagai et al 1992; Mechoso et al 1993; Neelin et al 1992). It should be noted that the divisions in the hierarchy are not sharp and some of the lowest-resolution CGCMs may not be much more complex than the best ICMs. Many of these models produce interannual variability through coupled interactions which have significant parallels to ENSO dynamics.

Our approach here is to summarize basic phenomenological features from a modeler's point of view (i.e. we do not attempt a complete review of the large observational literature) in the *Observations* section, and then to present a cross-section of model results in the *Simulation* section, which includes selected intermediate models as well as CGCMs and HCMs. The *Theory* section makes use of intermediate and simple models to outline basic mechanisms of interaction, describes the manner in which different mechanisms combine and contribute to the sensitivity of the coupled system, and details the current understanding of the bifurcation structure. Many of the theoretical considerations prove useful in understanding results of the more complex models.

Observations

Aspects of El Niño and the Southern Oscillation were known individually long before any connection was made. The term "El Niño," which originated with Peruvian fishermen, now refers to strong warmings of surface waters through the eastern and central equatorial Pacific that last about a year (e.g. Rasmusson & Carpenter 1982, Deser & Wallace 1990). Although it is common to refer to these as "events," they exhibit a distinct oscillatory behavior now understood to be part of a low frequency cycle. The Southern Oscillation was discovered by Walker (1923), and its global scale was inferred early on (Belarge 1957) from correlation maps of sea-level pressure anomalies which exhibit anomalies of opposite sign in the eastern and western hemisphere. The larger scale of this pattern relative to the SST anomaly is typical of the atmosphere's nonlocal response to boundary conditions. The strong relationship between interannual variability of SST and sea-level pressure may be seen in Figure 1.

As a background to understanding ENSO-related interannual variability, a brief description of the time-mean circulation is required. Differential forcing of the atmosphere by the SST boundary condition thermodynamically drives direct circulation cells: convection tends to organize roughly over the warmest SST, producing regions of strong surface convergence (known as intertropical convergence zones). The zonally-symmetric (i.e. averaged around latitude circles) component of this circulation

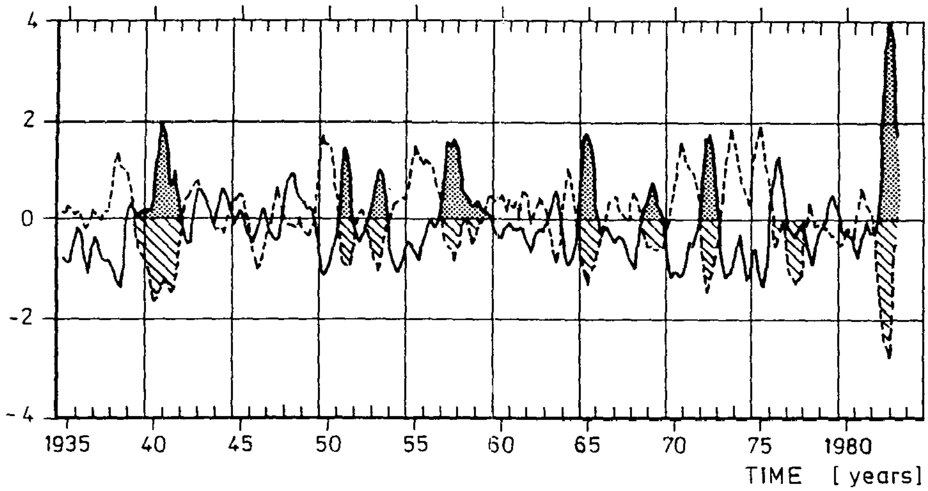

Figure 1 Time series of the Southern Oscillation Index, which measures the atmospheric sea-level pressure gradient across the tropical Pacific basin (*dashed curve*), and sea surface temperature (SST) anomalies at Puerto Chicama, Peru (*solid curve*). Both series are normalized by their standard deviation; shading indicates major ENSO warm phases (high SST, low Southern Oscillation Index). After Rasmusson (1984).

is referred to as the *Hadley circulation*, the zonally-asymmetric component as the *Walker circulation*. The Hadley circulation contributes an easterly (i.e. westward) component to tropical surface winds. This is strongly reinforced over the tropical Pacific by the Walker circulation driven by the strong SST gradient across the basin between the warm waters in the west and the cooler eastern waters.

The westward wind stress has a strong impact on the ocean circulation. The input of momentum is balanced, in a vertical average, largely by pressure gradients in the upper ocean. A sea-level gradient of about 40 cm across the Pacific is compensated by a slope in the *thermocline* (the interface that separates the well-mixed, warm surface waters from the cold waters at deeper levels) which slopes upward to the east. Within the upper ocean, the differential deposition of stress by vertical viscosity drives westward surface currents along the equator, and Ekman drift due to the Coriolis force to either side of the equator drives a narrow band of upwelling along the equator, especially under the regions of strong easterlies in the eastern/central Pacific. The combination of upwelling and shallow thermocline produces the *equatorial cold tongue* in the east, while the deep thermocline in the west is associated with warm SST—the western Pacific *warm pool*.

The important dependence of SST in the equatorial cold tongue region on wind-driven ocean dynamics (rather than just on air-sea heat exchange) and the Walker circulation response to anomalies in the SST pattern form the key elements of the Bjerknes hypothesis. Consider an initial positive SST anomaly in the eastern equatorial Pacific. This anomaly reduces the zonal SST gradient and hence the strength of the Walker circulation, resulting in weaker trade winds at the equator. This leads to a deeper thermocline and reduced currents and produces higher surface temperatures in the cold tongue region, further reducing the SST gradient in a positive feedback which can lead to instability of the climatological state via ocean-atmosphere interaction. The cyclic nature of the unstable mode depends on the time scales of response within the ocean. The details of what produces the cycle are subtle, as elaborated in the *Theory* section, but a concise observational picture motivated by theoretical considerations is provided by Latif et al (1993b).

Figure 2 shows characteristic anomaly patterns of three crucial quantities: zonal wind stress, SST, and the depth of the thermocline or upper ocean heat content, as measured by depth of the 20°C isotherm. The patterns represent an estimate of the dominant coupled ENSO mode as obtained by principal oscillation pattern analysis (Hasselmann 1988)—specifically, the leading eigenvector of the system matrix obtained by fitting a first-order Markov process to the data, where oscillations are represented by the cycle of patterns in temporal quadrature. The right panels show conditions during the warm phase of the ENSO cycle, i.e. during El Niño (the cold phase simply has reversed signs under this technique). Most of the tropical Pacific is covered by anomalously warm surface waters (Figure 2*d*), with maximum anomalies in the eastern equatorial Pacific. These SST anomalies are highly consistent with the patterns obtained by other techniques, including the well-known Rasmusson & Carpenter (1982) composites. The positive SST anomaly is accompanied by a westerly (eastward) zonal wind stress anomaly (Figure 2*b*) which reduces the mean Walker Circulation. Consistent with this feature, the tilt in the thermocline is reduced as indicated by the negative anomalies in the upper ocean heat content which are centered off the equator (Figure 2*f*).

The phase differences necessary to maintain the oscillation exist between sea surface temperature and wind on the one hand and upper ocean heat content on the other. As described in the *Theory* section, the ocean is not in equilibrium with the atmosphere and carries information associated with past winds that permits continuous oscillations. This feature is clearly seen during the transition phase in upper ocean heat content (Figure 2*e*) which shows a pronounced equatorially-trapped signal in the western Pacific. This signal appears not to be related to the contemporaneous winds (Figure 2*a*), but rather was generated by anomalous eastward winds of the preceding cold phase (Figure 2*b*, but with reversed signs). Equatorial wave dynamics dictate that the heat content anomalies at latitudes larger than a few degrees propagate westward and reflect at the western maritime boundary into the equatorial wave guide. The transition phase SST (Figure 2*c*) does not show a clear signal; variations in SST can therefore be described to first order as a standing oscillation. Thus, it is the subsurface memory of the ocean that is crucial to ENSO (see e.g. Latif & Graham 1992 and Graham & White 1990 for additional observations).

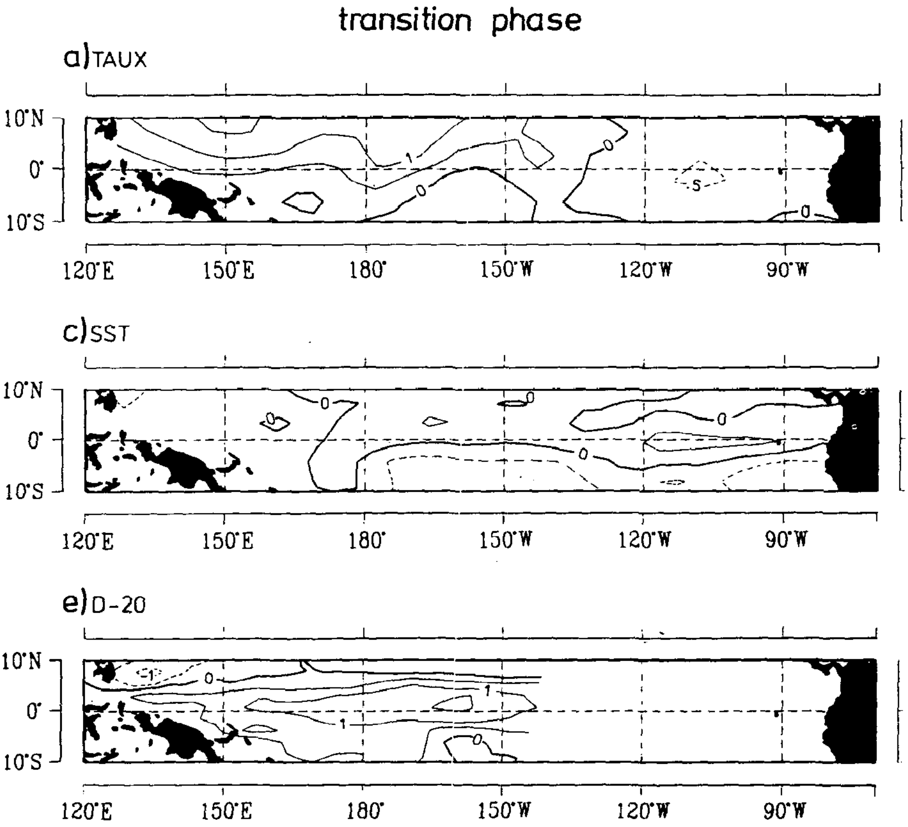

Figure 2 Spatial patterns of the dominant mode of ENSO variability as represented by the leading principal oscillation pattern (see text). The oscillation is represented by two time phases in quadrature during the cycle: transition phase (panels *a*, *c*, *e*) and extreme phase (panels *b*, *d*, *f*). (*a*), (*b*) wind stress anomaly, (*c*), (*d*) sea surface temperature anomaly, (*e*), (*f*) heat content anomaly as measured by the depth of the 20°C isotherm (blank areas in the eastern Pacific are due to lack of subsurface data). After Latif et al (1993b).

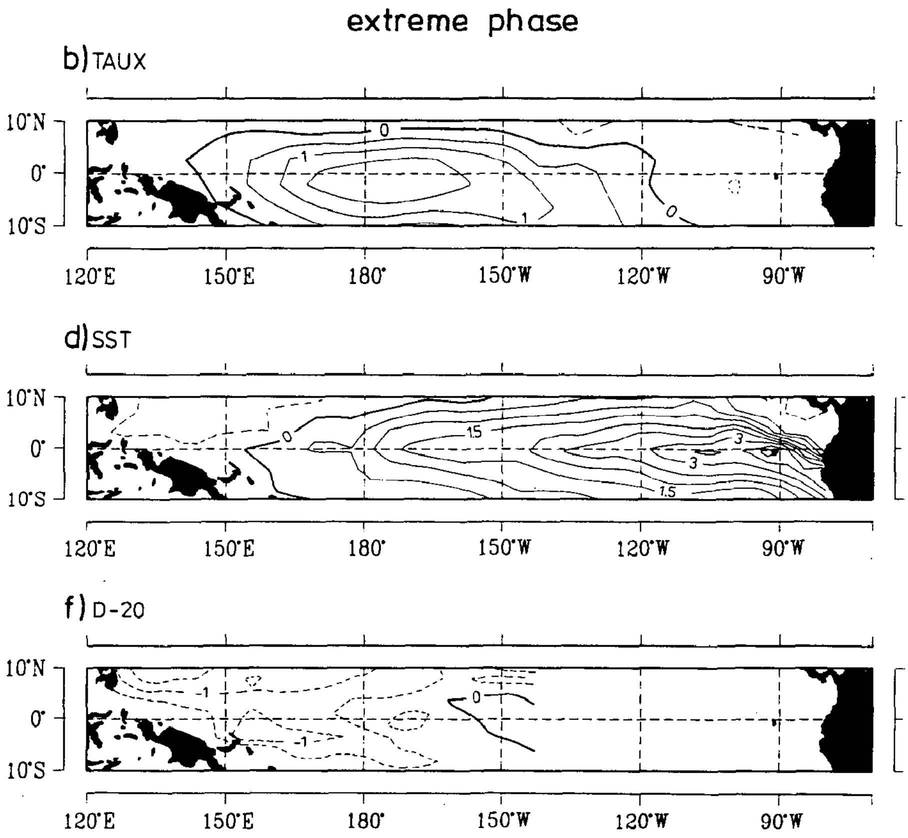

Figure 2 (continued).

The transition phase zonal wind stress (Figure 2*a*) shows a pronounced westerly anomaly centered over the northwestern Pacific so that the evolution in zonal wind stress is also characterized by a slowly eastward-propagating feature. The role of this propagation in maintaining the ENSO cycle, however, is still a controversial issue. Several authors have argued that this feature indicates a link to circulation systems over India, in particular the Monsoon (e.g. Barnett 1983).

A complementary view of the oceanic side of this feedback is provided by time-longitude plots of SST and a measure of thermocline depth anomalies along the equator (Figure 3). The time series is limited by the length of the records of ocean subsurface temperature. Even without statistical techniques, it is easy to pick out the dominant standing oscillation pattern in SST (although some hints of propagation may be noted—see e.g. Gill & Rasmusson 1983, Barnett et al 1991), and the characteristic signature of subsurface memory—the lead of the heat content anomalies in the western part of the basin relative to the eastern part. Several coupled ocean-atmosphere models simulate variability patterns to those described above.

There is evidence that the spectral peak associated with ENSO may have a quasi-biennial component in addition to the dominant low-frequency

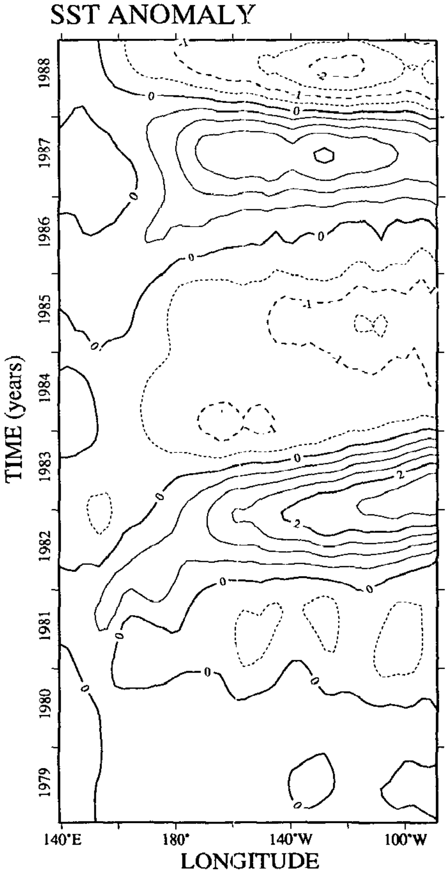

Figure 3 Time-longitude plot of observed anomalies along the equator. (*Left*) SST (contour interval 0.5°C). (*Right*) heat content integrated above 275 m (contour interval 100°C m). The data have been low-pass filtered to remove variability on time scales smaller than 17 months. Data sets are described in Reynolds (1988) and Barnett et al (1993), respectively.

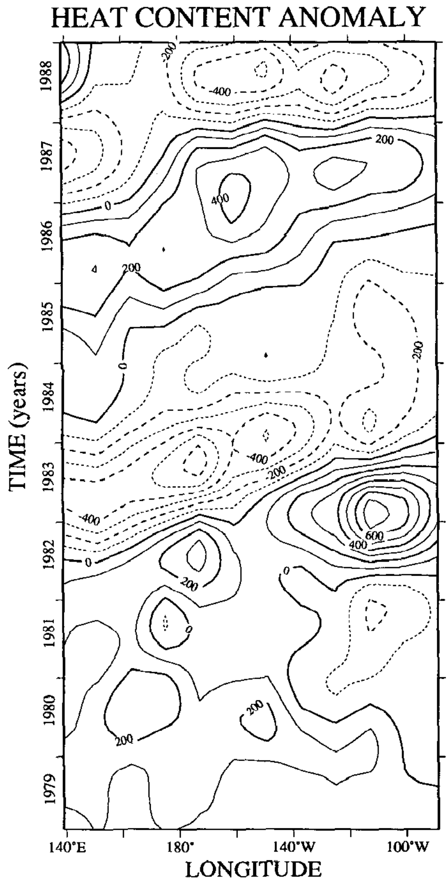

Figure 3 (continued).

(3–6 year) component, e.g. Rasmusson et al (1990), Latif et al (1993b). Spatial structures and interactions between assumed spectral bands have been examined, e.g. by Barnett 1991, Ropelewski et al 1992. For further discussion of ENSO observations see, for instance, Cane (1986), Rasmusson & Wallace (1983), Philander (1990), and references therein. Discussion of the seasonal cycle and interannual variability in the tropical Atlantic may be found in Lamb et al (1986), Lough (1986), Wolter (1989), Servain & Legler (1986), Philander & Chao (1991), Houghton & Tourre (1992), and Servain (1991), while Zebiak (1993) gives evidence that the latter may in part share similar dynamics to ENSO.

Models and Simulation

AN INTERMEDIATE MODEL The intermediate coupled model of Cane & Zebiak (1985, with Zebiak & Cane 1987; collectively CZ hereafter) has proven influential in ENSO studies and has provided the first successful ENSO forecasts with a coupled model (see *Prediction* section). A version of the ocean component is described in the *Theory* section. The atmosphere (Zebiak 1986) is one of several simple atmosphere models which attempt to improve on that of Gill (1980); drawbacks include lack of a moisture budget and formulation with discontinuous derivatives, but similar results are obtained with different atmospheric models (Jin & Neelin 1993a, N. Graham, personal communication). Figure 4 shows the SST and thermocline depth anomalies over one period of the simulated ENSO cycle from the linearized version of the CZ model used by Battisti & Hirst (1989) to examine the essential dynamics. The typical stationary oscillation in SST may be seen, with the lead of the western-basin thermocline-depth anomaly relative to the eastern basin characterizing the subsurface memory. The details of the transition between west and east differ from those observed because the simulated winds are shifted relative to observed winds, but the cycle is not strongly sensitive to this. Simulated ENSO events tend to resemble each other strongly in this model, and Battisti (1988) and CZ disagree over the degree of irregularity that can be generated by internal model dynamics, but there is reasonable consensus that basic elements of ENSO dynamics are captured.

INTERCOMPARISON OF GCM SIMULATIONS A recent comparison (Neelin et al 1992) of the tropical simulations of seventeen coupled ocean-atmosphere models, contributed by a dozen institutions worldwide, represents a snapshot at a relatively early stage of a rapidly developing field. We review some of the results, with the caveat that in the brief time since their collection, several of the models have made great progress in the accuracy of simulation and new models have been developed which are not yet published. The comparison was intended to give a feel for the sensitivity of the system modeled (possible in part because the models were not yet optimized), to point out common problems, and to provide a forum for discussing the broad range of coupled-model behavior.

The models were selected on the basis of having at least one component of sufficient complexity to be called a GCM, (i.e. CGCMs and HCMs), with two representatives of the ICMs—those of Cane & Zebiak (1985) and Schopf & Suarez (1988), the latter differing from a GCM principally by lack of a moisture budget. Some of the models are global, designed for global warming studies; others have a dynamically-active ocean only in the tropical Pacific, and were designed for the tropical problem. SST was

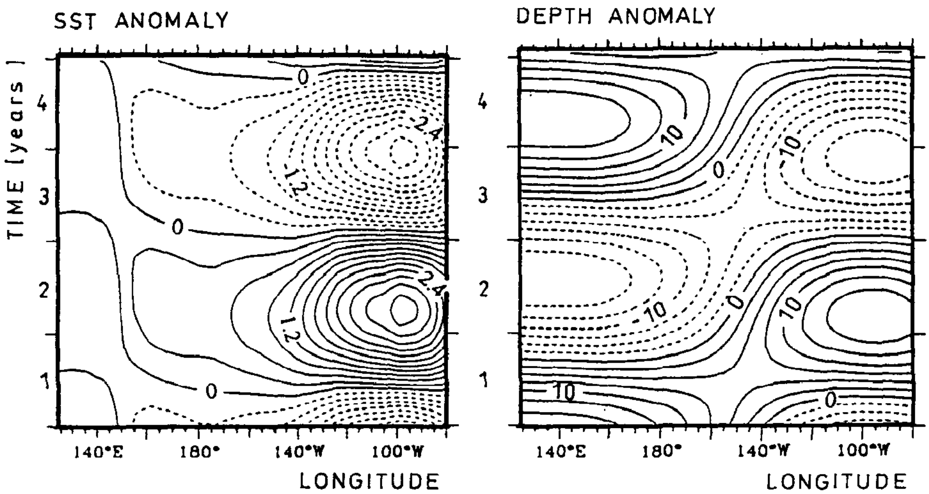

Figure 4 Time-longitude plot of anomalies along the equator from the Battisti & Hirst (1989) linearized version of the Cane & Zebiak (1985) intermediate coupled model. (*Left*) SST (contour interval 0.3°C). (*Right*) thermocline depth (contour interval 2.5 m). After Battisti & Hirst (1989).

chosen as the principle variable of comparison because of its crucial role in mediating the interactions.

Table 2 provides a summary of the results, augmented with more recent results where published, roughly classified according to the type of interannual variability and the simulation of climate in the equatorial Pacific. Models are listed as in Neelin et al (1992); the most closely related independent references available are Endoh et al (1991), Gent & Tribbia (1993), Gates et al (1985), Gordon (1989), Latif et al (1988), Latif et al (1993a), Lau et al (1992), Meehl (1990b), Neelin (1990), Philander et al (1992), Schopf & Suarez (1988), Sperber & Hameed (1991), Zebiak & Cane (1987). Many of the models exhibit climate drift. Some of the models, especially those with simplified atmospheres, sidestep this problem by flux correction. The category "Modest drift" as used here means only that the degree of drift in SST was relatively small by current (subjective) standards and comparable to that of uncoupled components. Interannual variability is weaker than observed in many of the models—the category "Weak interannual variability" means too weak to be classified.

Climate drift occurs in a variety of forms. A general cooling of large parts of the ocean basin is the most common form of slow drift. Fast climate drift is characteristically a coupled-dynamical effect leading to an

Table 2 Summary of models grouped according to common behavior for both tropical climatology and interannual variability as reflected in the sea surface temperature field (after Neelin et al 1992)

Variability	Climate		
	Modest drift	Flux corrected	Other
Weak interannual variability		Latif et al-1 Neelin-2[b,c] Oberhuber et al-1[d]	Gordon & Ineson-2[a] Gates et al[d,e] Cubasch et al[d,e] Oberhuber et al-2[d,e]
Interannual variability with zonal propagation of SST anomalies	Lau et al[d]	Neelin-1[b]	Meehl & Washington[a,d,f] Gates & Sperber[a,d,f] Tokioka et al[a,f]
Interannual variability with standing SST anomalies	Philander et al Gent & Tribbia Nagai et al (1992)[g] Mechoso et al (1993)[g]	Zebiak & Cane[b] Allaart et al[b,c]	Schopf & Suarez[a,b] Latif & Sterl[a]

[a] Slow cooling of warm regions.
[b] Model with simplified atmospheric component.
[c] Multiple climate states known or suspected.
[d] Model with global-domain ocean component.
[e] Weak zonal gradient; weak cold tongue.
[f] Cold tongue extended or cold tongue/warm pool boundaries displaced.
[g] Dates given for recently added references; otherwise see Neelin et al (1992).

overly-weak or overly-strong equatorial cold tongue. Three-dimensional feedbacks between SST, convection zones, wind stress, and ocean circulation qualitatively similar to those responsible for El Niño are seen to play a role in creating such drift or in exacerbating weaknesses in parameterizations controlling one-dimensional, vertical-column processes such as cloud-radiative interaction or vertical mixing. We note many situations where the position of the cold tongue migrates or extends within the basin, with a warm pool developing in the eastern part of the basin in some instances. The observed convection zone in the eastern Pacific stays north of the equator in all seasons; in some models it migrates across the equator with season. The similarities between the fast mode of climate drift to interannual phenomena of comparable time scale implies that, unlike numerical weather prediction—in which correction of climate drift was only addressed as the models matured—interannual climate forecasting with coupled GCMs must address the accurate simulation of certain aspects of the climatology at a relatively early stage.

We find that there is little relation between the presence of climate drift and the existence of significant interannual variability, so long as the cold tongue is present somewhere in the basin. Interannual variability tends to come in two varieties: cases in which anomalies in SST, wind, etc propagate in the longitudinal direction along the equator and cases in which anomalies develop as a standing oscillation in the cold tongue region. In the latter case, fine ocean model resolution is required near the equator and subsurface memory due to oceanic adjustment processes is believed to determine the time scales; in the former case, coarse ocean model resolution does not preclude interannual oscillations and the time scales of ocean wave dynamics are not essential to the period.

Figure 5 provides an example of interannual variability from one of the first coupled GCMs with a high-resolution tropical ocean component (Philander et al 1989, 1992; Philander et al in Table 2). While the spectrum of interannual time scales may not exactly match that observed (possibly due to the removal of the seasonal cycle in this model for hypothesis-testing purposes), the spatial form, again with dominant standing oscillations in SST and with subsurface phase lags, is reasonably close to the observed form; Chao & Philander (1993) also compare these results to the uncoupled ocean component forced with obsrved winds to provide a longer surrogate time series for the subsurface anomalies. A number of other CGCMs have variations on this spatial form, some having clearer propagation characteristics in SST, combined with significant subsurface phase lags (e.g. Nagai et al 1992, Latif et al 1993a).

The rich variety of coupled phenomena found in these models serves as an indication of the sensitivity of the coupled system and lends support to qualitative arguments that coupled feedbacks are crucial in establishing tropical climate features. Even the most important features, such as the extent and position of the equatorial cold tongue and western Pacific warm pool, are not guaranteed to be reproduced in coupled GCMs. The lack of robustness in these features does not necessarily imply major faults in the models since coupled feedbacks can turn a small deficiency in one of the components into a significant departure in the coupled climatology. For example, a tendency of the atmospheric model to give slightly weak easterlies can result in a weaker cold tongue which in turn further weakens the Trades. In some models this can lead to a permanent warm state, although in others, weak AGCM stresses do not adversely affect either climatology or interannual oscillations.

Because the behavior of the coupled system can be qualitatively different (and difficult to anticipate) from that of the individual components, coupling should be regarded as a crucial part of the testing and development procedure for AGCMs and OGCMs being used for climate studies. In particular, the simulation of the warm-pool/cold-tongue configuration in the equatorial Pacific can represent a stringent test of the combined effects

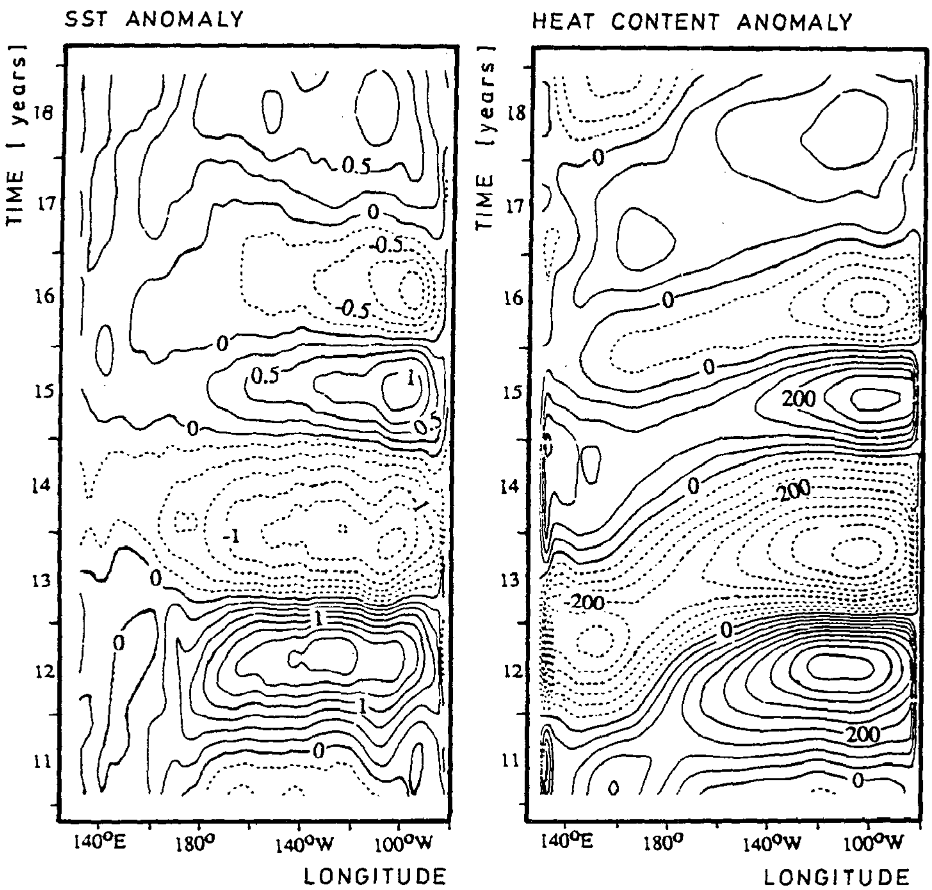

Figure 5 Time-longitude plot of anomalies along the equator from the Philander et al (1992) coupled GCM. (*Left*) SST (contour interval 0.25°C). (*Right*) heat content integrated above 300 m (contour interval 50 C° m). After Chao & Philander (1993). The data have been low-pass filtered to remove time scales less than 24 months.

of vertical-mixing parameterizations, interactive cloud-radiative schemes, and surface-flux parameterizations with the three-dimensional dynamics. The rate of improvement of recent model versions (both those in the table and currently unpublished models) is particularly encouraging in this respect.

Theory

CONTEXT AND HISTORY The considerable differences in the nature of the coupled variability produced by the different models above is related to the sensitivity and the rich variety of flow regimes found in ICMs and simple models, which exhibit multiple mechanisms of coupled interaction. The character of the interannual variability in nonlinear models is largely determined by the first bifurcation from the climate state (Neelin 1990, Münnich et al 1991)—in other words by the leading unstable mode of the system linearized about the climatological state. Many of the most pressing questions about the range of coupled variability found in coupled models can thus be addressed by understanding the relation between flow regimes in the linear problem. To keep this multiparameter bifurcation problem tractable, the key is to choose a few crucial parameters that capture the range of behavior of interest, and to map out the connections among regimes close to that of the real system and those that provide useful simplifications.

In the literature, the search for simple prototype systems to provide conceptual analogs for the modes of coupled variability has led in a number of apparently contradictory directions, and it is desirable to bring these together. We approach this by presenting first a version of the CZ ICM scaled to highlight parameters used to show these connections succinctly. We derive three important simple models from this and discuss the differing idealizations. We then return to the ICM to show how the simple models relate to the connections between eigenmodes in the coupled parameter space. This completes the discussion of the primary bifurcation, i.e. how the period and spatial form of the ENSO cycle are determined and its maintenance through instability of the climatological state. We conclude with a discussion of higher bifurcations and describe what is known about the sources of irregularity in the ENSO cycle.

In ordering the presentation to emphasize a unified view, the historical aspects are necessarily simplified, so we preface with a brief overview of the literature (see also McCreary & Anderson 1991, Ghil et al 1991). Early theoretical work includes low-order models by McWilliams & Gent (1978) and some nonrotating coupled cases (Lau 1981a). Models by McCreary (1983) and McCreary & Anderson (1984) have often been omitted from recent citation because of the use of a discontinuous switch in their atmosphere, but elements of their discussion of basin adjustment processes have been incorporated in later work. Philander et al (1984) presented the first linear instability study in a coupled modified-shallow-water system, and refinements and additional mechanisms were elaborated numerically in Gill (1985), Yamagata (1985), Hirst (1986, 1988), Battisti & Hirst (1989), Wakata & Sarachik (1991), and analytically in Neelin (1991). Nonlinear solutions in ICMs were introduced in Cane & Zebiak (1985) and Zebiak & Cane (1987), in a regime now felt to approximate that of the observed, and by Anderson & McCreary (1985), Yamagata & Masumoto (1989) in a different regime; hints at regime connections may be found in Xie et al (1989) and Wakata & Sarachik (1991).

Much of the terminology used in these papers is based on the Rossby and Kelvin modes of the uncoupled ocean in an infinite or periodic basin, presumably because these are most familiar to oceanographers. A significant step toward thinking in terms of the fully coupled problem was advanced by Schopf & Suarez (1988) and Suarez & Schopf (1988) using a simple model with a single spatial variable to explain the oscillation in their ICM; Battisti & Hirst (1989) showed that a version of this model could be fitted to a number of important aspects of the oscillation in the CZ model, and that the Hopf-bifurcation regime was the physically relevant one. Referred to hereafter as the *SSBH delayed-oscillator model*, it consists of a differential-delay equation representing the time evolution of SST averaged over a small eastern equatorial box, with a net growth tendency representing local positive feedback mechanisms due to coupling and a delayed negative feedback representing the equatorial-wave adjustment process; whether the latter can be interpreted literally in terms of off-equatorial Rossby wave packets reflecting from the western boundary has been the subject of debate (Graham & White 1988, Battisti 1989, Chao & Philander 1993). The model is designed to represent the regime in which SST variability occurs as a standing oscillation in the strongly-coupled eastern basin, and in which time scales of ocean wave dynamics provide the memory of the system essential to the oscillation.

On the other hand, there exists a large class of coupled regimes in which ocean wave dynamics is not essential to interannual oscillation. In an idealized limit (the *fast-wave limit*), coupled modes are associated with the time derivative of the SST equation, and hence referred to as *SST modes*. These do involve subsurface ocean dynamics, but the time-dependence of this component is secondary. A distorted-physics method for testing this (involving articicial multipliers on selected OGCM time derivatives) was employed in Neelin (1991) to show the relevance of this limit to oscillations in one flow regime of an HCM. Hirst (1986, 1988) and Neelin (1991) showed, by numerical and analytical methods respectively, that a number of physical processes cooperate in the destabilization of SST modes whereas they compete in terms of the direction of propagation. Propagation is essential to the period in these modes and they provide a good prototype for slowly-propagating modes in a number of intermediate models and GCMs (e.g. Anderson & McCreary 1985, Yamagata & Masumoto 1989, Meehl 1990b, Lau et al 1992).

Because the SSBH delayed-oscillator model is based on the SST equation, it was natural to hypothesize that nonpropagating SST modes away from the fast-wave limit might be perturbed by wave time scales to produce standing oscillations. Such a connection is inherent in the analysis of Wakata & Sarachik (1991) in which the relation between a propagating regime of Hirst (1988) and a standing oscillation regime is demonstrated. In an apparent contradiction, two models aimed at producing more rigorous derivations of the SSBH delayed oscillator (Cane et al 1990 plus Münnich et al 1991, MCZ hereafter; and Schopf & Suarez 1990) emphasize a rather different limit. These models also assume that the coupling occurs at a single point rather than across all or most of the basin. SST-mode solutions in the fast-wave limit allow an analytical approach to the spatial structure of the coupled modes, inclusion of several growth mechanisms, and a determination of their relation to propagating regimes, but at the cost of eliminating subsurface memory. Jin & Neelin (1993a,b) and Neelin & Jin (1993; collectively JN hereafter) outlined the complementarity between these approaches and the usefulness of analytical prototypes which include solutions for the spatial structure of the coupled modes in various limits.

INTERMEDIATE COUPLED MODEL We present here the JN "stripped-down" version of the CZ ICM, as a basis for deriving simpler models and discussing flow regimes. We nondimensionalize to bring out a few *primary parameters* from among the many lurking in the coupled system. These are:

μ: the relative coupling coefficient—strength of the wind-stress feedback from the atmosphere per unit SST anomaly, scaled to be order unity for the strongest realistic coupling; for $\mu = 0$ the model is uncoupled.
δ: the relative adjustment time coefficient—measures the ratio of the time scale of oceanic adjustment by wave dynamics to the time scale of adjustment of SST by coupled feedback and damping processes. It is scaled to be order unity at standard values of dimensional coefficients.
δ_s: surface-layer coefficient. This parameter governs the strength of feedbacks due to vertical-shear currents and upwelling, (u_s, v_s, w_s), created by viscous transfer between the surface layer and the rest of the thermocline. As $\delta_s \rightarrow 0$ the effects of these feedbacks become negligible.

A modified shallow-water model with an embedded, fixed-depth mixed layer (Cane 1979, Schopf & Cane 1983) provides the ocean-dynamics component:

$$\begin{aligned}(\delta\partial_t+\varepsilon_m)u'_m2-yv'_m+\partial_x h' &= \tau' \\ yu'_m+\partial_y h' &= 0 \\ (\delta\partial_t+\varepsilon_m)h'+\partial_x u'_m+\partial_y v'_m &= 0 \qquad (1)\end{aligned}$$

$$\varepsilon_s u_s' - y v_s' = \delta_s \tau'$$
$$\varepsilon_s v_s + y u_s' = 0, \tag{2}$$

where latitude, y, appears due to the nondimensionalized Coriolis force and the equations are applied here to departures (primed quantities), from a specified climatology (denoted by an overbar). Anomalous vertical mean currents above the thermocline, (u'_m,v'_m), and thermocline depth, h', are governed by the shallow-water component in the long-wave approximation (1), with suitable boundary conditions at basin boundaries (Gill & Clarke 1974); vertical shear currents, (u'_s,v'_s), are governed by local viscous equations (2). Both are driven by the zonal wind stress anomaly, τ'. The damping rates, ε_m and ε_s are not treated as primary parameters because the former is small and the latter can be largely absorbed into δ_s. For a more formal scaling see JN; for justification of several approximations, see Cane (1979) and CZ. Vertical velocities are given by the divergence of the horizontal velocities and the values of surface currents and upwelling into the surface layer by the sum of anomalous mean and shear contributions plus the climatology: $u = \bar{u} + u'_m + u'_s$, $w = \bar{w} + w'_m + w'_s$.

Because SST serves as a key interfacial variable, careful parameterization of processes that affect SST are largely responsible for the success of the CZ model. The direct effects of temperature variations in the surface layer on pressure gradients are neglected in (1), but a prognostic equation for SST is carried separately which contains all the essential nonlinearity of the CZ model:

$$\partial_t T + u\partial_x T + \mathrm{H}(w)w(T - T_{sub}) + v\partial_y T + \varepsilon_T(T - T_0) = 0 \tag{3}$$

in nondimensional form. Here T is total SST and H is an analytic version of the Heaviside function due to upstream differencing into the surface layer. The Newtonian cooling represents all physical processes that bring SST towards a radiative-convective-mixing equilibrium value, T_0. The subsurface temperature field, T_{sub}, characterizes values upwelled from the underlying shallow-water layer and is parameterized nonlinearly on the thermocline depth—deeper thermocline results in warmer T_{sub}. Motivated by the fact that the strongest SST response to upwelling, advection, and thermocline depth change are confined to a fairly narrow band along the equator for the phenonena of interest, Neelin (1991) applied this equation to the SST in an equatorial band, where each of the variables in (3) need only be evaluated at the equator, and where the $v\partial_y T$ term is replaced by a suitable upstream differencing. In the JN ICM, this captures all the essential behavior of the CZ model while permitting a number of analytical results to be generated in special cases.

The simple atmospheric models that provide a zeroth-order approximation to the wind-stress response to SST anomalies can be written

$$\tau' = \mu A(T'; x, y), \tag{4}$$

where μ is the coupling coefficient and $A(T'; x, y)$ is a linear but nonlocal function of T' over the entire basin. For a Gill (1980) model with a specified meridional profile of the forcing appropriate to the assumed SST y-dependence, A is a simple integral operator.

Coupling is carried out by a version of flux correction: running the ocean model with observed climatological wind stress to define the ocean climatological state ($\bar{u}$, $\bar{w}$, $\bar{T}$, . . .), then defining SST anomalies, T', with respect to this. A known climatological solution to the coupled system is thus constructed. For sufficiently small coupling, this state is unique and stable; interannual variability must arise by bifurcations from this state as μ increases.

USEFUL LIMITS We introduce terminology for limits that are useful for understanding how coupled modes relate to simpler cases and for comparing various theoretical models. The *weak-coupling limit* is reached at small μ, i.e. little wind-stress feedback per unit SST anomaly; these modes are found not to be good prototypes for fully coupled modes. At large μ, one obtains the *strong-coupling limit*. When the time scale of dynamical adjustment of the ocean is small compared to the time scale of SST change by coupled processes (i. e. small δ), one has the *fast-wave limit*; which is very useful for generating analytical results that provide understanding of spatial structure and growth characteristics. The *fast-SST limit* is reached at large δ; this is the converse to the fast-wave limit, i.e. sea surface temperature adjusts quickly compared to ocean dynamical processes.

In the uncoupled case or in the fast-wave limit, the modes of the ICM, linearized about its climatology, separate into a set associated purely with the time derivatives of the shallow-water equations, referred to as *ocean-dynamics modes,* and a set associated purely with the time derivative of the SST equation, referred to as *SST modes*. In an uncoupled, zonally-bounded basin, the ocean-dynamics modes consist of a set of ocean-basin modes (Cane & Moore 1981) and a scattering spectrum (JN). At low frequencies and basin scales, the ocean-dynamics modes are very different from the Rossby and Kelvin modes of the infinite-basin case. In the coupled system, the distinction between corresponding coupled modes is maintained in the idealized fast-wave and fast-SST limits, but in most of the parameter space the coupled modes will have a mixed nature, for which we use the descriptive term *mixed SST/ocean-dynamics modes* when it is necessary to be specific; otherwise the term *coupled modes* is taken to imply this.

WIND-DRIVEN OCEAN RESPONSE Extensive theory exists for the adjustment of the uncoupled shallow-water ocean to time-varying winds (see Moore & Philander 1978, Cane & Sarachik 1983, McCreary 1985 for reviews). Much of it is phrased in terms of adjustment to abrupt changes in wind; a much better prototype for understanding interannual coupled oscillations is the case of forcing by low-frequency, time-periodic winds (Cane & Sarachik 1981, Philander & Pacanowski 1981). Figure 6 shows a time-longitude plot of thermocline perturbations along the equator for such a case. The western Pacific leads the eastern Pacific by between 90 and 180

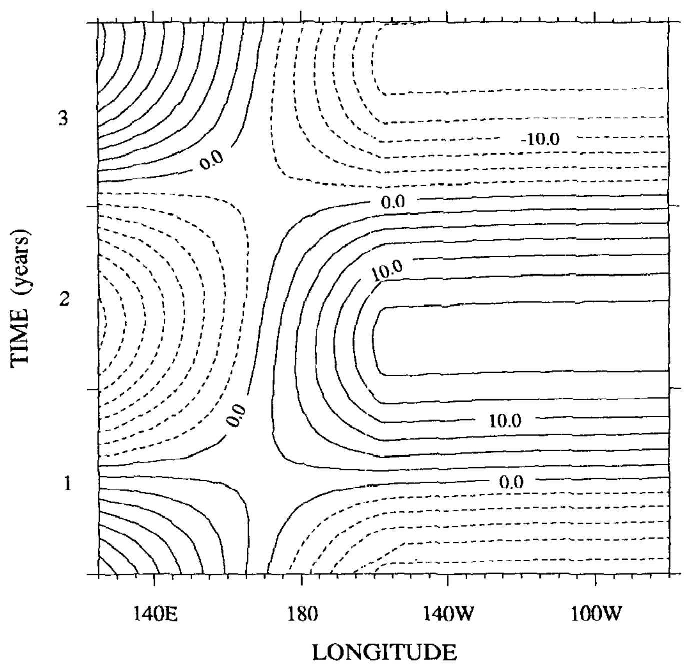

Figure 6 Time-longitude plot of thermocline depth anomalies along the equator from a shallow-water ocean model forced by specified wind stress: constant in longitude through the western half of the basin, zero in the eastern half, Gaussian in latitude (scale 5°), and periodic in time with period 3 years and amplitude 0.2 dyne/cm^2. Following Cane & Sarachik (1981) with specified modifications, and with frictional damping of time scale 0.5 yr in the ocean. Contour interval 2.5 m.

degrees in temporal phase. The fast-wave limit case, in which the ocean approaches equilibrium with the wind, would correspond to 180° phase difference between these, which would remove the apparent slow eastward "propagation." It should be emphasized that this is not a wave propagation in the sense of any individual free wave of the system, but rather the sum total of the ocean response which is not quite in equilibrium with the wind forcing. The slight departures from equilibrium, as measured by the difference from a 180° phase lag, characterize the oceanic memory which is so important to interannual variability.

SIMPLE MODELS: POINT COUPLING One special case where solutions of (1) can be carried forward is if the surface-layer feedbacks are dropped ($\delta_s = 0$) and if it is assumed that *coupling to the atmosphere occurs at a single point*, e.g. at the eastern boundary. The wind-stress magnitude is taken to be proportional to SST at that point and the spatial form of wind stress is fixed, for instance to be a patch of very small longitudinal extent (here placed at mid-basin for ease of presentation), with Gaussian y-dependence of curvature α. We give here an extended version of the MCZ model (or Schopf & Suarez 1990) using this approximation; the SST equation (3) and ocean shallow-water dynamics (1) can be reduced to

$$\delta^{-1}\partial_t T' + [(T' - T'_{\mathrm{sub}}(h')] = 0 \tag{5a}$$

$$h'(t) = \sum_{j=1}^{\infty} a_j(\varepsilon_{\mathrm{m}})h'(t-4j) + \mu \sum_{j=0}^{\infty} b_j(\alpha, \varepsilon_{\mathrm{m}})T'(t-1/2-2j), \tag{5b}$$

where T' and h' are SST and thermocline depth anomalies at the eastern point, respectively, and the coefficients a_j, b_j summarize information about the ocean dynamics, boundary conditions, and parameters. The reduction of ocean dynamics to sums over discrete transit times results, of course, from the point-wise coupling assumptions. In contrast to (1), here time has been normalized by the time scale characterizing ocean dynamics (the Kelvin-wave basin-crossing time), so δ appears in the SST equation and the integer lag dependences on the past history of h' and T' are due to wave transit times across the basin with reflection at basin boundaries. This rescaling is because the MCZ model has been used primarily in the fast-SST limit ($\delta \rightarrow \infty$), which results in dropping the time derivative in (5a). In this case, (5) becomes an iterated map of high order for modes related to the time derivatives of the shallow-water equations, which yielded the lags.

A simpler delay equation which has proven influential in the field can be derived from the above by a series of simplifications which cannot be rigorously justified but which retain essential features of the dynamics:

(*i*) Set all $a_j = 0$ in (5b) while retaining the b_j, which amounts to removing eastern-boundary wave reflections while keeping those at the western boundary. This does irreparable damage to the uncoupled ocean dynamics, so the usefulness of this simplification depends on the coupling dominating the spatial structure of the mode—we show below how this comes about in an ICM. (*ii*) Move the wind stress to the point of coupling to SST (the position of the eastern boundary is now immaterial); this is reasonable when the coupled frequency is much less than the Kelvin transit time across the separation. (*iii*) Truncate the sum over b_j to only two terms—a single westward Rossby wave plus an eastward Kelvin wave. Because the series converges slowly, this can only be justified qualitatively. We present the model linearized about the climatology, since this determines the period and the location of the bifurcation:

$$\partial_t T' = (\mu b_0 - 1)T'(t) - \mu b_1 T'(t - 4\delta) \tag{6}$$

This is the *SSBH delayed-oscillator model*. Note that we have restored the time nondimensionalization used in (1–3). Nonlinear versions are straightforward to derive from (5) and differ significantly from that given by SSBH for more realistic $T_{sub}(h')$; $dT_{sub}/dh'|_0 = 1$ is used without loss of generality for this simple case. When the model is uncoupled ($\mu = 0$), the solution is a purely decaying mode whose eigenvalue is determined by the SST equation, i.e. an SST mode, in contrast to the MCZ model which also has uncoupled ocean modes. In the fast-wave limit ($\delta \to 0$), the model has stationary (i.e. nonoscillatory) SST modes which become unstable for coupling above $\mu = (b_0 - b_1)^{-1}$. For realistic values of δ, this unstable mode becomes oscillatory due to the adjustment time scales of subsurface ocean dynamics, here represented by a single delay. The SSBH model may thus be summarized as an SST-mode whose growth can be understood from the fast-wave limit, perturbed to give oscillation by aspects of ocean dynamics which are *not* characteristic of the uncoupled case. We will show below that this interpretation can be carried over to an ICM.

On the other hand, consider the modes of the extended MCZ model (Equation 5), linearized about climatology, with time dependence $\exp(\sigma t)$. The infinite series in (5b) can be summed exactly (Cane & Sarachik 1981) under certain conditions (note the contrast to the severe truncation of the SSBH model which has sometimes been interpreted too strongly in terms of individual waves). Equation (5b) becomes, in the simplest case:

$$[\sinh(2\sigma)/\sinh(\sigma)]h' = \mu T' \tag{5b$'$}$$

In the fast-SST limit, there is a singularity at $\mu = 2$, with two equal eigenvalues, demarcating the boundary between oscillatory eigenvalues below and stationary eigenvalues above, one of which is strongly growing.

In this case, the singularity leads to a codimension 2 (double-zero) bifurcation. When any other destabilizing process is added, the bifurcation is oscillatory (Hopf); with damping and no other destabilizing process it is stationary (transcritical)—but note that stationary bifurcations must be treated with caution since they are not robust to relaxation of assumptions used to construct the climatological state.) The oscillatory case is the one that applies to ENSO, but it is worth asking why the ocean dynamics "break" from oscillatory behavior, as would be expected of wave-related modes in a bounded basin, into a growing stationary mode. Consider the case where coupled feedback processes are very strong; then local interactions dominate nonlocal wave-propagation processes yielding pure growth—the mode grows too fast to be affected by weak return signals from the western boundary. The transition has to occur at moderate coupling. The remarkable feature which will be shown in an ICM below is that the stationary mode, even in the fast-SST limit, shares more characteristics with the SST mode in the fast-wave limit than with the uncoupled ocean modes. In fact, at strong coupling, the stationary mode eigensurface is continuously connected across the whole range of δ.

SIMPLE MODELS: FAST-WAVE LIMIT Although the time scale of subsurface dynamics is the dominant factor in setting the period of ENSO, the strong simplifications that occur in the fast-wave limit permit insight into spatial structure. Setting $\delta = 0$ in (1), i.e. assuming that oceanic adjustment occurs fast compared to other time scales, and considering that the damping ε_m is very weak, reduces the shallow-water equations to Sverdrup balance along the equator:

$$\partial_x h' = \tau' \tag{7}$$

with negligible vertical mean currents. The off-equatorial ocean solution plays a significant role which can be summarized in boundary conditions to (7) suitably derived as the limit to wave adjustment processes, as discussed in JN and in Hao et al (1993), both of which provide further analysis of this fast-wave limit case. The multiple coupled feedback mechanisms can be seen from a linearized version of the SST equation (3), with h' given by combining Equations (7) and (4) (see Neelin 1991).

A number of physical mechanisms contribute to destabilization of SST modes. However, these mechanisms tend to compete in terms of whether the mode will be purely growing or will propagate slowly along the equator. For instance, a warm SST anomaly will lead to westerly wind anomalies above and to the west of the warm SST, which will lead to eastward current anomalies and reduced upwelling and thus to a warming of SST which will both enhance the original anomaly and cause it to shift slightly westward. In the ICM, these feedbacks are controlled by δ_s. On the other hand, the thermocline slope will tend to be reduced below and slightly to the east of the SST anomaly. The subsurface waters being carried to the surface will be warmer than normal, thus tending to enhance the initial anomaly. Since both the thermocline response and wind-stress response are nonlocal, the shape of the anomalies will evolve to satisfy basin boundary conditions, leading for a broad range of parameters to a stationary growing mode. Analytical results for both propagating and stationary cases can be obtained in the fast-wave limit (JN); Hao et al 1993 give nonlinear solutions. The mode with the largest spatial scale in the basin is always the most unstable, with SST and wind structure similar to observations. The larger SST anomalies in the east and central basin are produced partly by the shape of the climatological upwelling, and partly by purely dynamical effects, with east-west asymmetry introduced by the latitudinal derivative of the Coriolis force. The analytical results also indicate the role of the eastern boundary in keeping the mode from propagating; the point-coupling models emphasize the role of the western boundary on the ocean—but for spatial structure, eastern boundary effects enter mainly through the atmosphere.

The feedback loop described above sounds very similar to that described in the Bjerknes hypothesis. It gives the mechanism maintaining interannual variability and, suitably extended by the analytical results, the spatial form. However, it only gives the interannual period in regimes with coherent zonal propagation along the equator. This is a good prototype for the slowly-propagating modes in a number of coupled models (e.g. Meehl 1990b, Lau et al 1992), but to understand how this mode relates to the observed system, it is essential to see how time scales of subsurface ocean dynamics perturb it in the vicinity of the stationary regime to produce oscillations with standing-oscillatory SST anomalies.

PARAMETER DEPENDENCE OF LEADING MODES IN AN ICM A global picture of the connection of coupled modes in the ICM (1)–(3) can be delineated by tracing the behavior of the few leading (fastest-growing or slowest-decaying) eigenmodes as a function of parameters μ and δ, beginning with $\delta_s = 0$ for simplicity. In the fast-wave limit ($\delta = 0$), a stationary (i.e. purely growing) SST mode becomes unstable. Its spatial structure is suggestive: It looks like the warm phase of Figure 4, except that the thermocline component has eastern and western parts of the basin exactly out of phase, so there is no oscillation. As one moves from the fast-wave limit to realistic relative-time-scale ratios (larger δ) one finds that this stationary eigenmode is scarcely changed. In fact, for coupling values stronger than a certain threshold (where coupled processes dominate those associated with oceanic wave propagation, as discussed above), the eigensurface extends without substantial change from the fast-wave limit all the way to the fast-SST limit. This is pivotal in understanding the coupled system because 1. it allows the spatial form and growth mechanisms of important coupled regimes to be understood from the fast-wave limit; and 2. it implies that modes associated with ocean dynamics must connect somehow to this strongly-growing mode.

To illustrate how this happens, Figure 7 shows a typical slice through parameter space as a function of coupling, for a realistic value of δ. The eigenvalues of the five leading modes are plotted as dots on the complex plane (growth-rate, frequency), for evenly spaced coupling values in the range $\mu = 0$ to 0.8. Left-right symmetry occurs because oscillatory modes always exist as conjugate pairs. The strongly growing stationary (i.e. non-oscillatory) branch in the strong-coupling range is indicated as "stationary regime" on the figure, since it is the only unstable mode in this parameter range. This is the mode that is so closely related to the SST mode in the fast-wave limit. At a slightly lower coupling value a singularity occurs where this is converted into an oscillatory mode—this singularity (corresponding to a codimension-2 double-zero bifurcation) extends as a curve in the μ-δ parameter plane connecting the eigensurface associated with the gravest SST mode to surfaces that are associated with ocean dynamics modes at low coupling. To the lower-coupling side of this singularity one finds the regime with oscillations that have a standing SST component (denoted in the figure as "standing-SST oscillatory regime'; SSO regime), corresponding to that shown in Figure 4 for the CZ ICM. The spatial form is similar to the stationary SST mode, and the mode is destabilized by the same coupling mechanisms, but subsurface oceanic dynamics provide the memory for the oscillation, characterized by temporal phase lag of the thermocline across the basin as in observations (Figure 3). This regime extends across a large range of δ, from $\delta = O(1)$ to the fast-SST limit.

In contrast, the connection of this standing-SST oscillation regime to the uncoupled case is complicated. In Figure 7, the SSO regime eigensurface eventually connects to one of the modes from the discretized scattering spectrum, but as it does so the mode rapidly changes in spatial form. When one includes variations in δ, one finds that the standing-oscillation regime connects, not to a single mode from the uncoupled oceanic dynamics spectrum, but to a series of them: The low coupling end of the branch attaches first to the lowest-frequency scattering mode (as in Figure 7), then to sequentially higher-frequency scattering modes, and finally at large δ, near the fast-SST limit, it connects to the gravest ocean basin mode, much as in the MCZ point-coupling model. These successive connections are

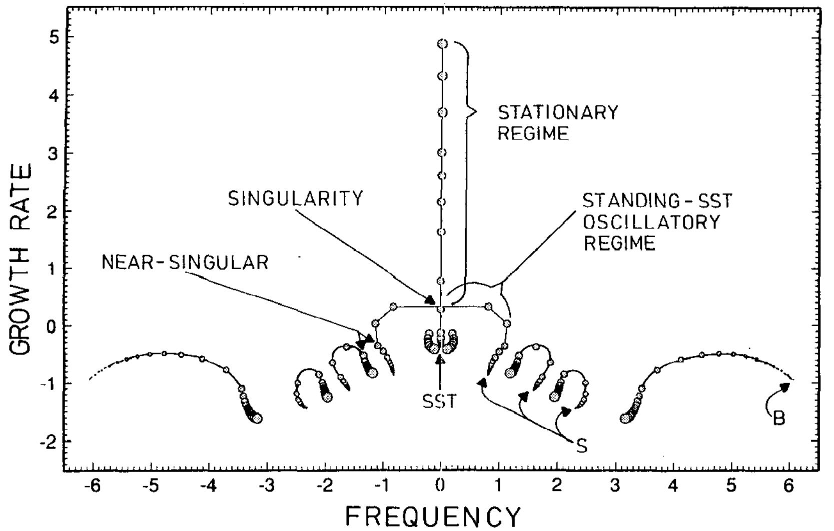

Figure 7 Eigenvalues of the five leading modes of the Jin & Neelin (1993a) intermediate coupled model as a function of coupling coefficient, μ, for a realistic value of the relative time scale coefficient, $\delta = 1.5$. Dots give frequency and growth rate of each mode on the complex plane, with dot size representing coupling for constant increments from $\mu = 0$ to $\mu = 0.8$. Eigenvalues trace out continuous paths as a function of coupling (indicated by interpolated lines for clarity). Uncoupled modes (ocean basin mode B, discretized scattering modes S, and an SST mode) are indicated at $\mu = 0$ (smallest dots). The modes have mixed character for larger μ: The purely growing mode which produces the *stationary regime*, indicated over the range of large μ, is closely related to the stationary SST mode; this is connected at a singularity to the important *standing-SST oscillatory regime* which extends over a range of moderate coupling values.

accomplished by a sequence of additional singularities; a "near-singular" point is shown, where the SSO regime connects to the next scattering mode at larger δ. However as δ varies, the characteristics of the SSO regime are almost completely insensitive to which uncoupled ocean mode it is attached to; its properties are fundamentally deterined by the coupling and it is thus best approached conceptually from the strong-coupling side.

It is thus much simpler to view the standing-oscillation regime as an extension of the strongly-growing stationary mode towards lower coupling, where ocean dynamics begin to regain some aspects of wavelike behavior. In this interpretation, one begins by understanding the spatial form and instability mechanisms of the mode in the fast-wave limit at fairly strong coupling. As one follows the stationary mode out to realistic values of the relative time-scale parameter and down to moderate coupling, it retains its form but acquires a frequency associated with "picking up" a degree of freedom from among the low-frequency part of the scattering spectrum on the low-coupling side. This view is consistent, in terms of physical content, with the original interpretation of the SSBH delayed oscillator model (Equation 6) (as long as it is understood that the subsurface memory is not associated with individual waves). Furthermore, the smooth connection from the fast-wave limit to the fast-SST limit implies that the seemingly contradictory approaches to the problem represented by (6) and (5b′) are just alternative approximations to the same eigensurface.

Finally, to make the connection to propagating regimes such as occur in some of the models, which may be relevant to the differences in evolution of certain ENSO events, consider reintroducing a third parameter, such as δ_s. As this changes, it is easy to move smoothly and gradually from the standing-oscillation regime to a regime of the mixed SST/ocean-dynamics modes where propagation occurs during parts of the cycle and contributes to the period (JN, Kleeman 1993). The standing-oscillation regime provides the clearest case emphasizing the role of subsurface dynamics in determining periodicity; the fast-wave-limit propagating cases provide alternate simple cases in which periodicity is provided by zonal phase lags. Between these continuously connected regimes, both characteristics can coexist within the same coupled mode. There is thus no contradiction between evidence for importance of subsurface dynamics in the ENSO cycle and indications of other contributing mechanisms.

TRANSITIONS TO IRREGULARITY The modeling consensus is thus that ENSO dynamics are fundamentally oscillatory. In particular, for models whose uncoupled components have no internal variability, interannual variability arises as a forward Hopf bifurcation of the coupled system, yielding a limit cycle. The obvious question is then the source of irregularity in the observed cycle: (*a*) transition to chaotic behavior by higher bifurcations associated with the coupled dynamics, and/or (*b*) stochastic forcing by atmospheric "noise" from shorter-lived phenomena which do not depend on coupling?

With regard to internal dynamics, CZ pointed out early on that their model achieved a degree of irregularity through deterministic coupled dynamics alone. Disagreement by Battisti & Hirst (1989) over whether this was due to the CZ numerical implementation seems to have been settled in the larger picture in favor of the original finding; for instance, the smoothly posed, simpler version of JN also possesses irregular regimes. Explicit discussion of the bifurcation structure of the coupled system and secondary bifurcations to regimes of complex behavior was given in an HCM in Neelin (1990), but the first clear demonstration of a bifurcation sequence into chaotic behavior was given by MCZ in the point-coupling model (5). An unfortunate footnote must be added for Vallis (1986), who attempted to raise these questions in an ad hoc model (now thought to lack essential physics) and instead illustrated the dangers of spurious chaos due to highly-truncated numerics (Vallis 1988). As to the scenario for the transition to chaos, MCZ cite the Ruelle-Takens-Newhouse scenario (e.g. Eckmann 1981) as a possibility, based upon their observing irregular behavior subsequent to one period doubling. However, it is clear that the presence of parametric forcing by the annual cycle plays an important role in the prevalence of chaotic regimes in parameter space (Zebiak & Cane 1987, MCZ) and in the widespread frequency-locked regimes (CZ, Battisti & Hirst 1989, Schopf & Suarez 1990, Barnett et al 1993). A "Devil's staircase" scenario (e.g. Jensen et al 1984) is among current postulates (F.-F. Jin et al, personal communication; E. Tziperman et al, personal communication).

On the stochastic forcing side, early discussions of ENSO were often phrased in terms of random wind events initiating an El Niño warm phase. Among modelers this has given way to the view that stochastic forcing more likely disrupts the cycle (or possibly excites a weakly-decaying oscillatory mode, if below the Hopf bifurcation). Zebiak (1989) indicates that such effects have only a minor impact in the CZ ICM; Latif & Villwock (1990) and T. P. Barnett et al (personal communication) indicate that the effects of randomized atmospheric forcing can be considerable on an uncoupled tropical ocean model. Problems in quantifying the importance of stochastic effects involve estimation of spatial coherence, which is extremely important to ocean response and to separation of the stochastic component from the atmospheric deterministic response to SST.

Prediction and Predictability

The underlying periodic aspects of ENSO and the above theoretical considerations imply a good deal of ENSO predictability. A hierarchy of ENSO prediction schemes has been developed which includes statistical and physical models (Inoue & O'Brien 1984, Cane et al 1986, Graham et al 1987a,b, Xu & von Storch 1990, Goswami & Shukla 1991, Keppene & Ghil 1992, Barnston & Ropelewski 1992, Latif et al 1993b, Penland & Magorian 1993). A more complete list of references can be found in the review papers by Barnett et al (1988) and Latif et al (1993c). The most succesful schemes, the coupled ocean-atmosphere models, show significant skill in predicting ENSO even at lead times beyond one year. Figure 8 shows the anomaly correlation of the observed with the predicted SST anomalies averaged over the region of greatest variability for the CZ ICM—the first coupled model used for ENSO forecasts. Comparable

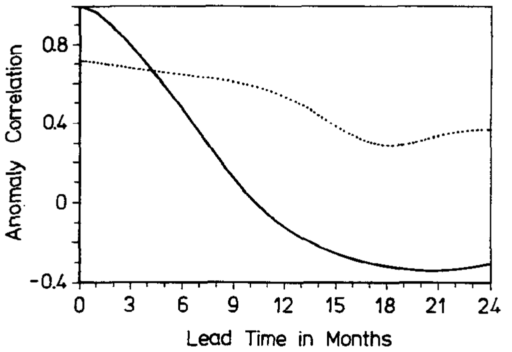

Figure 8 Skill scores of prediction ensembles as a function of lead time for forecasts by the Cane & Zebiak (1985) intermediate coupled model (*dotted curve*), compared with skill obtained by assuming persistence of anomalies (*solid curve*). The measure is correlation of predicted and observed SST averaged over the region of largest ENSO anomalies ($\pm 5°$ latitude, 150°W to 90°W longitude), during the period 1972 to 1991. Note that 0-month lead can differ from observed because SST data are not used in the initialization. Data from S. Zebiak (personal communication); for methodology see Cane et al (1986).

results have recently been obtained with CGCMs (Latif et al 1993b) and HCMs (Barnett et al 1993).

At lead times of a few months, the coupled models do not beat the persistence forecast which assumes that the SST anomalies remain constant throughout the forecast period. This is due to the fact that up to present no ocean observations are used in the initialization of the coupled models. Instead, the observed wind stresses are used to spin up to the ocean component, but errors in the forcing and the model formulation manifest themselves as considerable errors in the initial SST anomaly fields. Thus significant improvement of the forecasts at small lead times can be achieved by assimilating in situ ocean obervations (e.g. Leetmaa & Ji 1989) which are becoming increasingly available (e.g. Hayes et al 1991), and/or observations from space (e.g. Tai et al 1989). In the case of coupled GCMs, a further reduction of climate drift will greatly aid this process. Upper limits on predictability are an area of current investigation (Blumenthal 1991, Goswami & Shukla 1991, Keppenne & Ghil 1992).

THE EXTRATROPICAL PROBLEM

Already in the late 1950s and early 1960s possible large-scale air-sea interactions in midlatitudes over both the Pacific and Atlantic Oceans were described by several authors (e.g. Namias 1959, Bjerknes 1964). Theoretical work by Hasselmann (1976) showed that the ocean can convert the white noise forcing by the atmosphere into a red noise SST spectrum through its large heat capacity. Such low-frequency SST anomalies can potentially feed back onto the atmospheric circulation. Recent modeling results suggest that the midlatitudinal atmosphere may indeed be significantly influenced by midlatitudinal SST anomalies, especially on interdecadal time scales (e.g. Hense et al 1990). The response characteristics, however, appear to be much more complex than in the tropics. Since the understanding of the extratropical problem is still at a rather low level, we restrict ourselves to pointing out four of the most important differences from the tropical problem.

First, the midlatitudinal circulation is influenced not only by midlatitudinal but also by tropical SST anomalies, as shown in many observational and modeling studies (e.g. Shukla & Wallace 1983, Lau 1985). A characteristic response pattern, the Pacific/North-America pattern, describing the response of the atmospheric winter circulation to tropical Pacific SST anomalies associated with the extremes in the ENSO cycle, has been identified (Wallace & Gutzler 1981) and exploited for short-range climate predictions for the North Pacific/North American region (Barnett & Preisendorfer 1987).

Second, both atmosphere and ocean have a much higher level of uncoupled internal variability. Uncoupled ocean models can produce decadal- or centennial-scale variations (e.g. Weaver et al 1991, Mikolajewicz & Maier-Reimer 1990). The effect of slowly-varying ocean boundary conditions on the atmosphere can be overwhelmed by the atmospheric noise level; for instance, Lau (1985), comparing AGCM runs with observed and climatological SST, found that observed SST variations produced a significant increase in atmospheric variability only in the tropics, while the midlatitudinal atmosphere exhibits a realistic level of interannual variability even in the case with climatological SST (Lau 1981b). Both effects make it difficult to assess the role of coupling on observed interannual to interdecadal variability (e.g. Gordon et al 1992). For instance, Delworth et al (1993) provide an analysis in a coupled GCM integration of Atlantic interdecadal variability involving changes in the overturning thermohaline circulation and advection-induced changes in density. While these phenomena have signatures in SST and air temperature, they are hypothesized to be uncoupled oceanic phenomena, driven by stochastic forcing from the atmosphere.

Third, the response of the general circulation in midlatitudes to SST anomalies (tropical and extratropical) is highly nonlinear (Kushnir & Lau 1992), while the response of the tropical atmosphere can be approximated by linear dynamics (e.g. Gill 1980). Experiments with general circulation models provide an opportunity to further investigate the relationship between extratropical SST anomalies and atmospheric flow regimes. Palmer & Sun (1985) investigated the reponse of the atmosphere to SST anomalies in the northwestern Atlantic. They showed that the model response was consistent with data and concluded that a positive feedback between ocean and atmosphere is possible during certain times of the year which might contribute to persistent climate anomalies. Some evidence of impacts of extratropical SST anomalies on the general circulation is also provided by Lau & Nath (1990) and Kushnir & Lau (1992) but the relationships between midlatitudinal SSTs and atmospheric indices appear to be far more complicated than in the tropics, in part due to the importance of transient disturbances to the time-averaged response.

Finally, direct effects of local air-sea heat exchange on the ocean play a more active role at midlatitudes than in the tropics where SST anomalies result primarily from variations in the surface wind stress. Persistent large-scale midlatitudinal SST fluctuations can be identified in Atlantic, Pacific, and global time series (Wallace & Jiang 1987, Wallace et al 1990, Folland et al 1991, Ghil & Vautard 1991). These anomalies may be driven by anomalies in the surface heat flux (e.g. Alexander 1992a, Cayan 1992), at least on monthly-to-interannual time scales. Kushnir (1993) argues that ocean circulation is important on longer time scales. Part of the interannual variability of SST in the North Pacific is linked to the ENSO phenomenon (Weare et al 1976, Luksch et al 1990, Alexander 1992a) and results from changes in the atmospheric circulation over the North Pacific in response to tropical SST anomalies. During an El Niño (warm) phase, for instance, an anomalous low-pressure system develops over the North Pacific, thereby strengthening the Aleutian Low. The changes in surface wind stress and more importantly those in surface heat flux force negative SST anomalies in the central North Pacific; the reverse occurs during an ENSO cold phase. The temperature near the American west coast tends to vary in phase with the tropical anomaly and is probably related in part to coastal Kelvin waves, which are generated by the reflection of equatorial Kelvin wave packets. Anomalous warm air advection in response to the strengthening of the Aleutian Low also plays a significant role in the generation of these anomalies. Pitcher et al (1988) show that these North Pacific SST anomalies can contribute a considerable atmospheric response; on the other hand, Alexander (1992b) finds that the local ocean-atmosphere feedback tends to act as a damping on the North Pacific SST response to ENSO.

SUMMARY AND DISCUSSION

The past decade has seen our knowledge of ocean-atmosphere interaction for the tropical problem go from the level of hypothesis to that of a field with rapidly developing theoretical and numerical modeling components. Theory for the El Niño/Southern Oscillation phenomenon has reached a relatively mature level for understanding the mechanisms contributing to the maintenance and period of the cycle, as they relate to the primary bifurcation from the climate state in models of different levels of complexity. The relationship between several regimes of interannual variability found in models has been largely understood, as has the complementary relationship between simple prototypes for the modes of coupled variability. These illustrate both the importance of subsurface ocean dynamics in providing the memory of the system, and the fundamental impact of coupling in determining the spatial character of these modes. The exact mechanism of the two apparent time scales in the ENSO signal and the dominant sources of irregularity in the cycle are not yet understood, although hypotheses have been posed in terms of the higher bifurcations of the coupled system or stochastic forcing due to uncoupled variability.

Models that capture the primary bifurcation in a realistic regime have been used to skillfully predict ENSO-related tropical Pacific SST anomalies at lead times out to a year. The potential for predictability beyond this is not yet known; a major area of current endeavor is ascertaining to what degree such tropical predictability can translate into useful midlatitude climate predictions on seasonal-to-interannual time scales.

Simulation of tropical climate and ENSO-related variability with coupled GCMs is improving at a rapid rate. The climate drift and variety of regimes of variability in earlier versions of these models are characteristic of the sensitivity of the coupled system and provide an apt demonstration that a coupled model is more complex than the sum of its uncoupled components. Because three-dimensional feedbacks tended to exacerbate relatively small errors in physical parameterizations in some of the early versions, small improvements in these parameterizations have in several cases provided highly encouraging improvements in simulation. This rapid learning curve for the tropical problem is partly the result of not needing to explicitly simulate the global thermohaline circulation which maintains the deep-ocean temperature and salinity through high-latitude convective sinking. Coupled GCMs for phenomena involving this circulation may have a longer development time to achieve accurate simulation without flux correction.

Exciting new areas within the tropical problem include: ocean-atmosphere interactions within the Atlantic and Indian basins, multi-basin interactions, and possible interactions with neighboring land processes (e.g. Southeast Asian and Indian Monsoon circulations, Tibetan plateau snow cover, Sahel rainfall, and South American convergence zones). Monsoon-ENSO interactions have already received considerable speculation (e.g. Yasunari & Seki 1992, Webster & Yang 1992); given the complexity of coupled processes in the tropical Pacific basin alone, unraveling next-order linkages to other subsystems will be a true challenge to models at all levels in the hierarchy. Circumstantial evidence from the coupled GCMs suggests the importance of coupled interactions in maintaining major features of the tropical climate and seasonal cycle, for instance, the warm-pool/cold-tongue configuration in the Pacific, and that the mechanisms involved may be qualitatively similar to those active in interannual variability. Developing a theoretical understanding of how these apply to the climatology would be a valuable asset both from a conceptual point of view, and for distinguishing the plausible from the speculative in tropical aspects of global-change scenarios.

While the midlatitude coupled problem is complicated by large internal variability of both atmosphere and ocean, there is reason to hope that the enthusiasm and experience that have accumulated for coupled interactions in the tropics will be carried to higher latitudes. There is growing attention to internal climate variability at decadal and longer time scales both in the tropics and extratropics, due to its importance in the problem of detection of anthropogenic warming and as a new frontier in simulation and theory. This will no doubt lead to a plethora of hypothesized mechanisms which may take decades to refute or verify due to the lack of long observational time series of dynamically-important quantities. Nonetheless, we can look forward to the need for a review of coupled ocean-atmosphere dynamics for the extratropical problem and new aspects of the tropical problem within a relatively few years.

Acknowledgments

Support for this study has been provided by National Science Foundation ATM-9215090, Presidential Young Investigator award ATM-9158294, National Oceanographic and Atmospheric Administration NA16RC0178/26GP00114-01, CEC Environmental Programme EV5V-CT-0121, the Max-Planck-Institut für Meteorologie, and the Max-Planck Society (partial support during a sabbatical leave by JDN at MPIM). We thank M. Grunert and W. Weibel for assistance with graphics, and many colleagues for constructive comments.

Literature Cited

Alexander, M. A. 1992a. Midlatitude atmosphere-ocean interaction during El Niño. Part I: the North Pacific. *J. Climate* 5: 944–58

Alexander, M. A. 1992b. Midlatitude atmosphere-ocean interaction during El Niño. Part II: the Northern hemisphere atmosphere. *J. Climate* 5: 959–72

Allen, M. R., Davey, M. K. 1993. Empirical parameterization of tropical ocean-atmosphere coupling: the "inverse Gill problem." *J. Climate* 6: 509–30

Anderson, D. L. T., McCreary, J. P. 1985. Slowly propagating disturbances in a coupled ocean-atmosphere model. *J. Atmos. Sci.* 42: 615–29

Barnett, T. P. 1983. Interaction of the monsoon and Pacific Trade Wind system at interannual time scales. Part I: the equatorial zone. *Mon. Weather Rev.* 111: 756–73

Barnett, T. P. 1991. The interaction of multiple time scales in the tropical climate system. *J. Climate* 4: 269–85

Barnett, T. P., Graham, N., Cane, M., Zebiak, S., Dolan, S., O'Brien, J., Legler, D. 1988. On the prediction of El Niño of 1986–1987. *Science* 241: 192–96

Barnett, T. P., Latif, M., Graham, N., Flügel, M., Pazan, S., White, W. 1993. ENSO and ENSO-related predictability. Part I: prediction of equatorial Pacific sea surface temperature with a hybrid coupled ocean-atmosphere model. *J. Climate* 6: 1545–66

Barnett, T. P., Latif, M., Kirk, E., Roeckner, E. 1991. On ENSO physics. *J. Climate* 4: 487–515

Barnett, T. P., Preisendorfer, R. 1987. Origins and levels of monthly forecast skill for United States surface air temperatures determined by canonical correlation analysis. *Mon. Weather Rev.* 115: 1825–50

Barnston, A. G., Ropelewski, C. F. 1992. Prediction of ENSO episodes using canonical correlation analysis. *J. Climate* 5: 1316–45

Battisti, D. S. 1988. The dynamics and thermodynamics of a warming event in a coupled tropical atmosphere/ocean model. *J. Atmos. Sci.* 45: 2889–919

Battisti, D. S. 1989. On the role of off-equatorial oceanic Rossby waves during ENSO. *J. Phys. Oceanogr.* 19: 551–59

Battisti, D. S., Hirst, A. C. 1989. Interannual variability in the tropical atmosphere/ocean system: influence of the basic state, ocean geometry and nonlinearity. *J. Atmos. Sci.* 46: 1687–712

Berlage, H. P. 1957. Fluctuations in the general atmospheric circulation of more than one year, their nature and prognostic value. *K. Ned. Meteor. Inst. Meded. Verh.* 69: 152 pp.

Bjerknes, J. 1964. Atlantic air-sea interaction. *Adv. Geophys.* 10: 1–82

Bjerknes, J. 1969. Atmospheric teleconnections from the equatorial Pacific. *Mon. Weather Rev.* 97: 163–72

Blumenthal, M. B. 1991. Predictability of a coupled ocean-atmosphere model. *J. Climate* 4: 766–84

Bretherton, F. P. 1982. Ocean climate modeling. *Prog. Oceanogr.* 11: 93–129

Bryan, K., Manabe, S., Pacanowski, R. C. 1975. A global ocean-atmosphere climate model. Part II: the ocean circulation. *J. Phys. Oceanogr.* 5: 30–46

Bryan, K., Manabe, S., Spelman, M. J. 1988. Interhemispheric asymmetry in the transient response of a coupled ocean-atmosphere model to a CO_2 forcing. *J. Phys. Oceanogr.* 18: 851–67

Cane, M. A. 1979. The response of an equatorial ocean to simple wind stress patterns: I. Model formulation and analytic results. *J. Mar. Res.* 37: 233–52

Cane, M. A. 1986. El Niño. *Annu. Rev. Earth Planet. Sci.* 14: 43–70

Cane, M. A., Moore, D. W. 1981. A note on low-frequency equatorial basin modes. *J. Phys. Oceanogr.* 11: 1578–84

Cane, M. A., Münnich, M., Zebiak, S. E. 1990. A study of self-excited oscillations of the tropical ocean-atmosphere system. Part I: linear analysis. *J. Atmos. Sci.* 47: 1562–77

Cane, M. A., Sarachik, E. S. 1977. Forced baroclinic ocean motions: II. The linear equatorial bounded case. *J. Mar. Res.* 35: 395–432

Cane, M. A., Sarachik, E. S. 1981. The response of a linear baroclinic equatorial ocean to periodic forcing. *J. Mar. Res.* 39: 651–93

Cane, M. A., Sarachik, E. S. 1983. Equatorial oceanography. *Rev. Geophys. Space Phys.* 21: 1137–48

Cane, M. A., Zebiak, S. E. 1985. A theory for El Niño and the Southern Oscillation. *Science* 228: 1084–87

Cane, M. A., Zebiak, S. E., Dolan, S. C. 1986. Experimental forecasts of El Niño. *Nature* 321: 827–32

Cayan, D. R. 1992. Latent and sensible heat flux anomalies over the northern oceans: driving the sea surface temperature. *J. Phys. Oceanogr.* 22: 859–81

Chao, Y., Philander, S. G. H. 1993. On the structure of the Southern Oscillation. *J. Climate* 6: 450–69

Charnock, H., Philander, S. G. H., eds. 1989. The dynamics of the coupled atmosphere and ocean. *Proc. of a Royal Society Discussion Meeting*. London: R. Soc. 315 pp.

Cubasch, U., Hasselmann, K., Höck, H., Maier-Reimer, E., Mikolajewicz, U., et al. 1992. Time-dependent greenhouse warming computations with a coupled ocean-atmosphere model. *Climate Dyn.* 8: 55–69

Delworth, T. L., Manabe, S., Stouffer, R. J. 1993. Interdecadal variations of the ther-

mohaline circulation in a coupled ocean-atmosphere model. *J. Climate* In press
Deser C., Wallace J. M. 1990. Large-scale atmospheric circulation features of warm and cold episodes in the tropical Pacific. *J. Climate* 3: 1254-81
Dickinson, R. E. 1983. Land surface processes and climate-surface albedos and energy balance. *Adv. Geophys.* 25: 305–53
Eckmann, J. P. 1981. Roads to turbulence in dissipative dynamical systems. *Rev. Mod. Phys.* 53: 643–53
Endoh, M., Tokioka, T., Nagai, T. 1991. Tropical Pacific sea surface temperature variations in a coupled atmosphere-ocean general circulation model. *J. Mar. Syst.* 1: 293–98
Folland, C., Owen, J., Ward, M. N., Colman, A. 1991. Prediction of seasonal rainfall in the Sahel region using empirical and dynamical methods. *J. Forecasting* 10: 21–56
Gates, W. L., Han, Y. J., Schlesinger, M. E. 1985. The global climate simulated by a coupled atmosphere-ocean general circulation model: preliminary results. See Nihoul 1985, 40: 131–51
Gates, W. L., Mitchell, J. F. B., Boer, G. J., Cubasch, U., Meleshko, V. P. 1992. Climate modeling, climate prediction and model validation. In *Climate Change* 1992: *The supplementary report to the Intergovernmental Panel on Climate Change Scientific Assessment*, ed. J. T. Houghton, B.A. Callander, S.K. Varney, pp. 97–133. Cambridge: Univ. Cambridge Press
Gent, P. R., Cane, M. A. 1989. A reduced gravity, primitive equation model of the upper equatorial ocean. *J. Comput. Phys.* 81: 444–80
Gent, P. R., Tribbia, J. J. 1993. Simulation and predictability in a coupled TOGA model. *J. Climate* In press
Ghil, M., Kimoto, M., Neelin, J. D. 1991. Nonlinear dynamics and predictability in the atmospheric sciences. *Rev. Geophys.*, Suppl., pp. 46–55, U.S. Natl. Rep. to the Int. Union of Geodesy and Geophys. 1987–1990
Ghil, M., Vautard, R. 1991. Interdecadal oscillations and the warming trend in global temperature time series. *Nature* 350: 324–27
Gill, A. E., 1980. Some simple solutions for heat induced tropical circulation. *Q. J. R. Meteorol. Soc.* 106: 447–62
Gill, A. E., 1985. Elements of coupled ocean-atmosphere models for the tropics. See Nihoul 1985, 40: 303–28
Gill, A. E., Clarke, A. J. 1974. Wind-induced upwelling, coastal current, and sea-level changes. *Deep-Sea Res.* 21: 325–45
Gill, A. E., Rasmusson, E. M. 1983. The 1982–1983 climate anomaly in the equatorial Pacific. *Nature* 306: 229–34
Gordon, C. 1989. Tropical ocean-atmosphere interactions in a coupled model. *Phil. Trans. R. Soc. London Ser. A* 329: 207–23
Gordon, A. L., Zebiak, S. E., Bryan, K. 1992. Climate variability and the Atlantic Ocean. *EOS, Trans. Am. Geophys. Union* 73(15): 161
Goswami, B. N., Shukla, J. 1991. Predictability of a coupled ocean-atmosphere model. *J. Climate* 4: 3–22
Graham, N. E., Michaelsen, J., Barnett, T. P. 1987a. An investigation of the El Niño-Southern Oscillation cycle with statistical models. 1. Predictor field characteristics. *J. Geophys. Res.* 92: 14,251–70
Graham, N. E., Michaelsen, J., Barnett, T. P. 1987b. An investigation of the El Niño-Southern Oscillation cycle with statistical models. 2. Model results. *J. Geophys. Res.* 92: 14,271–89
Graham, N. E., White, W. B. 1988. The El Niño cycle: Pacific ocean-atmosphere system. *Science* 240: 1293–302
Graham, N. E., White, W. B. 1990. The role of the western boundary in the ENSO cycle: experiments with coupled models. *J. Phys. Oceanogr.* 20: 1935–48
Hao, Z., Neelin, J. D., Jin, F.-F. 1993. Nonlinear tropical air-sea interaction in the fast-wave limit. *J. Climate* 6: 1523–44
Hasselmann, K. 1976. Stochastic climate models. Part I: theory. *Tellus* 28: 289–305
Hasselmann, K. 1988. PIPs and POPs: the reduction of complex dynamical systems using Principal Interaction and Principal Oscillation Patterns. *J. Geophys. Res.* 93D: 11,015–21
Hayes, S. P., Mangum, L. J., Picaut, L. J., Sumi, A., Takeuchi, K. 1991. TOGA-TAO: a moored array for real-time measurements in the tropical Pacific ocean. *Bull. Am. Meteorol. Sci.* 72: 339
Hense, A., Glowienka-Hense, R., von Storch, H., Stähler, U. 1990. Northern Hemisphere atmospheric response to changes of Atlantic Ocean SST on decadal time scales: a GCM experiment. *Climate Dyn.* 4: 157–74
Hirst, A. C. 1986. Unstable and damped equatorial modes in simple coupled ocean-atmosphere models. *J. Atmos. Sci.* 43: 606–30
Hirst, A. C. 1988. Slow instabilities in tropical ocean basin-global atmosphere models. *J. Atmos. Sci.* 45: 830–52
Houghton, J. T., Jenkins, G. J., Ephraums, J. J., eds. 1990. *Climate Change: The Intergovernmental Panel on Climate Change Scientific Assessment*. Cambridge: Univ. Cambridge Press. 364 pp.

Houghton, R. W., Tourre, Y. 1992. Characteristics of low frequency sea surface temperature fluctuations in the tropical Atlantic. *J. Climate* 5: 765–71

Inoue, M., O'Brien, J. J. 1984. A forecasting model for the onset of a major El Niño. *Mon. Weather Rev.* 112: 2326–37

Jensen, M. H., Bak, P., Bohr, T. 1984. Transition to chaos by interaction of resonances in dissipative systems. Part I. Circle maps. *Phys. Rev. A* 30: 1960–69

Jin, F.-F., Neelin, J. D. 1993a. Modes of interannual tropical ocean-atmosphere interaction—a unified view. Part I: numerical results. *J. Atmos. Sci.* 50: 3477–503

Jin, F.-F., Neelin, J. D. 1993b. Modes of interannual tropical ocean-atmosphere interaction—a unified view. Part III: analytical results in fully-coupled cases. *J. Atmos. Sci.* 50: 3523–40

Keppenne, C. L., Ghil, M. 1992. Adaptive filtering and prediction of the Southern Oscillation index. *J. Geophys. Res.* 97: 20,449–54

Kleeman, R. 1993. On the dependence of hindcast skill on ocean thermodynamics in a coupled ocean-atmosphere model. *J. Climate* In press

Kushnir, Y. 1993. Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. *J. Climate* 6: In press

Kushnir, Y., Lau, N. C. 1992. The general circulation model response to a North Pacific SST anomaly: dependence on time scale and pattern polarity. *J. Climate* 4: 271–83

Lamb, P. J., Peppler, R. A., Hastenrath, S. 1986. Interannual variability in the tropical Atlantic. *Nature* 322: 238–40

Latif, M., Barnett, T. P., Cane, M. A., Flügel, M., Graham, N. E., et al. 1993c. A review of ENSO prediction studies. *Climate Dyn.* In press

Latif, M., Biercamp, J., von Storch, H. 1988. The response of a coupled ocean-atmosphere general circulation model to wind bursts. *J. Atmos. Sci.* 45: 964–79

Latif, M., Graham, N. E. 1992. How much predictive skill is contained in the thermal structure of an OGCM? *J. Phys. Oceanogr.* 22: 951–62

Latif, M., Sterl, A., Maier-Reimer, E., Junge, M. M. 1993a. Climate variability in a coupled GCM. Part I: the tropical Pacific. *J. Climate* 6: 5–21

Latif, M., Sterl, A., Maier-Reimer, E., Junge, M. M. 1993b. Structure and predictability of the El Niño/Southern Oscillation phenomenon in a coupled ocean-atmosphere general circulation model. *J. Climate* 6: 700–8

Latif, M., Villwock, A. 1990. Interannual variability as simulated in coupled ocean-atmosphere models. *J. Mar. Syst.* 1: 51–60

Lau, K. M. 1981a. Oscillations in a simple equatorial climate system. *J. Atmos. Sci.* 38: 248–61

Lau, N. C. 1981b. A diagnostic study of recurrent meteorological anomalies appearing in a 15-year simulation with a GFDL general circulation model. *Mon. Weather Rev.* 109: 2287–311

Lau, N. C. 1985. Modelling the seasonal dependence of the atmospheric response to observed El Niños in 1962–76. *Mon. Weather Rev.* 113: 1970–96

Lau, N. C., Nath, M. J. 1990. A general circulation model study of the atmospheric response to extratropical SST anomalies observed during 1950–79. *J. Climate* 3: 965–89

Lau, N. C., Philander, S. G. H., Nath, M. J. 1992. Simulation of El Niño/Southern Oscillation phenomena with a low-resolution coupled general circulation model of the global ocean and atmosphere. *J. Climate* 5: 284–307

Leetmaa, A., Ji, M. 1989. Operational hindcasting of the tropical Pacific. *Dyn. Atmos. Oceans* 13: 465–90

Lindzen, R. S., Nigam, S. 1987. On the role of sea surface temperature gradients in forcing low level winds and convergence in the tropics. *J. Atmos. Sci.* 44: 2418–36

Lough, J. M. 1986. Tropical Atlantic sea surface temperature and rainfall variations in Subsaharan Africa. *Mon. Weather Rev.* 114: 561–70

Luksch, U., von Storch, H., Maier-Reimer, E. 1990. Modeling North Pacific SST anomalies as a response to anomalous atmospheric forcing. *J. Mar. Systems* 1: 155–68

Manabe, S., Bryan, K. 1969. Climate calculations with a combined ocean-atmosphere model. *J. Atmos. Sci.* 26: 786–89

Manabe, S., Bryan, K., Spelman, M. J. 1975. A global ocean-atmosphere climate model. Part I: the atmospheric circulation. *J. Phys. Oceanogr.* 5: 3–29

Manabe, S., Bryan, K., Spelman, M. J. 1979. A global ocean-atmosphere climate model with seasonal variation for future studies of climate sensitivity. *Dyn. Atmos. Oceans* 3: 393–426

Manabe, S., Bryan, K., Spelman, M. J. 1990. Transient response of a global ocean-atmosphere model to a doubling of atmospheric carbon dioxide. *J. Phys. Oceanogr.* 20: 722–49

Manabe, S., Spelman, M. J., Stouffer, R. J. 1992. Transient responses of a coupled ocean-atmosphere model to gradual changes of atmospheric CO_2. Part II: seasonal response. *J. Climate* 5: 105–26

Manabe, S., Stouffer, R. J. 1988. Two stable equilibria of a coupled ocean-atmosphere model. *J. Climate* 1: 841–66

Matsuno, T. 1966. Quasi-geostrophic motions in the equatorial area. *J. Meteorol. Soc. Jpn.* Ser. II 44: 25–43

McCreary, J. P. 1976. Eastern tropical ocean response to changing wind systems with application to El Niño. *J. Phys. Oceanogr.* 6: 632–45

McCreary, J. P. 1983. A model of tropical ocean-atmosphere interaction. *Mon. Weather Rev.* 111: 370–87

McCreary, J. P. 1985. Modeling equatorial ocean circulation. *Annu. Rev. Fluid Mech.* 17: 359–409

McCreary, J. P., Anderson, D. L. T. 1984. A simple model of El Niño and the Southern Oscillation. *Mon. Weather Rev.* 112: 934–46

McCreary J. P., Anderson D. L. T. 1991. An overview of coupled ocean-atmosphere models of El Niño and the Southern Oscillation. *J. Geophys. Res.* 96: 3125–50

McWilliams, J. C., Gent, P. R. 1978. A coupled air and sea model for the tropical Pacific. *J. Atmos. Sci.* 35: 962–89

Mechoso, C. R., Kitoh, A., Moorthi, S., Arakawa, A. 1987. Numerical simulations of the atmospheric response to a sea surface temperature anomaly over the equatorial eastern Pacific ocean. *Mon. Weather Rev.* 115: 2936–56

Mechoso, C. R., Ma, C.-C., Farrara, J. D., Spahr, J., Moore, R. W. 1993. Parallelization and distribution of a coupled atmosphere-ocean general circulation model. *Mon. Weather Rev.* In press

Meehl, G. A. 1990a. Development of global coupled ocean-atmosphere general circulation models. *Climate Dyn.* 5: 19–33

Meehl, G. A. 1990b. Seasonal cycle forcing of El Niño-Southern Oscillation in a global coupled ocean-atmosphere GCM. *J. Climate* 3: 72–98

Mikolajewicz, U., Maier-Reimer, E. 1990. Internal secular variability in an ocean general circulation model. *Climate Dyn.* 4: 145–56

Mitchell, J. F. B. 1989. The "greenhouse effect" and climate change. *Rev. Geophys.* 27: 115–39

Moore, D. W. 1968. *Planetary-gravity waves in an equatorial ocean.* PhD thesis. Harvard Univ. 207 pp.

Moore, D. W., Philander, S. G. H. 1978. Modeling of the tropical oceanic circulation. In *The Sea*, ed. E. D. Goldberg et al, 6: 319–61. New York: Wiley-Intersci.

Münnich, M., Cane, M. A., Zebiak, S. E. 1991. A study of self-excited oscillations in a tropical ocean-atmosphere system. Part II: nonlinear cases. *J. Atmos. Sci.* 48: 1238–48

Nagai, T., Tokioka, T., Endoh, M., Kitamura, Y. 1992. El Niño–Southern oscillation simulated in an MRI atmosphere-ocean coupled general circulation model. *J. Climate* 5: 1202–33

Namias, J. 1959. Recent seasonal interactions between North Pacific waters and the overlying atmospheric circulation. *J. Geophys. Res.* 64: 631–46

Neelin, J. D. 1989a. A note on the interpretation of the Gill model. *J. Atmos. Sci.* 46: 2466–68

Neelin, J. D. 1989b. Interannual oscillations in an ocean general circulation model coupled to a simple atmosphere model. *Phil. Trans. R. Soc. London Ser. A* 329: 189–205

Neelin, J. D. 1990. A hybrid coupled general circulation model for El Niño studies. *J. Atmos. Sci.* 47: 674–93

Neelin, J. D. 1991. The slow sea surface temperature mode and the fast-wave limit: analytic theory for tropical interannual oscillations and experiments in a hybrid coupled model. *J. Atmos. Sci.* 48: 584–606

Neelin, J. D., Held, I. M. 1987. Modelling tropical convergence based on the moist static energy budget. *Mon. Weather Rev.* 115: 3–12

Neelin, J. D., Jin, F.-F. 1993. Modes of interannual tropical ocean-atmosphere interaction—a unified view. Part II: analytical results in the weak-coupling limit. *J. Atmos. Sci.* 50: 3504–22

Neelin, J. D., Latif, M., Allaart, M. A. F., Cane, M. A., Cubasch, U., et al. 1992. Tropical air-sea interaction in general circulation models. *Climate Dyn.* 7: 73–104

Nihoul, J. C. J., ed. 1985. *Coupled Ocean-Atmosphere Models.* Amsterdam: Elsevier Oceanogr. Ser.

Nihoul, J. C. J., ed. 1990. *Coupled Ocean-Atmosphere Modeling.* Amsterdam: Elsevier. 313 pp.

Palmer, T. N., Mansfield, D. A. 1986. A study of wintertime circulation amomalies during past El Niño events using a high-resolution general circulation model. Part I: influence of model climatology. *Q. J. R. Meteorol. Soc.* 112: 613–38

Palmer, T. N., Sun, Z.-B. 1985. A modeling and observational study of the relationship of sea surface temperature in the northwestern Atlantic and the atmospheric general circulation. *Q. J. R. Meteorol. Soc.* 111: 947–75

Penland, C., Magorian, T. 1993. Prediction of Niño-3 sea surface temperatures using linear inverse modeling. *J. Climate* 6: 1067–76

Philander, S. G. H. 1981. The response of the equatorial oceans to a relaxation of the trade winds. *J. Phys. Oceanogr.* 11: 176–89

Philander, S. G. H. 1990. *El Niño, La Niña, and the Southern Oscillation.* San Diego: Academic. 293 pp.

Philander, S. G. H., Chao, Y. 1991. On the contrast between the seasonal cycles of the equatorial Atlantic and Pacific oceans. *J. Phys. Oceanogr.* 21: 1399–406

Philander, S. G. H., Lau, N. C., Pacanowski, R. C., Nath, M. J. 1989. Two different simulations of Southern Oscillation and El Niño with coupled ocean-atmosphere general circulation models. *Phil. Trans. R. Soc. London Ser. A* 329: 167–78

Philander, S. G. H., Pacanowski, R. C. 1980. The generation of equatorial currents. *J. Geophys. Res.* 85: 1123–36

Philander, S. G. H., Pacanowski, R. C. 1981. Response of the equatorial ocean to periodic forcing. *J. Geophys. Res.* 86: 1903–16

Philander, S. G. H., Pacanowski, R. C., Lau, N. C., Nath, M. J. 1992. Simulation of ENSO with a global atmospheric GCM coupled to a high-resolution, tropical Pacific ocean GCM. *J. Climate* 5: 308–29

Philander, S. G. H., Yamagata, T., Pacanowski, R. C. 1984. Unstable air-sea interactions in the tropics. *J. Atmos. Sci.* 41: 604–13

Pitcher, E. J., Blackmon, M. L., Bates, G. T., Muñoz, S. 1988. The effect of North Pacific sea surface temperature anomalies on the January climate of a general circulation model. *J. Atmos. Sci.* 45: 173–88

Rasmusson, E. M. 1984. El Niño: the ocean/atmosphere connection. *Oceanus* 27: 5–13

Rasmusson, E. M., Carpenter, T. H. 1982. Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation/El Niño. *Mon. Weather Rev.* 110: 354–84

Rasmusson, E. M., Wallace, J. M 1983. Meteorological aspects of El Niño/Southern Oscillation. *Science* 222: 1195–202

Rasmusson, E. M., Wang, X., Ropelewski, C. F. 1990. The biennial component of ENSO variability. *J. Mar. Syst.* 1: 71–96

Ropelewski, C. F., Halpert, M. S., Wang, X. 1992. Observed tropospheric biennial variability and its relation to the Southern Oscillation. *J. Climate* 5: 594–614

Sausen, R., Barthels, K., Hasselmann, K. 1988. Coupled ocean-atmosphere models with flux correction. *Climate Dyn.* 2: 154–63

Schlesinger, M. E., ed. 1990. *Climate-Ocean Interaction.* Dordrecht: Kluwer. 385 pp.

Schlesinger, M. E., Gates, W. L., Han, Y. J. 1985. The role of the ocean in CO_2-induced climate change: preliminary results from the OSU coupled atmosphere-ocean general circulation model. See Nihoul 1985, 40: 447–78

Schopf, P. S., Cane, M. A. 1983. On equatorial dynamics, mixed layer physics and sea surface temperature. *J. Phys. Oceanogr.* 13: 917–35

Schopf, P. S., Suarez M. J. 1988. Vacillations in a coupled ocean-atmosphere model. *J. Atmos. Sci.* 45: 549–66

Schopf, P. S., Suarez, M. J. 1990. Ocean wave dynamics and the time scale of ENSO. *J. Phys. Oceanogr.* 20: 629–45

Seager, R., Zebiak, S. E., Cane, M. A. 1988. A model of the tropical Pacific sea surface temperature climatology. *J. Geophys. Res.* 93: 1265–80

Servain, J. 1991. Simple climatic indices for the tropical Atlantic Ocean and some applications. *J. Geophys. Res.* 96: 15137–46

Servain, J., Legler, D. M. 1986. Empirical orthogonal function analyses of tropical Atlantic sea surface temperature and wind stress: 1964–1979. *J. Geophys. Res.* 91: 14181–91

Shukla, J., Fennessey, M. J. 1988. Prediction of time-mean atmospheric circulation and rainfall: influence of Pacific sea surface temperature anomaly. *J. Atmos. Sci.* 45: 9–28

Shukla, J., Wallace, J. M. 1983. Numerical simulation of the atmospheric response to equatorial Pacific sea surface temperature anomalies. *J. Atmos. Sci.* 40: 1613–30

Sperber, K. R., Hameed, S. 1991. Southern Oscillation in the OSU coupled upper ocean-atmosphere GCM. *Climate Dyn.* 6: 83–97

Sperber, K. R., Hameed, S., Gates, W. L., Potter, G. L. 1987. Southern Oscillation simulated in a global climate model. *Nature* 329: 140–42

Stouffer, R. J., Manabe, S., Bryan, K. 1989. Interhemispheric asymmetry in climate response to a gradual increase of atmospheric CO_2. *Nature* 342: 660–62

Suarez, M. J., Schopf, P. S. 1988. A delayed action oscillator for ENSO. *J. Atmos. Sci.* 45: 3283–87

Tai, C. K., White, W. B., Pazan S. E. 1989. Geosat crossover analysis in the tropical Pacific. 2. Verification analysis of altimetric sea level maps with expendable bathythermograph and island sea level data. *J. Geophys. Res.* 94C: 897–908

Trenberth, K. E., ed. 1993. *Climate System Modeling.* New York: Cambridge Univ. Press. 817 pp.

Vallis, G. K., 1986: El Niño: a chaotic dynamical system? *Science* 232: 243–45
Vallis, G. K. 1988. Conceptual models of El Niño and the Southern Oscillation. *J. Geophys. Res.* 93C: 13,979–91
Wakata, Y., Sarachik, E. S. 1991. Unstable coupled atmosphere-ocean basin modes in the presence of a spatially varying basic state. *J. Atmos. Sci.* 48: 2060–77
Walker, G. T. 1923. Correlation in seasonal variations of weather, VIII: a preliminary study of world weather. *Mem. Indian Meteorol. Dep.* 24: 75–131
Wallace, J. M., Gutzler, D. S. 1981. Teleconnections in the potential height field during the Northern Hemisphere winter. *Mon. Weather Rev.* 109: 784–812
Wallace, J. M., Jiang, Q. 1987. On the observed structure of the interannual variability of the atmosphere/ocean climate system. In *Atmospheric and Oceanic Variability*, ed. H. Cattle, pp. 17–43. London: Roy. Meteorol. Soc.
Wallace, J. M., Smith, C., Jiang, Q. 1990. Spatial patterns of atmosphere-ocean interaction in the northern winter. *J. Climate* 3: 990–98
Washington, W. M., Meehl, G. A. 1989. Climate sensitivity due to increased CO_2: experiments with a coupled atmosphere and ocean general circulation model. *Climate Dyn.* 4: 1–38
Washington, W. M., Parkinson, C. L. 1986. *An Introduction to Three-Dimensional Climate Modeling*. Mill Valley, Calif: Univ. Sci. Books. 422 pp.
Washington, W. M., Semtner, A. J., Meehl, G. A., Knight, D. J., Mayer, T. A. 1980. A general circulation experiment with a coupled atmosphere, ocean and sea ice model. *J. Phys. Oceanogr.* 10: 1887–908
Weare, B., Navato, A., Newell, R. F. 1976. Empirical Orthogonal analysis of Pacific Ocean sea surface temperatures. *J. Phys. Oceanogr.* 6: 671–78
Weaver, A. J., Sarachik, E. S., Marotzke, J. 1991. Freshwater flux forcing of decadal and interdecadal oceanic variability. *Nature* 353: 836–38
Webster, P. J., Yang, S. 1992. Monsoon and ENSO: selectively interactive systems. *Q. J. R. Meteorol. Soc.* 118: 877–926
Wolter, K. 1989. Modes of tropical circulation, Southern Oscillation, and Sahel rainfall anomalies. *J. Climate* 8: 149–72
Xie, S.-P., Kubokawa, A., Hanawa, K. 1989. Oscillations with two feedback processes in a coupled ocean-atmosphere model. *J. Climate* 2: 946–64
Xu, J.-S., von Storch, H. 1990. Predicting the state of the Southern Oscillation using principal-oscillation-pattern analysis. *J. Climate* 3: 1316–29
Yamagata, T. 1985. Stability of a simple air-sea coupled model in the tropics. See Nihoul 1985, pp. 637–57
Yamagata, T., Masumoto, Y. 1989. A simple ocean-atmosphere coupled model for the origin of warm El Niño Southern Oscillation event. *Phil. Trans. R. Soc. London Ser. A* 329: 225–36
Yasunari, T., Seki, Y. 1992. Role of Asian monsoon on the interannual variability of the global climate system. *J. Meteorol. Soc. Jpn.* 70: 177–89
Zebiak, S. E. 1986. Atmospheric convergence feedback in a simple model for El Niño. *Mon. Weather Rev.* 114: 1263–71
Zebiak, S. E. 1989. On the 30–60 day oscillation and prediction of El Niño. *J. Climate* 2: 1381–87
Zebiak, S. E. 1993. Air-sea interaction in the equatorial Atlantic region. *J. Climate* 6: 1567–86
Zebiak, S. E., Cane, M. A. 1987. A model El Niño Southern Oscillation. *Mon. Weather Rev.* 115: 2262–78