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Instability and predictability in coupled atmosphere–ocean models

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We review simple instabilities in linear theories of coupled atmosphere–ocean models in both bounded and unbounded ocean basins and describe the mechanisms for instability in these linear theories. We then review nonlinear coupled atmosphere–ocean simulations of the El Niño Southern Oscillation (ENSO) phenomenon and relate the instabilities seen in linear theory to the fully nonlinear ENSO simulations. We present a general discussion of the relation between instability and predictability in the ENSO problem and review some recent work on predictability in coupled models. Finally, we comment on some recent predictions in light of our discussion of predictability.

1. Introduction

When an atmosphere, which itself has no instabilities, interacts with an ocean, which itself has no instabilities, instabilities of long time and space scales can arise. The surface of the eastern equatorial Pacific, for example, is normally colder than the western Pacific, and is under the influence of atmospheric mean divergence driven by the (convergent) region of persistent precipitation over the 'maritime continent' in the far west Pacific. If a warm sea surface temperature (SST) anomaly alters the atmospheric circulation in such a way that the induced anomalous surface winds enhance the anomalous SST, then the anomalies will grow in both the atmosphere and ocean.

In general, this will happen by inducing westerly surface wind anomalies to the west of the warm SST anomaly, which could then warm the surface of the ocean by a number of different processes. The westerly surface wind anomaly could force anomalous eastward ocean currents which, because of a pre-existing mean eastward SST gradient, would then warm by anomalous zonal advection. The westerly surface wind anomaly could weaken the normally poleward currents on both sides of the Equator which, because of a pre-existing mean equatorward SST gradient, would then warm by anomalous meridional advection. The westerly surface wind anomaly could reduce the magnitude of the upwelling which would then warm by anomalous vertical advection. The westerly surface wind anomaly could deepen the thermocline, thereby weakening the effective vertical temperature gradient, and the surface would then warm because of mean vertical advection acting on an anomalously weak vertical temperature gradient. These, and other permutations of terms in the ocean thermodynamic equation, could act singly, or in concert, to anomalously warm the surface of the ocean and thereby cause a growing coupled atmosphere–ocean instability. As it turns out, the nature of the induced instability depends sensitively on the mix of thermodynamic processes in the ocean that contribute to the anomalous SST changes.

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It is the purpose of this paper to review the types of large-scale instability that arise from coupled atmosphere–ocean interactions and to point out their relevance to the El Niño Southern Oscillation (ENSO) phenomenon. In the second section we will review the work done on instabilities in simple linear models of the atmosphere interacting with similarly simple linear models of the ocean in both bounded and unbounded domains. In the third section we will discuss the ENSO cycles as simulated in more complicated models and, especially, the basic linearized instability in the Cane–Zebiak model and its relation to the model generated ENSO phenomenon. In the fourth section we will emphasize how ubiquitous coupled instabilities seem to be and discuss the implications of this for the predictability of the ENSO phenomenon. We conclude in the fifth section with suggestions for future research in the applicability of these ideas to more complex models.

2. Instabilities in simple coupled models

We here review simple coupled instabilities in unbounded basins with the notation and theory of Hirst (1986). Many other authors contributed to this theory, beginning with the work of Philander *et al.* (1984), and we refer the reader to the references in Hirst (1986) and in the papers and references in Nihoul (1985).

The atmosphere is taken to be a simple, thermally driven baroclinic fluid on an equatorial β-plane similar to the model originally considered by Gill (1980). The equations of the atmosphere are

$$\left.\begin{aligned} U_t-\beta yV+\phi_x+AU&=0,\\ V_t+\beta yU+\phi_y+AV&=0,\\ \text{and}\quad \phi_t+c_a^2(U_x+V_y)+B\phi&=Q,\end{aligned}\right\}\tag{1}$$

where (U,V) are the baroclinic components of the wind, ϕ is the geopotential thickness, c_a is the baroclinic gravity wave speed, Q is the thermal forcing of the atmosphere, and A and B are Rayleigh damping coefficients.

The ocean is taken to be a shallow-water model of constant (unit) density on a motionless basic state of unperturbed upper layer depth H whose dynamics is given by

$$\left.\begin{aligned} u_t-\beta yv+g'h_x+au&=\tau^{(x)}/H,\\ v_t-\beta yu+g'h_y+av&=\tau^{(y)}/H,\\ \text{and}\quad h_t+H(u_x+v_y)+bh&=0,\end{aligned}\right\}\tag{2}$$

where (u, v) are the currents in the layer, h is the perturbation layer depth, g' is the value of reduced gravity (so that $(g'H)^{\frac{1}{2}}$ is the Kelvin wave speed), a and b are Rayleigh damping coefficients, and $(\tau^{(x)}, \tau^{(y)})$ are the components of the surface wind stress.

The ocean thermodynamics is given by an equation for the sea surface temperature T, which may be thought of as the temperature of the active surface layer:

$$T_t+\bar{T}_x\,u-K_T\,h+\alpha T=0.\tag{3}$$

Equation (3) indicates that SST can be changed by anomalous advection on the mean zonal SST gradient $\bar{T}_x$, by changes in the thermocline depth anomaly h, or by damping. The second term may be thought of as a weakening of the mean cooling by entrainment of cold water into the upper layer as a result of upwelling (i.e. as an anomalous heating) as the thermocline deepens.

The coupling must relate the thermal forcing Q on the right-hand side of the atmospheric equations (1) to the oceanic quantities, and the wind stress forcing $\boldsymbol{\tau}$ on the right-hand side of the oceanic equations (2) to the atmospheric quantities. The simplest coupling scheme is

$$Q = K_Q\, T \quad \text{and} \quad (\tau^{(x)}/H, \tau^{(y)}/H) = -K_S(U, V), \tag{4}$$

so that the anomalous thermal forcing occurs over anomalously warm or cold water, and the wind stresses are linearly related to the surface winds (which themselves are related linearly and negatively to the baroclinic wind components U and V).

The system given by equations (1)–(4) is solved by assuming normal mode solutions $\exp[\mathrm{i}(kx-\sigma t)]$ and solving for the complex eigenvalues σ subject to the boundary condition that all quantities die away as $y \rightarrow \infty$. The results turn out to depend sensitively on the ocean thermodynamic equation (3) and it proves convenient, for ease of interpretation, to divide (3) into separate categories, each one of which defines a model (Hirst 1986), as follows.

(I) $T = Kh$ (SST anomalies proportional to h anomalies);

(II) $T_t + \bar{T}_x\, u + \alpha T = 0$ (no entrainment);

(III) $T_t + \bar{T}_x\, u - K_T\, h + \alpha T = 0$ (full model);

(IV) $T_t - K_T\, h + \alpha T = 0$ (entrainment dominated).

The energetics of the coupled model indicates that instabilities are possible, for all the thermodynamic models, if and only if $\langle \phi Q \rangle > 0$ and $\langle \boldsymbol{\tau} \cdot \boldsymbol{u} \rangle > 0$, i.e. if the heating in the atmosphere takes place where the geopotential thickness is already large, and if the wind stress over the ocean does work on the ocean.

For all of the thermodynamic models I–IV, the following general statements can be made.

(i) For large enough values of the atmosphere–ocean coupling coefficients K_Q and K_S, taken together as the product $K_Q\, K_S$, instabilities of the coupled system arise.

(ii) For representative (i.e. typical) values of the coupling coefficients $K_Q\, K_S$, all the models have coupled instabilities.

(iii) All the coupled instabilities are long-wavelength instabilities, are equatorially confined, and have growth rates of order months.

(iv) Assuming that the atmosphere is everywhere in local equilibrium with the ocean (i.e. assuming that $U_t = V_t = \phi_t = 0$ in equation (1)) makes no difference for the slow, long-wavelength unstable modes.

(v) Assuming the long-wavelength approximation for the ocean (i.e. that $v_t = 0$ in equation (2)) makes no difference for the slow, long-wavelength unstable modes.

For models I and II, the form of the thermal equation allows the SST to be expressed entirely in terms of dynamical quantities and the resulting instabilities have characteristics of modified ocean waves travelling with almost free-ocean wave speeds. For model I, the SST is in phase with the thermocline depth and for model II, the SST lags the thermocline depth by a quarter of a cycle. Figure 1 then indicates that the instability of model I is basically an eastward-propagating Kelvin wave in the ocean with the atmospheric thermal forcing dragged along with it. Only for the eastward-propagating Kelvin wave is the surface wind stress in phase with the ocean surface currents so that the atmosphere can do work on the ocean and the perturbation can grow. Similarly, for model II, only the westward-propagating ocean Rossby

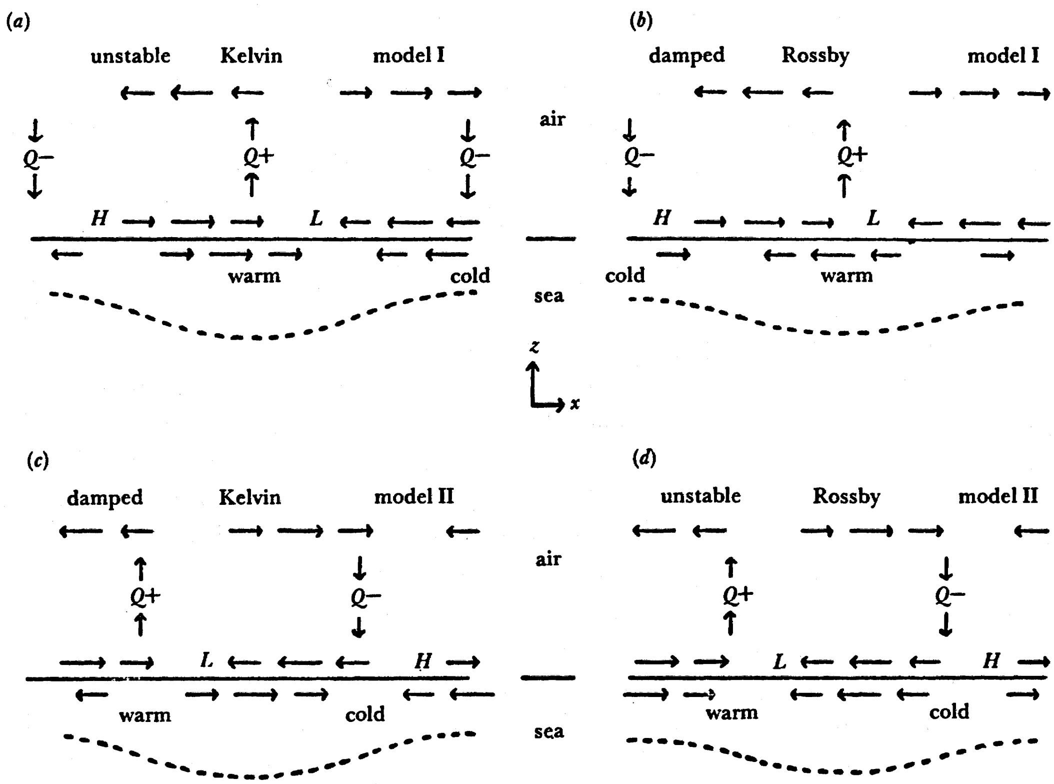

FIGURE 1. Schematic illustration of unstable ((*a*) and (*d*)) and damped ((*b*) and (*c*)) modes for the models I and II. Arrows indicate ocean currents and surface wind perturbations. Maximum positive and negative perturbations of SST are indicated by warm and cold, of thermal forcing by $Q+$ and $Q-$, and of surface pressure by H and L. (After Hirst (1986).)

wave has the surface winds in phase with the ocean surface currents. This instability resembles a free-ocean Rossby wave with the atmospheric thermal forcing dragged along with it.

The frequencies Re σ and the growth rates Im σ for the instabilities induced in the more realistic models III and IV are shown in figure 2. Both models have long-wave instabilities (wavelengths greater than 10000 km) that oscillate slowly (periods of the order of years), move slowly eastward or westward (speeds of the order of centimetres per second), and grow slowly (growth times of the order of months). The coupled instabilities no longer resemble free modes in the ocean and are modes that arise uniquely from the coupling of the atmosphere and the ocean. Clearly the coupling of the sea surface temperature to the atmospheric winds is the crucial one: the unstable modes can oscillate and/or propagate only as fast as the SST can change.

When boundaries at the east and west are added to the ocean, for the oceanographically relevant models III and IV, it turns out (Hirst 1988) that the instabilities in the bounded basin of length L are very similar to the instabilities of wavelength L in the unbounded case. In particular, instabilities will arise in basins the size of the Pacific but will not arise in basins the size of the Atlantic or Indian Ocean. These instabilities may be considered as modifications of the unbounded instabilities and are characterized by slow eastward propagation. Because they appear so similar to the unbounded instabilities, we may conclude that the boundaries are not essential for their existence, but that propagation is. As we will see in the next section, the

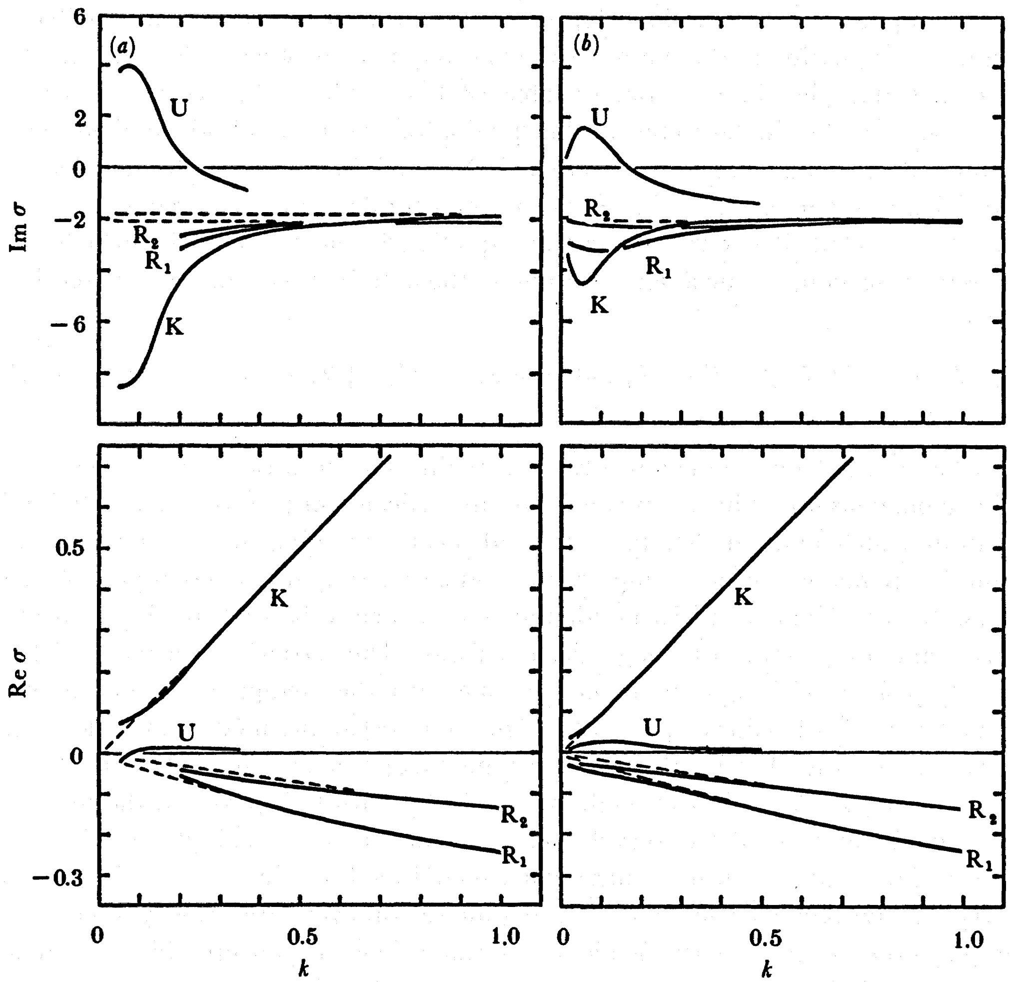

FIGURE 2. Growth rates (Im σ) and oscillation frequencies (Re σ) against wavelength (k) of eigenmodes for thermodynamics characteristic of (*a*) model III and (*b*) model IV. U represents the unstable mode in both cases, K and R are the damped Kelvin and Rossby modes. The non-dimensionalization is such that that Re $\sigma = 0.1$ corresponds to a period of 131 days, Im $\sigma = 1$ corresponds to a growth rate of (208 days)$^{-1}$, and $k = 0.1$ corresponds to a wavelength of 16000 km. (After Hirst (1986).)

coupled instability characterizing the dominant ENSO cycle in the Cane–Zebiak model (Cane & Zebiak 1985; Zebiak & Cane 1987) is not propagating and the boundaries play an essential role.

3. Instabilities in more complex models

The model originally constructed by Cane & Zebiak (1985) has been shown to contain a reasonable ENSO cycle (Zebiak & Cane 1987) and to have skill in predictions out to as long as two years (Cane *et al.* 1986; Cane & Zebiak 1987). The coupled model calculates anomalies about a specified annual cycle in both the atmosphere and ocean in a rectangular basin the size of the Pacific: in the atmosphere, the seasonal cycle of surface wind and convergence is given by observations; in the ocean, only the annual cycle of SST is presently observed and the annual cycle of currents, vertical velocity, and thermocline depth is generated by driving the ocean model with the observed winds. The atmospheric model consists of a simple thermally driven

(Gill 1980) atmosphere, as in equation (1), except that a form of convergence feedback is utilized, i.e. the precipitation in the thermal sources is required to be consistent with the low-level convergence that the thermal sources force (Zebiak 1986). The ocean consists of a shallow-water ocean and is similar to the one in equation (2) except that horizontal advections due to the annually varying mean currents are included, and, in addition, a frictional surface layer of fixed depth h_s whose convergence and divergences determine the magnitude of the vertical velocities is included. The thermodynamic equation for the SST anomalies, which, as we saw in the previous section, is crucial for determining the nature of the coupled instabilities, is in this model

$$T_t + u(\bar{T}+T)_x + v(\bar{T}+T)_y + \bar{u}T_x + \bar{v}T_y + \delta[M(\bar{w}+w) - M(\bar{w})]\ \bar{T}_z + \delta M(\bar{w}+w)\ T_z = -\alpha T. \tag{5}$$

The SST can be changed by anomalous advection acting on the mean and anomalous SST gradients, by anomalous upwelling acting on the mean vertical temperature gradient (with the standard entrainment condition that the total and mean upwelling are positive), by total upwelling on the anomalous vertical temperature gradient, and by thermal damping of SST by surface fluxes. $M(x)$ is the modified Heaviside function defined to be x when x is positive and zero otherwise, and δ represents a mixing efficiency factor. The vertical gradients are defined as $T_z = (T - T_{sub})/h_s$ where T_{sub} is a function giving the subsurface temperature in terms of the mean and anomalous thermocline depths and h_s. The model ocean thermodynamics clearly has a richer variety of ways to change SST than the simple models defined by equation (3).

A close version of the Cane–Zebiak model was built by Battisti (1988) and the dynamics and thermodynamics of the model ENSO cycle were examined in great detail. By examining the full nonlinearly simulated ENSO using a meridional modal analysis, Battisti was able to show that the initial downwelling warming signal during the El Niño year grew slowly in place (i.e. without propagation) in the eastern Pacific at the same time that an upwelling signal was excited in the central Pacific, propagated freely off to the western boundary in the gravest meridional Rossby mode, returned to the warming area in an equatorially trapped Kelvin mode, and destroyed the downwelling signal to end the ENSO warm phase and begin the ENSO cold phase. The sequence of events suggested that a warm instability grew in the eastern Pacific, and got turned around by cooling effects connected somehow with the initial growth.

This possibility was examined in detail by Battisti & Hirst (1989) by linearizing the coupled model and examining the growth of the resulting coupled instabilities in detail. Because the subsurface T_{sub} could be linearized to the form $a(\bar{h})\ h$, the linearized thermodynamic equation for anomalous SST becomes

$$T_t + u\bar{T}_x + v\bar{T}_y + \bar{u}T_x + \bar{v}T_y + w\bar{T}_z + M(\bar{w})\ (T - a(\bar{h})h)/h_s = -\alpha T, \tag{6}$$

and we see that the thermodynamic equation is related to, but still more complicated than, the simple thermodynamic equation (3). A stability analysis assuming that the mean quantities $\bar{u}$, $\bar{v}$, $\bar{w}$, and $\bar{h}$ are constants (Wakata 1989), gives unstable coupled modes very similar to those of Hirst (1986).

In the full linearized coupled model, the mean quantities $\bar{u}$, $\bar{v}$, $\bar{w}$, and $\bar{h}$, however, all vary in time with the annual cycle and are, in addition, functions of latitude and longitude so that the stability analysis becomes considerably more complicated. Battisti & Hirst (1989) specified these mean quantities as in Zebiak & Cane (1987) and performed the stability analysis of the linearized system in the bounded basin. In the simplest case, the mean state from which anomalies are calculated was taken to be the annually averaged mean state so that the annual cycle was temporarily suppressed. The results indicated the following.

(i) There is an exponentially growing unstable mode that has a frequency quite close to that of the ENSO cycle in the full coupled nonlinear model.

(ii) The unstable mode has most of its amplitude in SST in the eastern Pacific and in surface zonal wind in the east-central Pacific. The anomalies grow and oscillate in place and have negligible propagation characteristics. The spatial character of the growing and oscillating modes, in SST, thermocline depth, and surface winds, are very close to those of the ENSO cycle in the full coupled nonlinear model.

(iii) The latitudinal and (especially) longitudinal inhomogeneities in the mean state are fundamental in localizing the coupled instabilities to the eastern and central Pacific: the coupled instabilities so generated have a different physics (see point (iv) below) than those generated with no latitudinal or longitudinal inhomogeneities which resemble the propagating instabilities of Hirst (1986).

(iv) The sea surface temperature of the growing and oscillating instability could, with a high degree of accuracy, be described by the retarded oscillator equation

$$\mathrm{d}T(t)/\mathrm{d}t = cT(t) - bT(t-\tau), \tag{7a}$$

where c and b are derived from the linearized coupled model to be $c = 2.2$ per year and $b = 3.9$ per year, and τ is the retarded time having the value 6 months in this model (to be explained more fully below).

(v) The retarded oscillator equation (7 a) has the property that growing oscillations of period small compared to 2τ can exist only when $b > c$.

(vi) As the parameters of the full nonlinear model are changed, equation (7 a) gives a very good representation of the changes of growth rate and oscillation period of both the linearized coupled model and the full nonlinear coupled model.

The interpretation of equation (7 a) can be given with the aid of figure 3 as follows. Consider the coupled instability in the eastern part of the Pacific such that, in the absence of boundaries, the SST would grow purely exponentially in time with a growth rate c and would be confined latitudinally to the region of the equator. The exponentially growing surface winds induced by the growing SST lie to the west of the SST, as is usual in models of Gill type (and in the observations). Because the winds cover a finite longitudinal extent, they tend to lower the

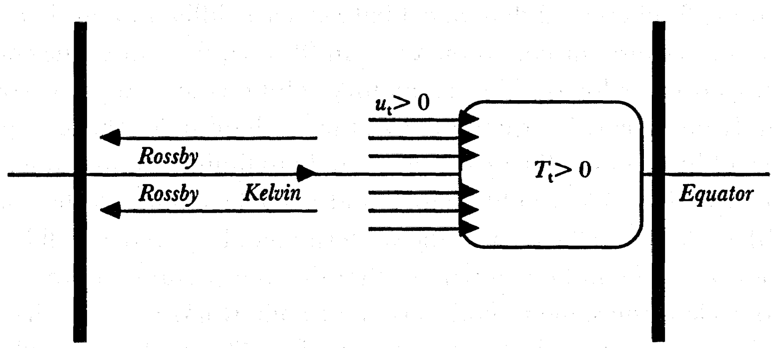

FIGURE 3. Schematic diagram of the retarded oscillator mechanism. See text for explanation.

thermocline to their east and raise the thermocline to their west. The signal to the east of the winds is consistent with the warming in the eastern Pacific whereas the upwelling (cooling) signal propagates freely westward behind signals propagating with speeds and meridional structures characteristic of grave Rossby modes, hits the western boundary, and returns behind wavefronts with speeds and meridional structure characteristic of Kelvin modes.

After a time τ (the length of time it takes for the signal to return to the region of growing SST) the cooling signal, growing exponentially with an amplitude characteristic of its *retarded* time $t-\tau$, reaches the east Pacific and begins to erode the warming signal, eventually turning it around to cooling. Clearly, this is only possible when the remote signal has an amplitude that exceeds the direct signal; this gives an interpretation of the condition $b > c$. We see that c involves all the terms of the thermal equation that locally change the temperature, including the damping by fluxes, whereas b involves only that term of the linearized thermal equation (6) that arises from thermocline depth, i.e. mean upwelling on anomalous vertical temperature gradient. Because the cooling signal travels a time τ before it shows up as SST changes, it is effectively shielded from damping by surface fluxes (which is of the form $-\alpha T$) and reaches the eastern Pacific undiminished. This is the reason why the remote signal, given by b, has a greater ability to change SST than the direct effects, given by c (the precise arguments are given in detail in Battisti & Hirst (1989)). Because the solutions of equation (7 *a*) grow exponentially, nonlinearities must act to limit the amplitude and produce an equilibrium ENSO cycle. The leading-order nonlinear analogue model can be derived to be

$$\mathrm{d}T(t)/\mathrm{d}t = cT(t) - bT(t-\tau) - e[T(t) - rT(t-\tau)]^3, \qquad (7\,b)$$

where r and e are constants defined by the basic state. The addition of such a nonlinear term has only a small effect on the period of the slow oscillation.

Schopf & Suarez (1988) found similar non-propagating disturbances of long period confined to the eastern Pacific that they also identified with ENSO cycles by utilizing a coupled model in which a Gill atmosphere was coupled to an ocean similar to the one in the Cane–Zebiak model but having the additional complexity that the surface layer was treated as an interacting mixed layer with variable depth. On the basis of an analysis of this model, they proposed an equation similar to equation (7 *b*), (Suarez & Schopf 1988)

$$\mathrm{d}T(t)/\mathrm{d}t = T(t) - \alpha T(t-\tau) - T^3(t), \qquad (8)$$

where they argued that $a < 1$ because the retarded effects would return diminished by reflections at the western boundary. Although equations (7 *b*) and (8) look similar, it should be pointed out that the fundamental dynamical balances are different and therefore the solutions are vastly different. Because the condition $\alpha < 1$ in (8) is equivalent to the condition $c > b$, the equation (8) has no *linear* long-period oscillatory solutions and any long-period periodicity must be inherently nonlinear. In equation (7 *b*), the mechanism for the long-period oscillations is linear and the addition of nonlinearities enters only to limit the amplitude. The character of the solutions to equation (7 *b*) would change and become similar to that of the solutions to equation (8) if the reflection efficiency at the western boundary were to fall below 55%, so that b would fall below c. It should be pointed out that the interpretation of the retardation and its role in the ENSO cycle is the same in both sets of authors' works and was first given by Schopf & Suarez (1988), made explicit in Suarez & Schopf (1988), and expanded upon in detail in Schopf *et al.* (1989).

We have noted two rather different types of instability in simple coupled models: those that propagate and those that do not. The mechanisms are rather distinct and have been investigated in detail in a paper by Hirst (1989). The propagating instabilities are a result of local interactions and do not depend on the existence of the boundaries even when they exist. The oscillatory property of the non-propagating instabilities depend in an essential way on the existence of boundaries, as the retarded oscillator mechanism makes explicit.

At the moment, we know of no comparable analyses of the instabilities in much more complex models, namely atmospheric general circulation models (GCMS) coupled to oceanic GCMS. One of the most interesting examples of fully coupled GCMS is the model of Latif *et al.* (1989), which exhibits no significant interannual variability characteristic of the ENSO cycle in the coupled mode, this despite the ability of each of the atmosphere and the ocean to reproduce its respective effects of ENSO when forced by *observed* ENSO forcing from the other medium. Although it is tempting to speculate, on the basis of the stability analyses of simpler systems, that either the effective coupling is too weak or the effective τ is too small (say by having the thermocline too deep so that the equatorial timescale of the lowest baroclinic mode is too large), it is clear that we can learn much about instability from systems that fail to exhibit it.

4. Instability and predictability

It appears, from the work reviewed in the previous sections, that the ENSO phenomenon in simple models is primarily a result of an instability of the coupled atmosphere–ocean system and that this instability is characteristic of a wide variety of physical situations. Although it is too early to extend this conclusion either to ENSO cycles in more complicated coupled GCM models or to ENSO cycles in Nature, let us assume that we can make this leap and see what we can conclude about the predictability of ENSO.

We know that, in classical weather prediction, the aperiodicity, and therefore the non-predictability, of mid-latitude motions can be identified with instability and nonlinearity. Errors in initial conditions grow rapidly, with timescales of days, and the ultimate limit of predictability is therefore believed to be of the order of two weeks.

The ENSO cycle also is aperiodic and therefore is not perfectly predictable. If the basic mechanism for the irregularity in the cycle were nonlinear (as in the 'chaos' theory of Vallis (1986)) then small errors in initial conditions will grow (albeit with timescales of months) and the situation would be akin to the classical one, with the ultimate limit of predictability being on the order of a year.

But if the basic mechanism for the oscillation is linear (as in equation (7*a*)), then linear instabilities lead to a perfectly regular cycle (even when equilibrated by nonlinearities) and instability leads to perfect predictability. A source of the irregularity of the ENSO cycle in such a case could be stochastic perturbations, presumably because of higher-frequency weather and other fluctuations not perfectly coupled to the slow SST variations characteristic of ENSO. Unstable configurations of the coupled atmosphere and ocean would grow into a perfectly periodic, and therefore perfectly predictable, ENSO cycle unless random noise somehow stabilized the system and prevented it from growing. If the system were stable, the ENSO oscillations would not grow and it too would be perfectly predictable.

We see that, unlike the weather prediction paradigm, instability is sometimes associated with perfect predictability; the lack of perfect predictability is caused by two factors: the inherent unpredictability of external noise and the possibility that this noise causes transitions between stable and unstable states.

Even in the relatively simple models the situation is not quite so simple, a basic complication being the annual cycle. The simple retarded oscillator equation, as derived by Battisti & Hirst (1989) is valid only for anomalies on an annually averaged mean state, with the annual cycle suppressed. The annually averaged mean state is indeed unstable, and the unstable modes on this mean state looks very much like an ENSO cycle of period 3.5 years. The system is highly regular because it is very far from any stable states, although it is conceivable that an external finite amplitude perturbation could deform the thermocline in such a way that the resulting state will be stable.

When the annual cycle is included in the mean basic state, tests of the stability of the mean state indicate that the unperturbed full nonlinear coupled system is stable during the early part of the year. The perturbed system, oscillating in the ENSO mode, continues to oscillate through the early part of the year because the ENSO limit cycle is a stable one (even though the unperturbed basic state is sometimes stable and sometimes unstable).

The predictions of Cane *et al.* (1986) and Cane & Zebiak (1987) may be interpreted as tests of the stability of the initial state. They basically initialize a state at time t_0 and then allow the model to run free to make predictions for times $t > t_0$. (In practice, it is impossible to initialize the ocean from available data so they run their ocean model with (imperfectly) observed winds for time up to t_0 then allow the atmosphere to come into equilibrium with the ocean at t_0, and then allow the coupled system to run free for time $t > t_0$.) Inspection of a large number of their hindcasts indicates that, over time periods of order two years, the predictions from a given state either do or do not go into an ENSO cycle. We would hypothesize that stability tests of their states at time t_0 would indicate which path the coupled model would choose. The success of the Cane–Zebiak model at prediction times up to two years indicate that a comprehensive and systematic understanding of the stability properties of different states throughout the year and throughout the ENSO cycle, and the ability of noise to allow the system to reach different states in this model would be very worthwhile.

5. Conclusion

We note that although progress in understanding instabilities in simple coupled models has been rapid, many issues remain in applying these results to more realistic and comprehensive numerical models (and to nature). Coupled atmosphere–ocean modelling with GCMs is a field in its infancy, primarily because of the enormous computing resources that these models require, so that not many examples of coupled ENSO cycles exist to be analysed. We can, however, anticipate some of the questions that will arise in applying these stability ideas to the more complicated models.

First, we should note that the thermodynamic equation for SST, so important for determining the nature of the coupled instabilities, is subtle in ocean GCMs and not easily interpreted. Ocean GCMs do not usually have embedded mixed layers so that mixing processes very near the surface may have to simulate the effects of upwelling (which must go to zero) in the uppermost grid level. Clearly, embedding mixed layers in ocean GCMs would not only bring the models closer to reality, but also would aid in interpreting coupled instabilities.

Second, ocean GCMs allow a number of vertical modes. How the existence of a number of vertical modes affects SST changes in response to winds is not thoroughly understood. Clearly oceans consisting of simple two layer models with surface layers embedded could be used to investigate this question and the instabilities that arise in coupling such an ocean to atmospheric models could be looked at.

Third, the effect of resolution in ocean GCMs is likely to be important for determining the form of the coupled atmosphere–ocean instabilities. If, for example, in the presence of adequate meridional resolution the mechanism of instability in the model would be that of the retarded oscillator, the coarsening of meridional resolution may no longer allow the resolution of the grave meridional modes and the signal propagation mechanisms inherent in the instability may no longer be operative. The nature of the coupled instability in the reduced resolution case is then likely to be radically different.

Finally, we note the importance of the western boundary reflections in the retarded oscillator mechanism. Even when meridional resolution is high, it is not known what is the effect of realistic geography on the reflection coefficients at the western boundary. As we saw, low-reflection coefficients were able to change the fundamental balances in the model ENSO cycle.

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