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Targeted Training Activity: ENSO-Monsoon in the Current and Future Climate

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THEORIES OF ENSO

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THEORIES OF ENSO

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1. TWO GENERAL CATEGORIES (NOT AS DIFFERENT AS THEY SEEM)

UNSTABLE
STABLE

2. ATMOSPHERE-OCEAN PROCESSES IN TROPICS

THE HEAT FLUX AT THE OCEAN SURFACE
THE PROCESSES THAT CHANGE SST
HOW THE THERMOCLINE CHANGES IN RESPONSE TO WINDS
HOW SURFACE WINDS CHANGE IN RESPONSE TO SST
THE CANE-ZEBIAK MODEL

3. UNSTABLE ENSO

THE BATTISTI VERSION OF THE CANE-ZEBIAK MODEL
THE DELAYED OSCILLATOR
WHAT SETS THE TIME SCALE?

4. STABLE ENSO

NON-NORMAL EVOLUTION: THE ROLE OF ATMOSPHERIC NOISE

5. THE INTERRELATIONSHIP OF TIME SCALES:

INTRASEASONAL

INTERANNUAL

DECADAL

CENTENNIAL

b. Atmosphere-Ocean Processes in Tropics

Heat Fluxes at the Sea Surface:

►Surface Heat Balance (all quantities positive upward):

$$-R_{net} = LE + S + Q$$

Where Q is the (sensible) heat flux through the sea surface.

[-Q therefore can be considered as the amount of net radiation NOT balanced by evaporation and sensible heating to the atmosphere]

►-Q Opposes SST Anomalies in the Equatorial Tropical Pacific

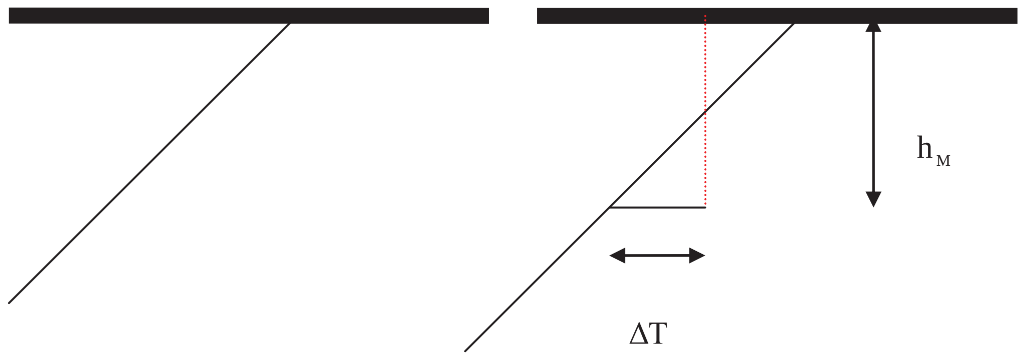
[When the SST gets cooler, LE decreases thereby increasing –Q;
When the SST gets warmer, LE increases thereby decreasing –Q]

►Since -Q opposes SST changes, the processes that change SST cannot be due to -Q and must therefore be sought IN THE OCEAN.

How SST Changes:

► Assume there is a mixed layer everywhere and the SST is the temperature of the mixed layer

► Assume there is a stable discontinuity at bottom of mixed layer:



►**The entrainment velocity at the base of the mixed layer**

$$w_e = dh_M/dt + w$$

is the volume of water crossing the (possibly moving) bottom of mixed layer per unit area per unit time.

►**The heat flux at the bottom of the mixed layer is**

$$w_e \Delta T$$

► **SST is calculated as the temperature of the mixed layer:**

$$\rho C h_M [\frac{\partial T}{\partial t} + v \bullet \nabla T + w_e \Delta T] = -Q$$

where w_e is the entrainment velocity through the bottom of the mixed layer. [ΔT is the difference between the SST and a subsurface temperature which depends on thermocline depth].

How Surface Winds Change Thermocline Depth:

Consider a two layer ocean with very deep (effectively infinite) lower layer of mean depth H and density $\rho - \Delta\rho$

$$u_t - \beta y v = -g\, h_x$$

$$v_t + \beta y u = -g\, h_y$$

$$h_t + H\,(u_x + v_y) = 0$$

where H is the equivalent depth (about 1m in equatorial Pacific).

Effective scales in the tropics are:

$L = (c/\beta)^{1/2}$, $T = (\beta c)^{-1/2}$ where $c = (gH)^{1/2}$ so that $c = L/T$

[The mid-latitude radius of deformation is $c/(\beta y)$. As we approach the equator, the equatorial radius of deformation grows until it reaches a value $c/(\beta L) = L$ which gives an interpretation of the equatorial radius of deformation L]

The shallow water equations scaled by L and T are:

$$u_t - \beta y v = - h_x + F$$

$$v_t + \beta y u = - h_y + G$$

$$h_t + (u_x + v_y) = -Q$$

Free Waves (F=G=Q=0):

A single equation can be derived for v:

If we now consider solutions exp[i(kx-ωt)], the equation becomes

$$V_{yy} + [\omega^2 - k^2 - k/\omega - y^2] = 0$$

The solutions on an infinite equatorial beta plane are the parabolic cylinder functions:

$$v_n = H_n(y) \exp[-y^2]$$

with solutions only when

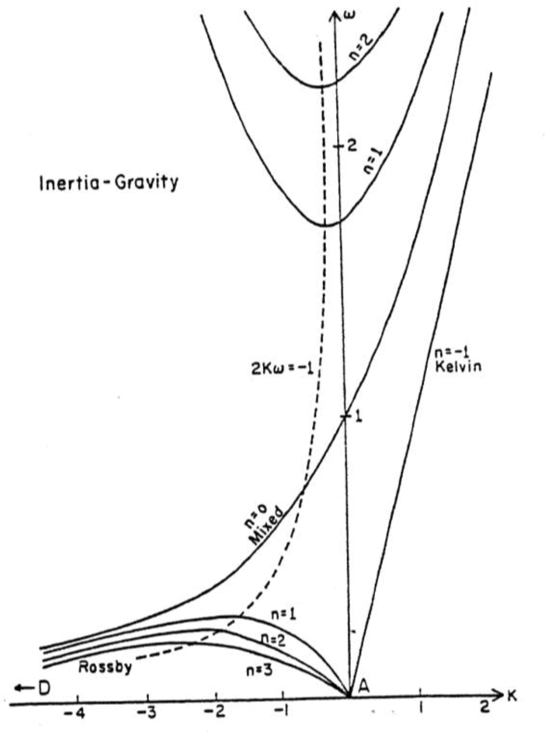
$$\omega^2 - k^2 - k/\omega = 2n+1$$

At low frequencies, large scales, $- k/\omega = 2n+1$ and the speed of Rossby waves is westward with speed $c_n = -c/(2n+1)$.

There is another solution, v=0, which has the form

$$u = h = \exp[-y^2]$$

and is the Kelvin wave: it travels eastward with speed c.



We will mostly be interested in low Frequency, long wavelength motion:

$$c_n = -c/(2n+1)$$

c_K=c=2.7m/sec in Pacific

[L~250km, T~1day]

Reflections:

► At the Western Boundary [where is the western boundary in the tropical Pacific?] any incoming (westward) Rossby waves reflects as a Kelvin wave plus a thinning boundary layer composed of short Rossby waves with group velocity to the east.

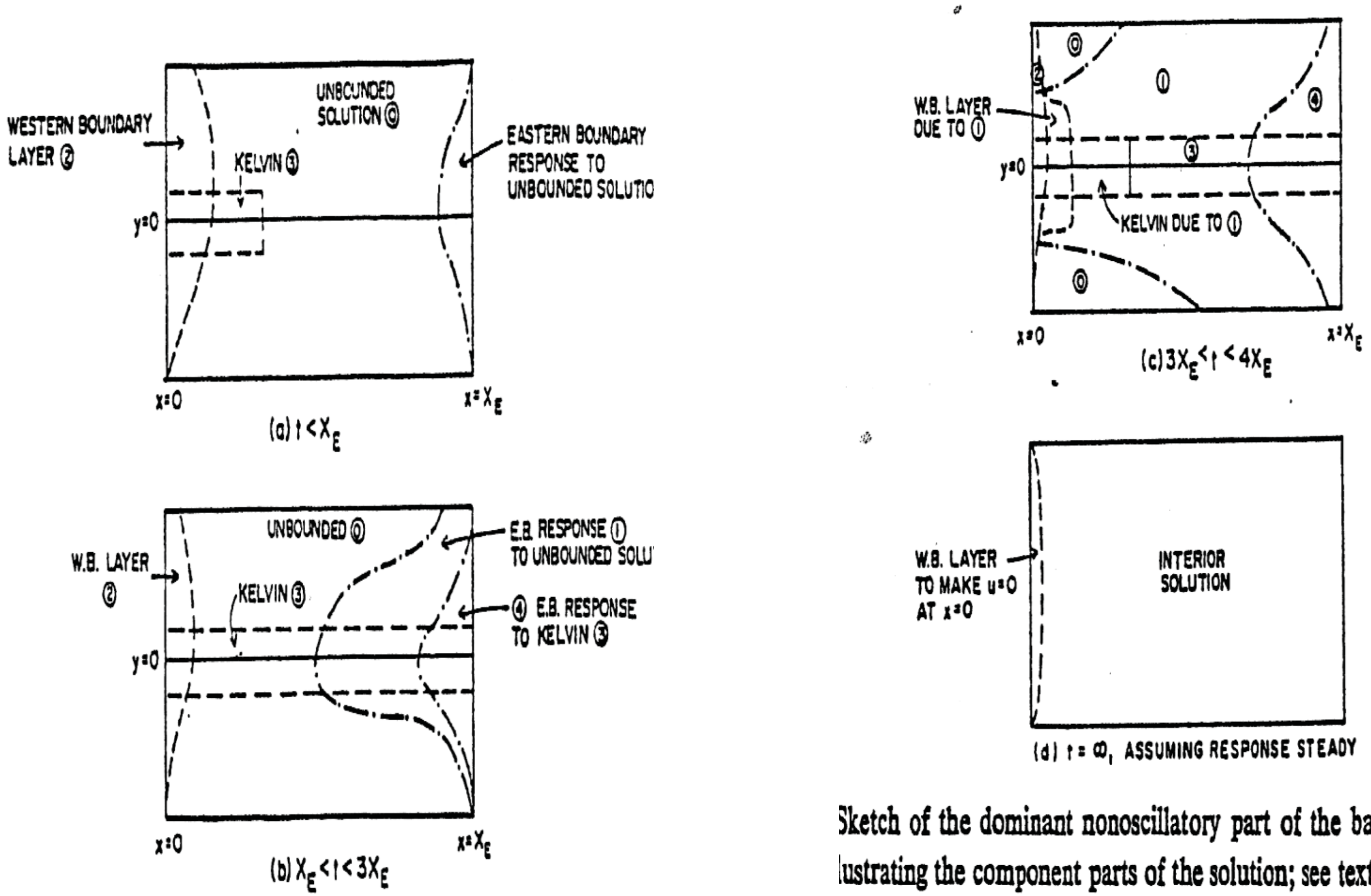
► At the Eastern Boundary, any incoming Kelvin wave gets reflected as an infinite sum of westward propagating Rossby waves.

Adjustment:

► For a constant forcing F, a solution to the shallow water equations is:

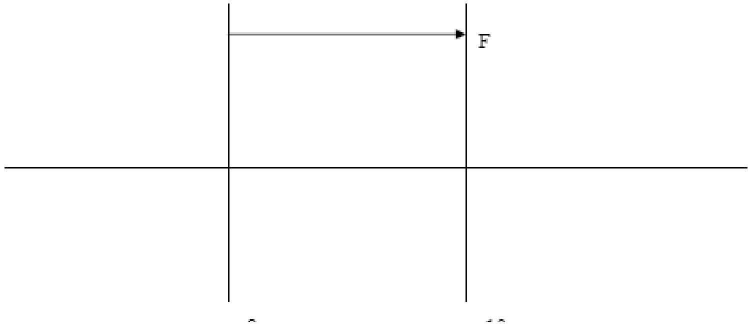
$$h_x = F,\ u = v = 0\ \ [\text{but also } h_t = F \text{ if } h_x = 0\}$$

► The adjustment proceeds behind wave fronts traveling with the Kelvin and Rossby wave speeds.

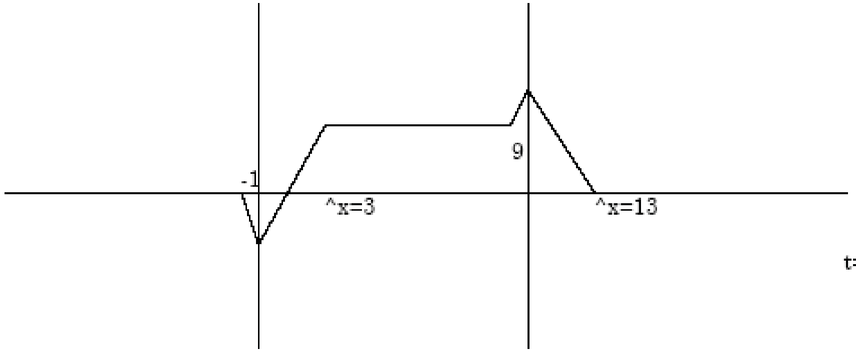


Sketch of the dominant nonoscillatory part of the baroclinic respon
lustrating the component parts of the solution; see text.

► **After a time it takes a Kelvin + Rossby wave to travel across the forcing region, the adjustment is complete but wave fronts continue outside the forcing region.**

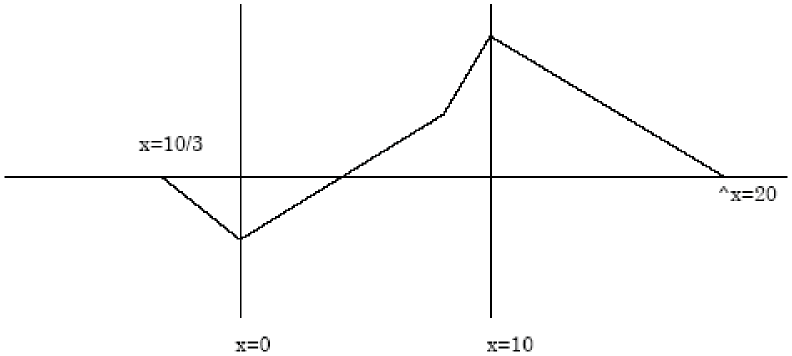


t=0

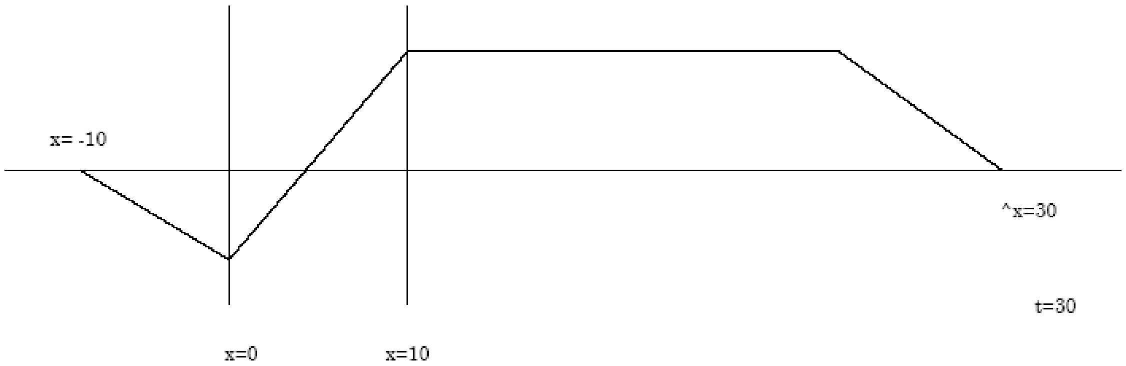


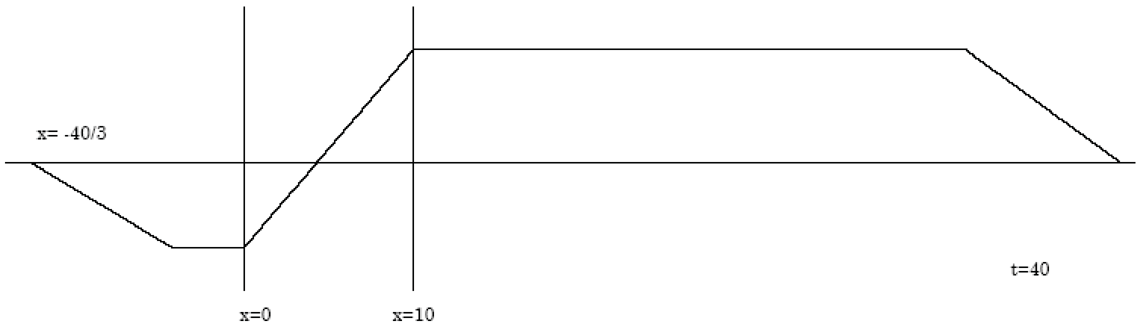
t=3

At t=30/4=7.5. the fronts meet in the wind region and there is no longer any flat part of the thermocline



t=10


x= -10
^x=30
t=30
x=0
x=10


x= -40/3
t=40
x=0
x=10

How Surface Winds Change

► We have a region on persistent precipitation over warm water (>28º).

► The heat released in the deep cumulonimbus clouds forces the atmosphere.

► The Gill model is the standard tool for looking at the response.

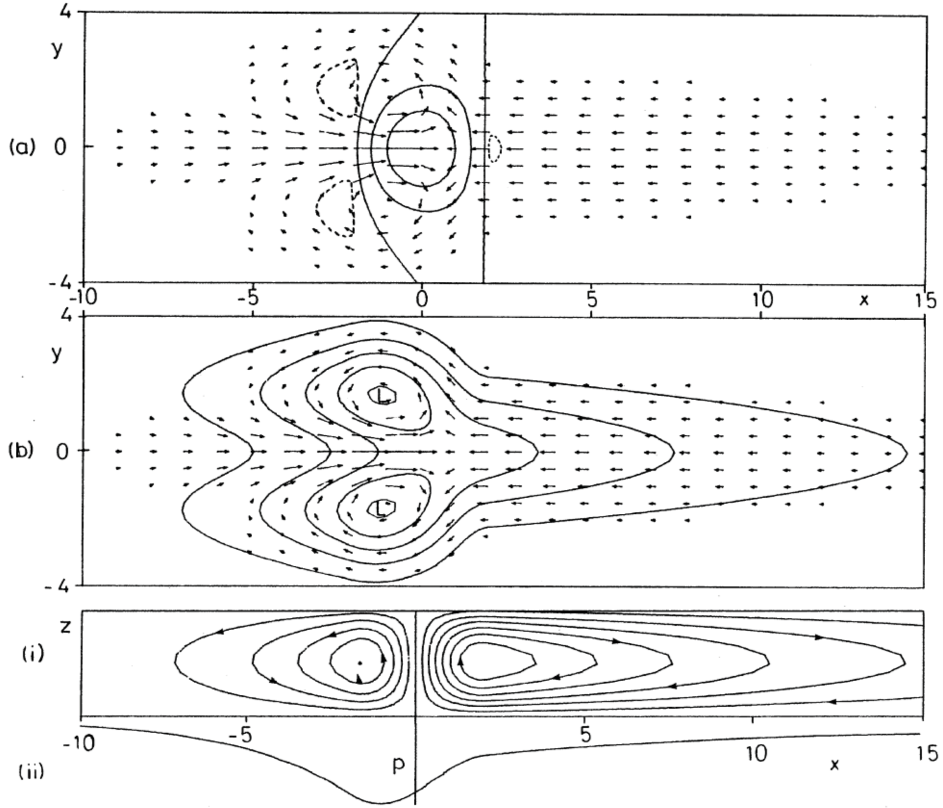
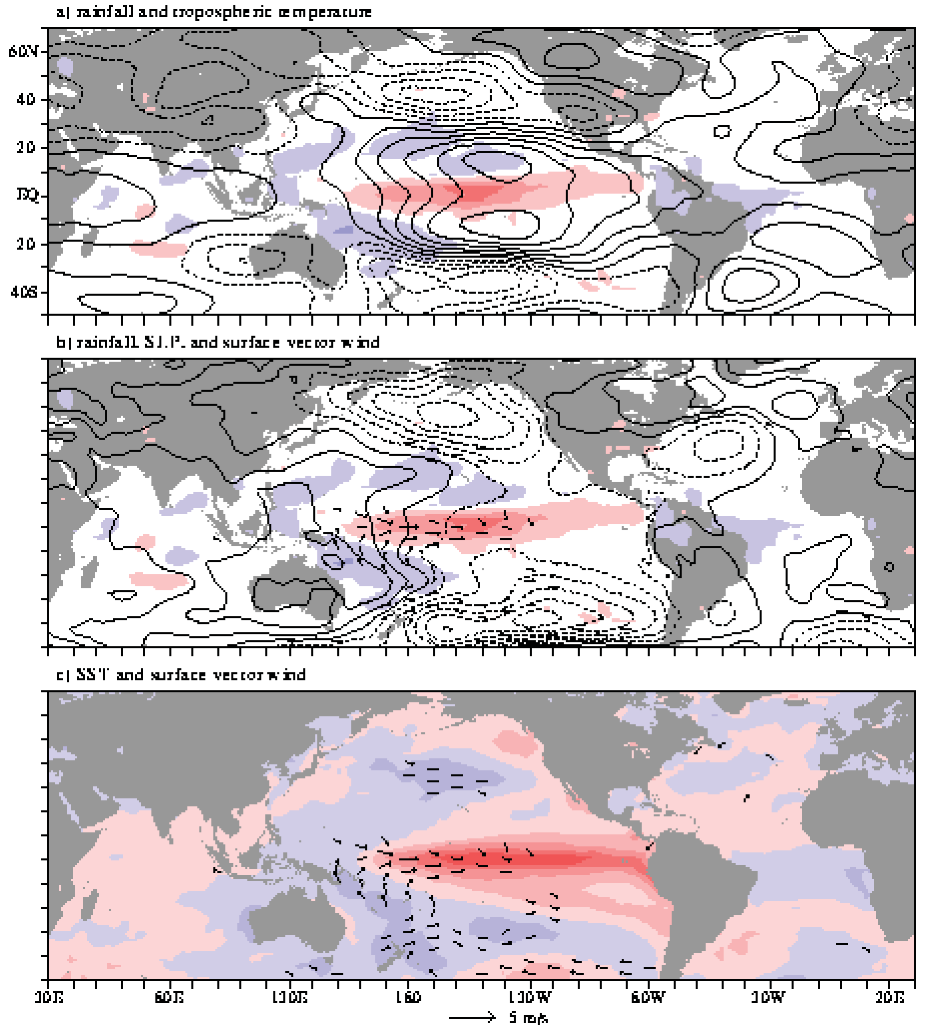


Figure 1. Solution for heating symmetric about the equator in the region $|x| < 2$ for decay factor $\varepsilon = 0{\cdot}1$.
(a) Contours of vertical velocity w (solid contours are 0, 0·3, 0·6, broken contour is −0·1) superimposed on the velocity field for the lower layer. The field is dominated by the upward motion in the heating region where it has approximately the same shape as the heating function. Elsewhere there is subsidence with the same pattern as the pressure field.

(b) Contours of perturbation pressure p (contour interval 0·3) which is everywhere negative. There is a trough at the equator in the easterly régime to the east of the forcing region. On the other hand, the pressure in the westerlies to the west of the forcing region, though depressed, is high relative to its value off the equator. Two cyclones are found on the north-west and south-west flanks of the forcing region.

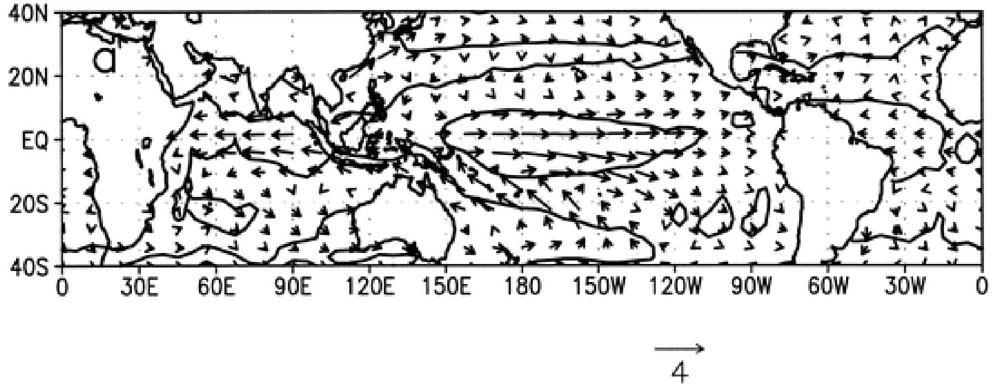
(c) The meridionally integrated flow showing (i) stream function contours, and (ii) perturbation pressure. Note the rising motion in the heating region (where there is a trough) and subsidence elsewhere. The circulation in the right-hand (Walker) cell is five times that in each of the Hadley cells shown in (c).

►The Gill model implies westerly winds to the west of the heat sources and strong easterly winds to the east: the easterly winds to the east are not seen



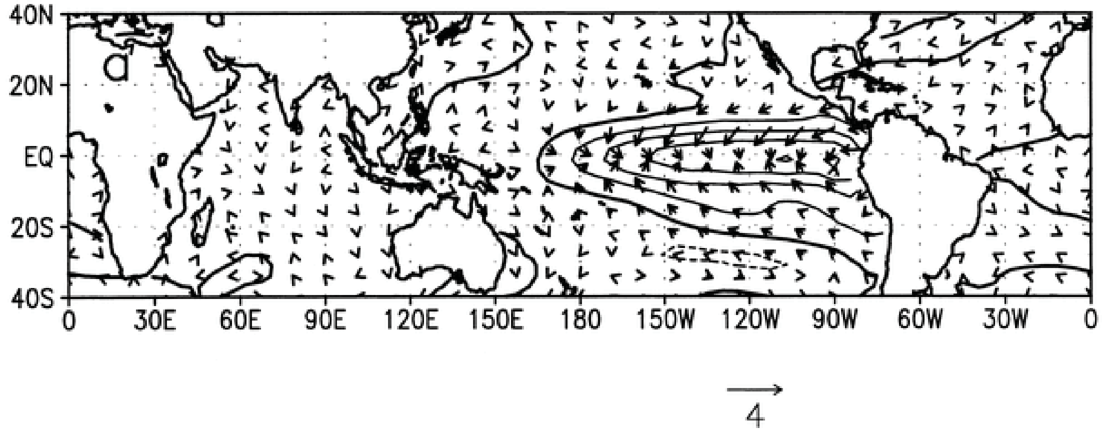
►The Gill model is wrong in its essentials: the actual surface winds have to be radiated (or otherwise transmitted) down from the cloud base which is usually at 600m—this problem solved by Z. Wu et al.

►The wind response is well calculated by models (Chiang, 2001)

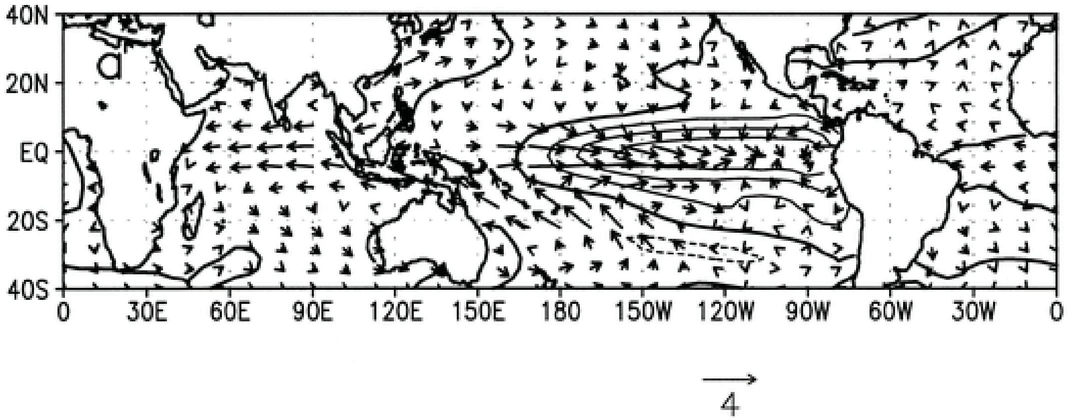


►Atmospheric Boundary layer gradients also affect surface winds.

Chiang, 01



►The total surface response is the sum of these and compares well with observations:



►Basic point: The surface response is westerly winds to the west of the heat source.

The Cane-Zebiak Model

► The CZ model is the simplest model that contains all the elements of atmosphere-ocean interactions in the tropics

► It has:

- A shallow water model in a square basin to calculate the response of the thermocline in response to wind anomalies
- A fixed 50 m mixed layer to calculate SST anomaly evolution
- A parameterization of subsurface temperature in terms of the thermocline depth:

$$T_{sub} = [\tanh \{\lambda(h_{mean}+1.5h)\} - \tanh \lambda h]$$

- A slightly modified Gill model parameterization of the surface wind response to SST anomalies

► What makes this all work is that the annual cycle is specified and only the anomalies are calculated.

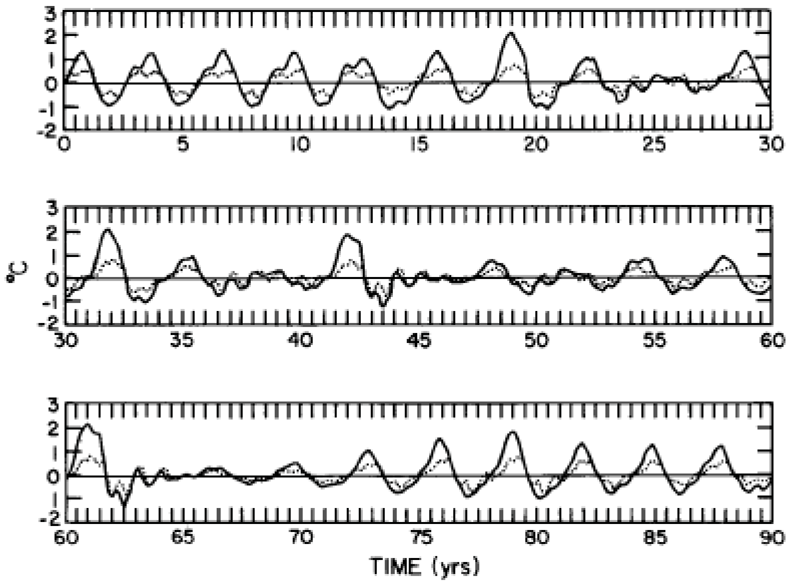


FIG. 1. Area-averaged SST anomalies for the 90-year model simulation. The solid line is NINO3 (5°N–5°S, 90°–150°W), and the dotted line is NINO4 (5°N–5°S, 150°W–160°E).

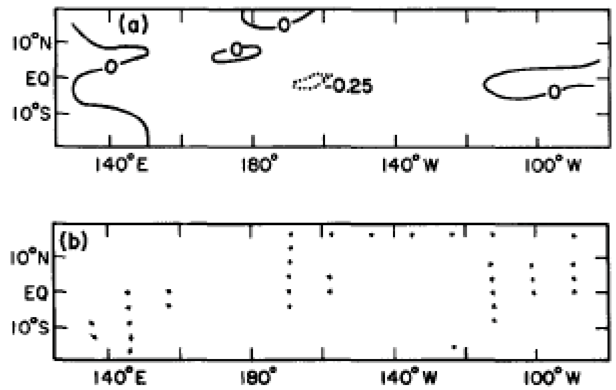


FIG. 4. (a) SST anomalies and (b) wind anomalies in December of year 30 of the model simulation.

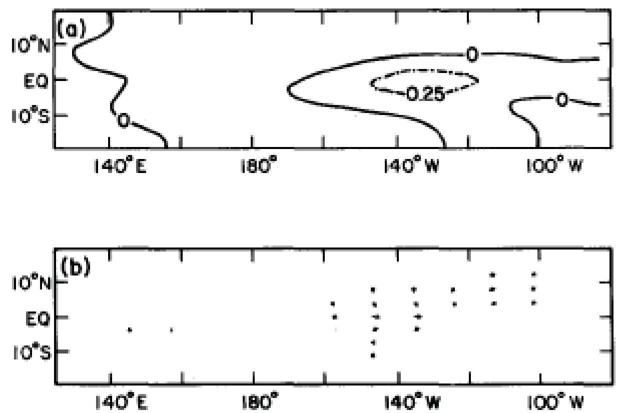


FIG. 5. As in Fig. 4, except for March of year 31.

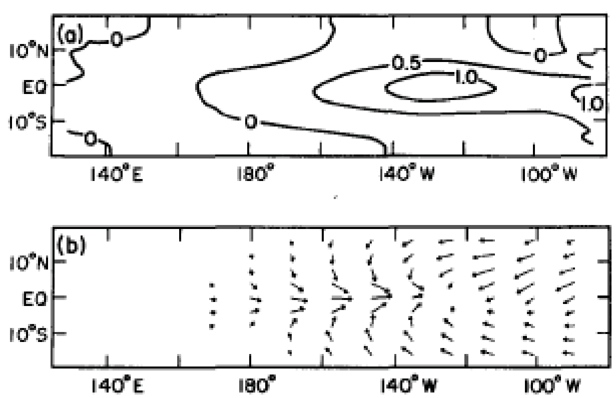


FIG. 6. As in Fig. 4, except for June of year 31.

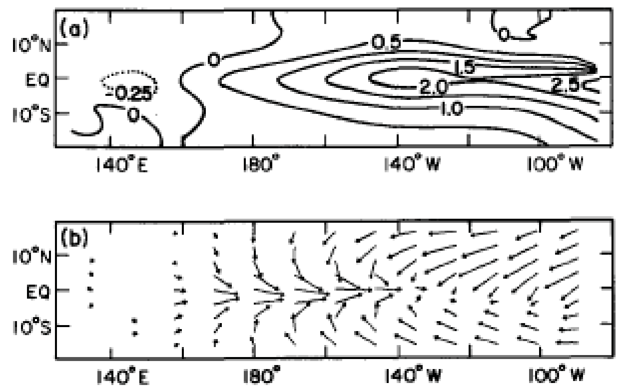


FIG. 7. As in Fig. 4, except for September of year 31.

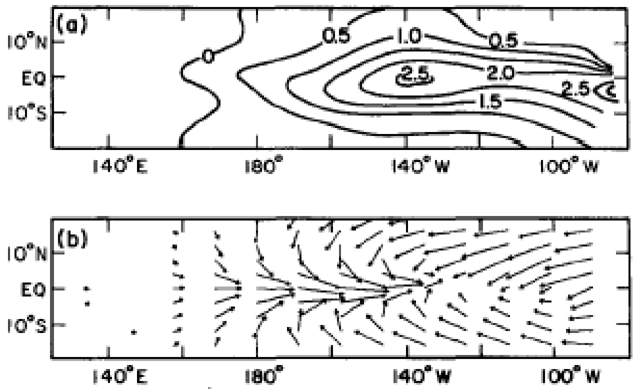


FIG. 8. As in Fig. 4, except for December of year 31.

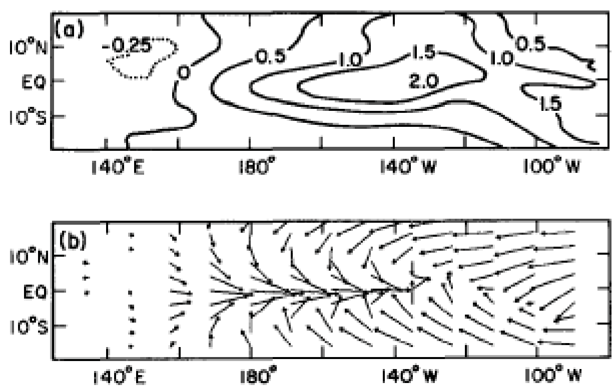
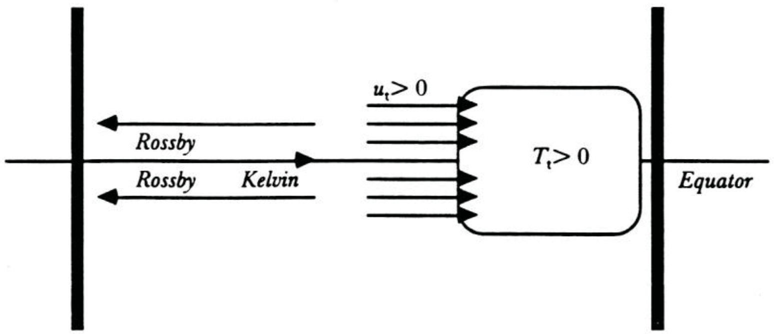


FIG. 9. As in Fig. 4, except for March of year 32.

►The great advantage of the CZ model is that it is simple enough to be run for very long times and simple enough to be analyzed for the ENSO mechanisms

c. Unstable ENSO

►A slight variant of the CZ model was built by Battisti and the subsequent analysis of the model indicated that the ENSO mechanism was well described by The Delayed Oscillator:



1. The warm anomaly induces westerly (eastward) winds to the west of the warm anomaly. The westerly winds forces the warm anomaly warmer and the unstable system grows.

2. The thermocline adjusts to deep in the eastern part of the wind patch (downwelling) and shallow in the western part (upwelling).
3. The downwelling signal is carried eastward by the Kelvin mode and is then dissipated in the mode fronts at the eastern boundary.
4. The upwelling signal travels westward as a Rossby mode, reflects as an upwelling Kelvin mode, and travels back to the scene of the growing instability.
5. The returning Kelvin wave eventually overcomes the effect of the growth if it is large enough.
6. The time scale is determined by the competition between the returning Kelvin mode and the growing instability.
7. The process is well represented by the delayed oscillator equation:

$$\frac{dT}{dt} = aT(t) - bT(t-\tau)$$

with b>a.

8. **An additional (non-linear) term is needed in the delayed oscillator to equilibrate the amplitude.**

9. **The mechanisms is essentially different from the one by Schopf and Suarez which has b<a.**

► **There are other mechanisms for getting interannual variability depending on the thermodynamic relationship [e.g. $T = h$, $T_t = vT_x$ ...] and some coupled GCMs seem to have these alternate versions of ENSO.**

► **Large complex coupled GCMs (e.g. GFDL, CCSM) have problems getting the annual cycle correct in the tropics and mostly do not get ENSO correct.**

► **The CZ model remains a good laboratory for testing ideas and for prediction (as we will see, its ability to predict ENSO is as good as any other model).**

► **Its mechanism for irregularity is an interactions with another mode.**

d. Stable ENSO

► The CZ model can be linearized and formulated as a linear matrix equation:

$$\frac{dx}{dt} = Ax + Noise$$

where all the eigenvalues of A are negative so there are no unstable normal modes

► So how can there be ENSO in a stable system?

The solution of the linear equation is

$$x(t) = R(t)\ x(t=0)$$

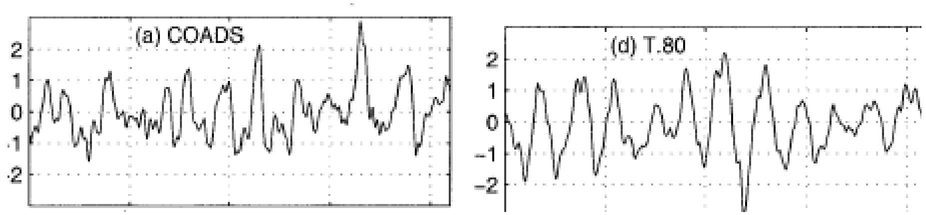
where R(t) = exp [At] is the propagator (which takes the initial solution to the solution at time t).

►The propagator R is non-normal [$RR^+ \neq R^+R$] and the eigenvectors are not orthogonal. This allows growing modes for a limited amount of time which then decay.

►A non-normal system can maintain a very much larger variance for a given input of noise than a normal system.

►The growing non-normal disturbances depend on the background flow.

►Thompson and Battisti showed that the linearization of the CZ model driven by random noise has many of the same properties as ENSO



►The growing modes have features of the delayed oscillator but irregularity is inherent and doesn’t have to be explained.

e. The interrelationship of time scales

► ENSO is modulated on scales of decades.

►It is not yet known how ENSO will react to global warming

►ENSO does contribute to global warming—the tropical troposphere warms by about 1K during warm phases of ENSO

►Because the tropics is so crucial to the climate of the world, ENSO must be gotten correctly for any coupled GCM to work

►There is a limited amount of predictability of the world's climate characteristic of ENSO.