

Climate Predictability

Part I: Predictability of The Second Kind:

An Introduction to the Role of External Forcing

Part II: Seasonal Forecasting: The mid-latitude response to ENSO

Part III: Inter-annual predictability of ENSO

Part IV: Decadal Predictability

Part V: Role of greenhouse gases in climate change

What physical factors are responsible for the *variability* in seasonal means?

- (1) Flow instabilities (internal dynamics) which always exist
- (2) Variations in the slowly changing boundary conditions themselves - **Why?**

SST- because of the higher heat capacity of the ocean, the surface air temperature $T_s = \text{SST}$. Higher SST means high T_s , hence higher saturation vapor pressure e_s , defined as the maximum partial pressure of water vapor possible without condensation taking place (alternatively, as the maximum weight of water vapor that a unit weight of air can hold).

Higher gradients of T_s also lead to surface which in turn increases the likelihood of condensation \Rightarrow **more rainfall !**

The evaporation from the ocean removes energy from the surface, and the condensation releases that latent energy into the atmosphere.

But, the latent heat release will in general occur at a different (x,y,p) compared to the location of the evaporation - ***a heat source has been added somewhere to the interior of the atmosphere!***

Soil Moisture - Increased soil moisture also leads to increased evaporation - and so also gives a path to increased precipitation and a latent heat source for the atmosphere

Vertical heat transport: increased soil moisture means that the incoming (solar + downward long wave) radiation goes towards evaporation, and not towards heating the ground directly. So from the atmosphere's point of view, the increased heating takes place in the interior (latent heating) rather than at the surface

In summary, both SST and soil moisture changes lead to changes in the latent heat sources in the atmosphere. The subsequent change in atmospheric seasonal means occurs for 2 reasons:

- (1) Direct Response of the mean flow to changes in heating
- (2) Indirect Response; the geographically preferred region of instabilities will shift as the mean flow changes (vertical shear, horizontal shear)

Indirect Mechanism:

- The flow instabilities (internal dynamics) plays less of a role in causing variability of seasonal means in the tropics (compared to mid-latitudes) because the instabilities are weaker.
- Baroclinic instability (growth of baroclinic waves) is strongest in regions of vertical wind shear / horizontal temperature gradients
- Horizontal temperature gradients are much weaker in the tropics; vertical wind shears are therefore also weaker (thermal wind relationship)

Conventional View:

- Convective instabilities in the tropics grow and saturate in a very short time - and here there *is* a distinct time scale separation- convective time scale is much shorter than a season.
- Seasonal tropical mean rainfall is due to mean of all convective instabilities, and this is well controlled by SST

How does boundary forced variability lead to predictability?

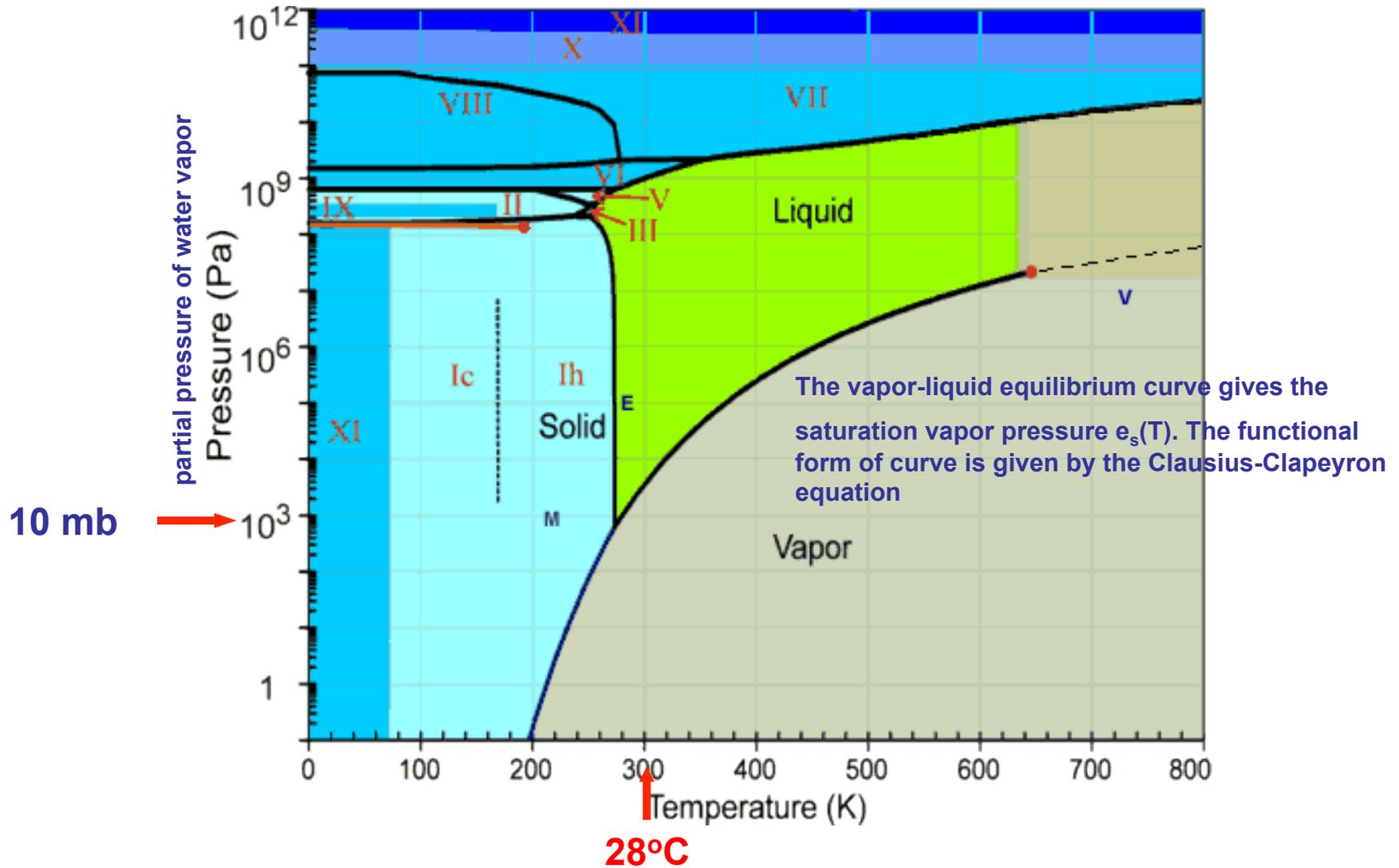
Variability due to **internal dynamics** is expected to be **unpredictable** on the seasonal time scales. This time scale is beyond the the predictability limit for predictability based on initial conditions - *in fact internal dynamics severely limits the predictability of seasonal means in mid-latitudes!*

Variability due to **boundary condition** changes **is potentially predictable**, since the boundary condition changes occur on a very slow time scale.

How can internal dynamics (instabilities, wave-wave interaction), lead to (unpredictable) variations of the seasonal mean?

- Although individual instabilities or mid-latitude storms have time scales less than a season, there are lower frequency phenomena whose time scale is not much less than a season. These include blocking events and variability of the jet stream, which controls the number of disturbances and their paths.
- A simple example of chaos causing unpredictable variations in long-time means:
Lorenz 1964

Phase Diagram of Water



Presence of a time scale “window”

SST and soil moisture are variables in the climate system, but they vary on time scales longer than months. So during a season, they do not change very much.

One of the most significant changes that occurs in tropical SSTs, with large consequences for mid-latitudes, is the

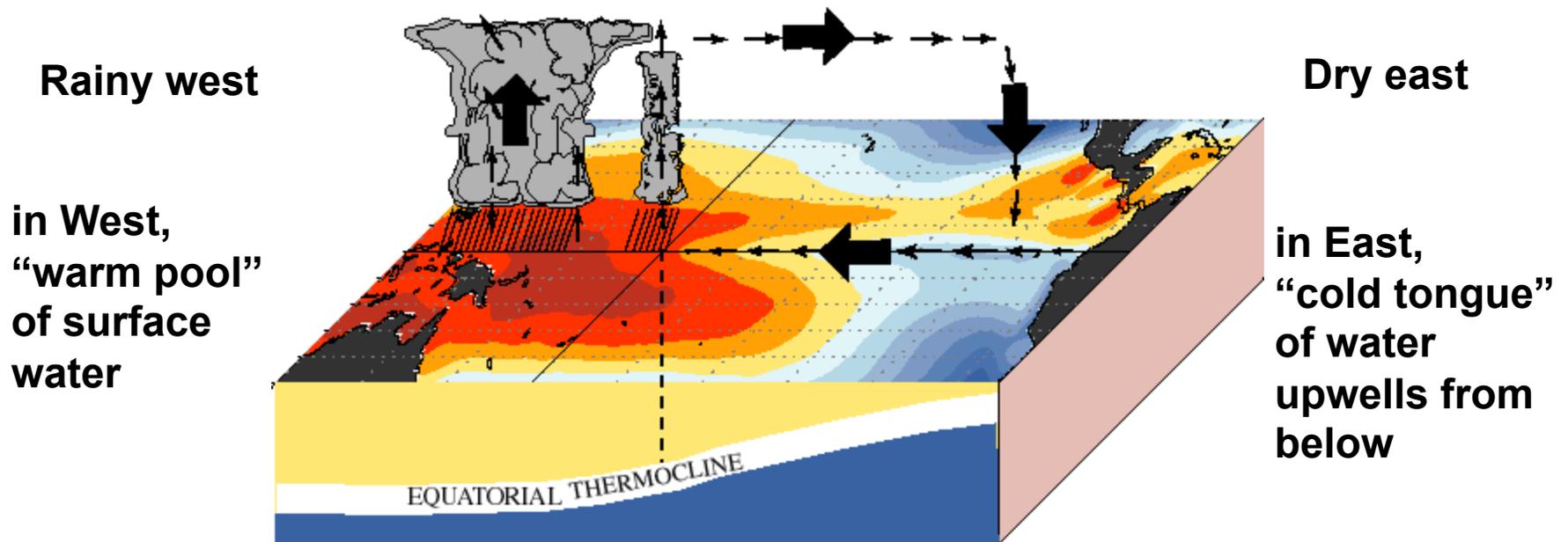
ENSO = “El-Nino Southern Oscillation”

ENSO is an irregular oscillation of the tropical atmosphere / ocean system (period of ~ 2 - 7 years)

ENSO involves *coupled ocean-atmosphere dynamics*

Normal Winter Conditions, Equatorial Pacific

“Walker Circulation” driven by sea surface temperature gradient

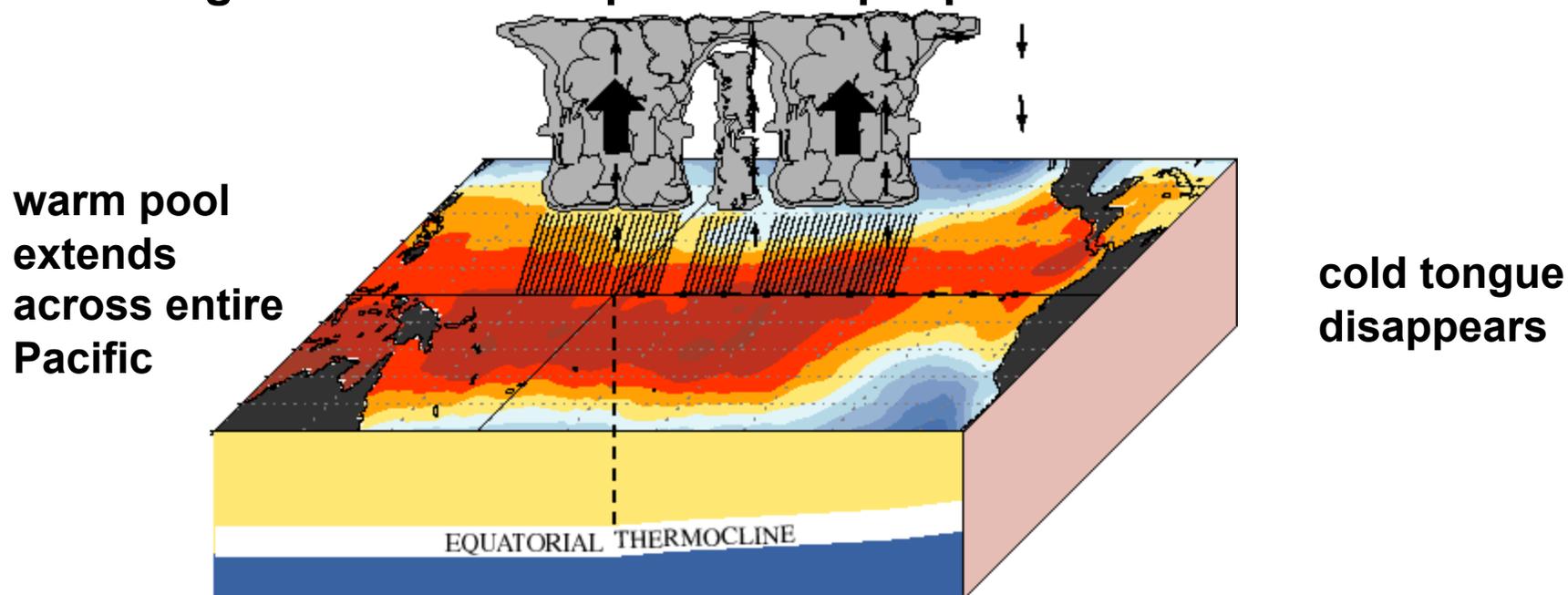


Thermocline tilt/upwelling driven by westward wind stress

www.cpc.ncep.noaa.gov/products/analysis_monitoring/impacts/warm_impacts.html

El Nino Winter Conditions, Equatorial Pacific

Warm SSTs in the Eastern Pacific --> Increased evaporation in the Eastern Tropical Pacific, increased deep convection and rainfall, increased rising motion, and finally increased tropical divergence near the top of the troposphere



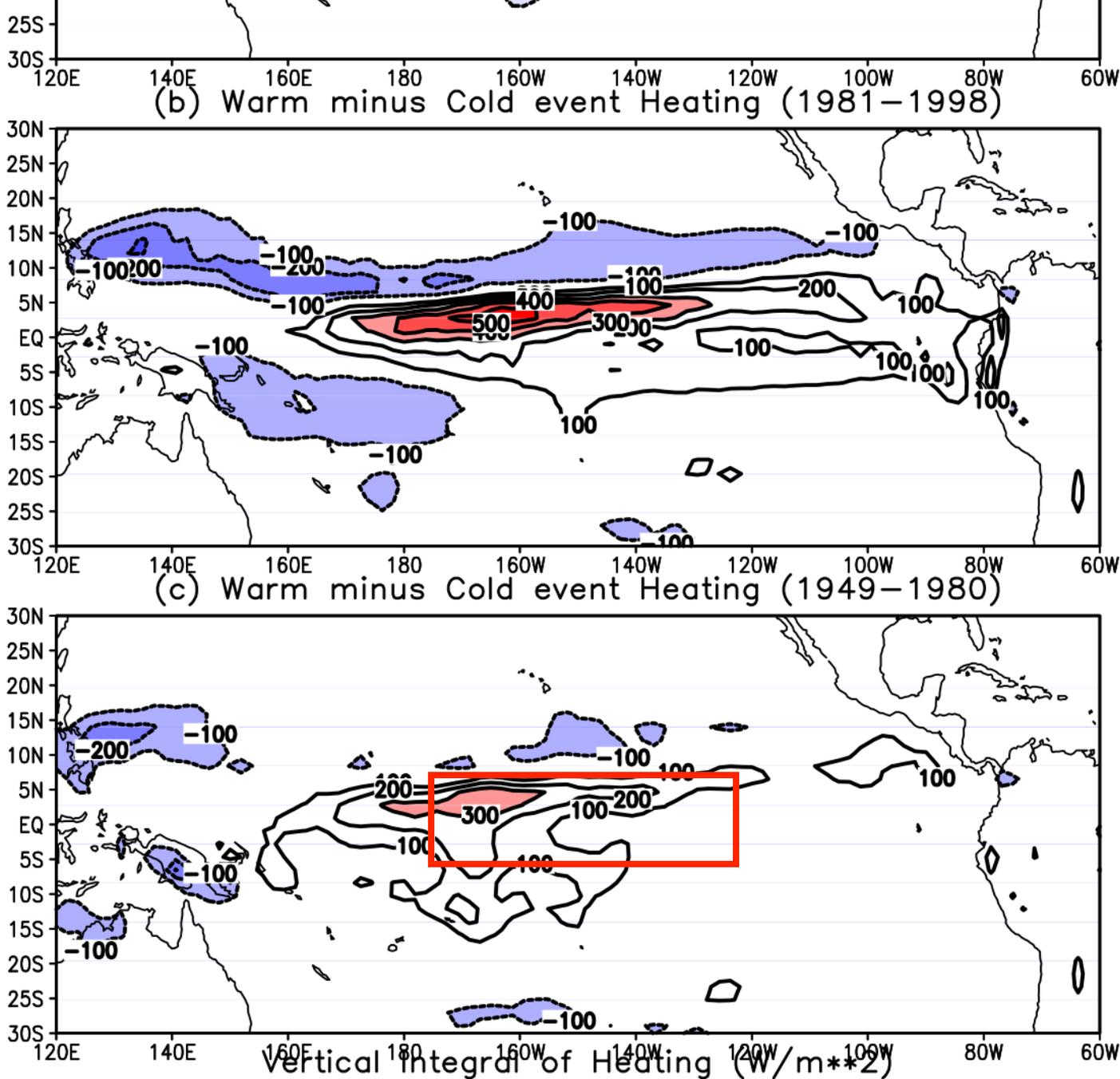
weakening of westerlies allows flattens thermocline, weakens upwelling

www.cpc.ncep.noaa.gov/products/analysis_monitoring/impacts/warm_impacts.html

ICTP Climate Predictability II

David Straus (GMU) David Straus

George Mason University / COLA



C20C
simulations
with COLA
AGCM

Warm and Cold
events identified
by value of SST
averaged over
box (Nino3.4
index)

ENSO SST variability is caused by coupled atmosphere-ocean interaction

BUT

The *effects* of these SST variations can be studied by atmospheric models (AGCMS) forced by given SST variability.

Why?

In the central and eastern Pacific, although the atmosphere and ocean interact strongly on inter-annual time scales, the ocean forces the atmosphere fairly strongly on seasonal and sub-seasonal time scales.

There are many regions where the atmosphere forces the ocean on seasonal and intra-seasonal time scales:

- The Indian Ocean - where once the summer Monsoon winds have been set up, they can lead to increased evaporation from the Indian Ocean, which leads to cooling of the ocean
- The North Pacific, where atmospheric wind stress is an important forcing of the ocean

The path for ENSO SSTs to affect mid-latitudes

Step 1: During El-Nino, the (anomalous) latent heating due to deep convection in the mid-Pacific is balanced by upward vertical motion.

To see this we examine the thermodynamic equation: $T \frac{ds}{dt} = Q$ or

$$\frac{\partial \theta}{\partial t} + u \frac{1}{a \cos(\phi)} \frac{\partial \theta}{\partial \lambda} + v \frac{1}{a} \frac{\partial \theta}{\partial \phi} + \omega \frac{\partial \theta}{\partial p} = \frac{Q}{c_p}$$

where θ is potential temperature, s is entropy per unit mass, $s = c_p \log(\theta)$

$\omega = dp/dt$ (the vertical velocity in pressure coordinates), Q is the rate of latent heating, and we also have:

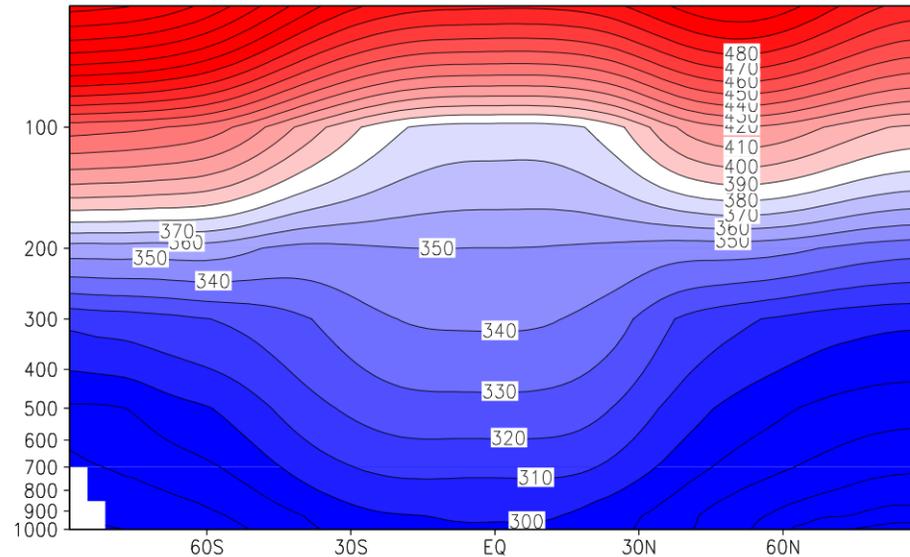
$$\theta = \left(\frac{p_0}{p} \right)^\kappa T \quad \kappa = R/c_p$$

where R is the gas constant and $c_p = 7/2 R$ the specific heat at constant pressure of an ideal gas.

The dominant balance is given in the blue box (for the tropics). Since $\frac{\partial \theta}{\partial p} < 0$

we have that $\omega < 0$, which corresponds to *upward motion*.

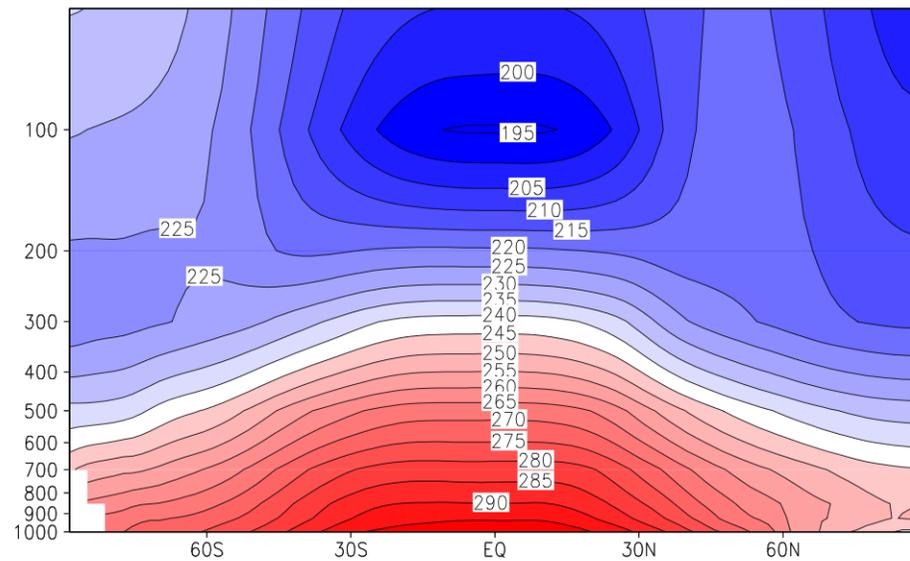
$$[\theta] = [T] \left(\frac{p_0}{p} \right)^\kappa$$



Both θ and T decrease towards the poles - but while T decreases with altitude, θ increases with altitude - hence θ decreases with pressure

$[T]$

- vertical coordinate is log (p)
- zonal means are computed at constant pressure.

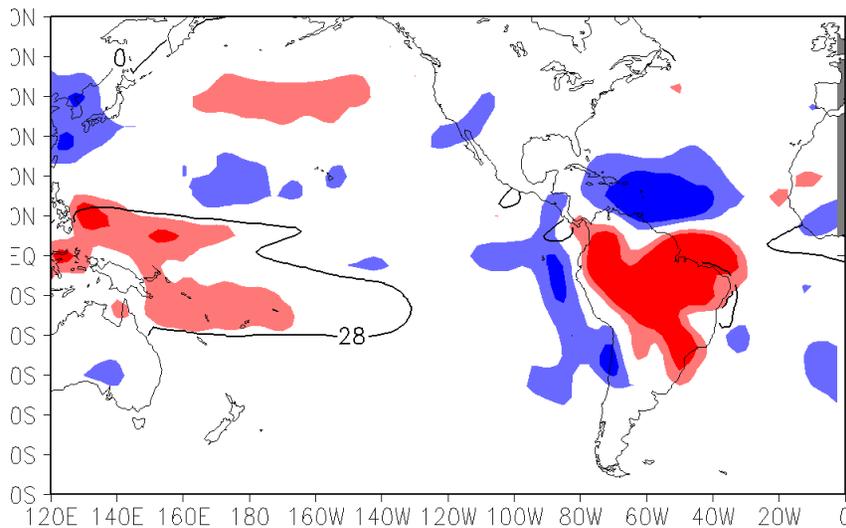


The path for ENSO SSTs to affect midlatitudes: Step 1 - continued

Increased Rising Motion in the Tropical Eastern Pacific Leads to increased upper-level (200 hPa) divergence of flow.

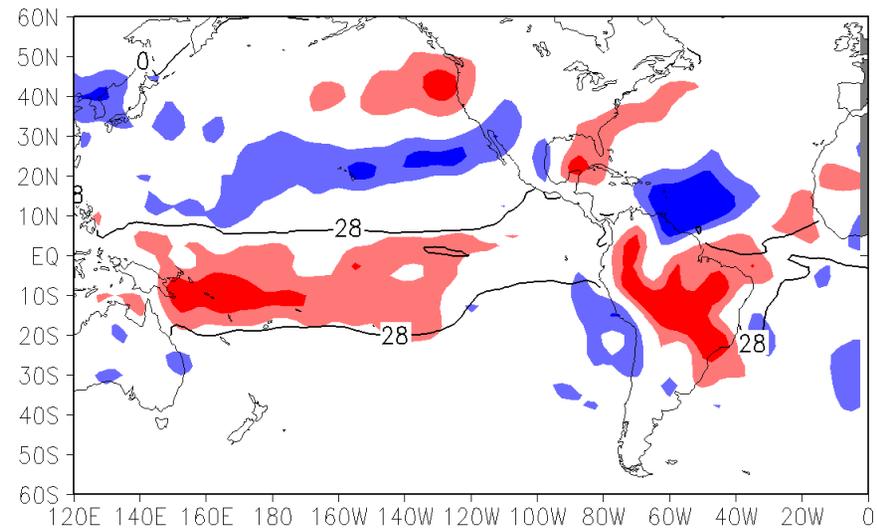
JFM seasonal mean 1982-2002

Normal Conditions for winter



Mean of JFM 1983 and 1998

Warm Event (avg. of obs)



Shading: Winter Mean upper level divergence ($Cl=2.0 \times 10^{-6} \text{ 1/s}$)

SST Isotherm of 28 C is shown (warmer water allows convection)

Step 2 of the path:

How do we understand the response to deep heating / upper level divergence anomalies?

In nature, we refer to the “stationary response” as the seasonal mean of a time-varying response.

In idealized “stationary wave” models, we look for the true steady state (time-independent) response. (In these models baroclinic instability is eliminated by using large damping / dissipation.)

Another approach:

Turn on the heating at $t=0$ in a full primitive equation model where the initial conditions correspond to the observed time mean circulation, and *measure the the mid-latitude response to the heating after 1 - 2 weeks* (before baroclinic instability has grown too strong).

The Direct Response to Tropical Heating in a Baroclinic Atmosphere by Feifei Jin and Brian J. Hoskins, *Journal of the Atmospheric Sciences* Volume 52, Issue 3 (February 1995) pp. 307–319

An example of stationary Rossby waves at upper levels

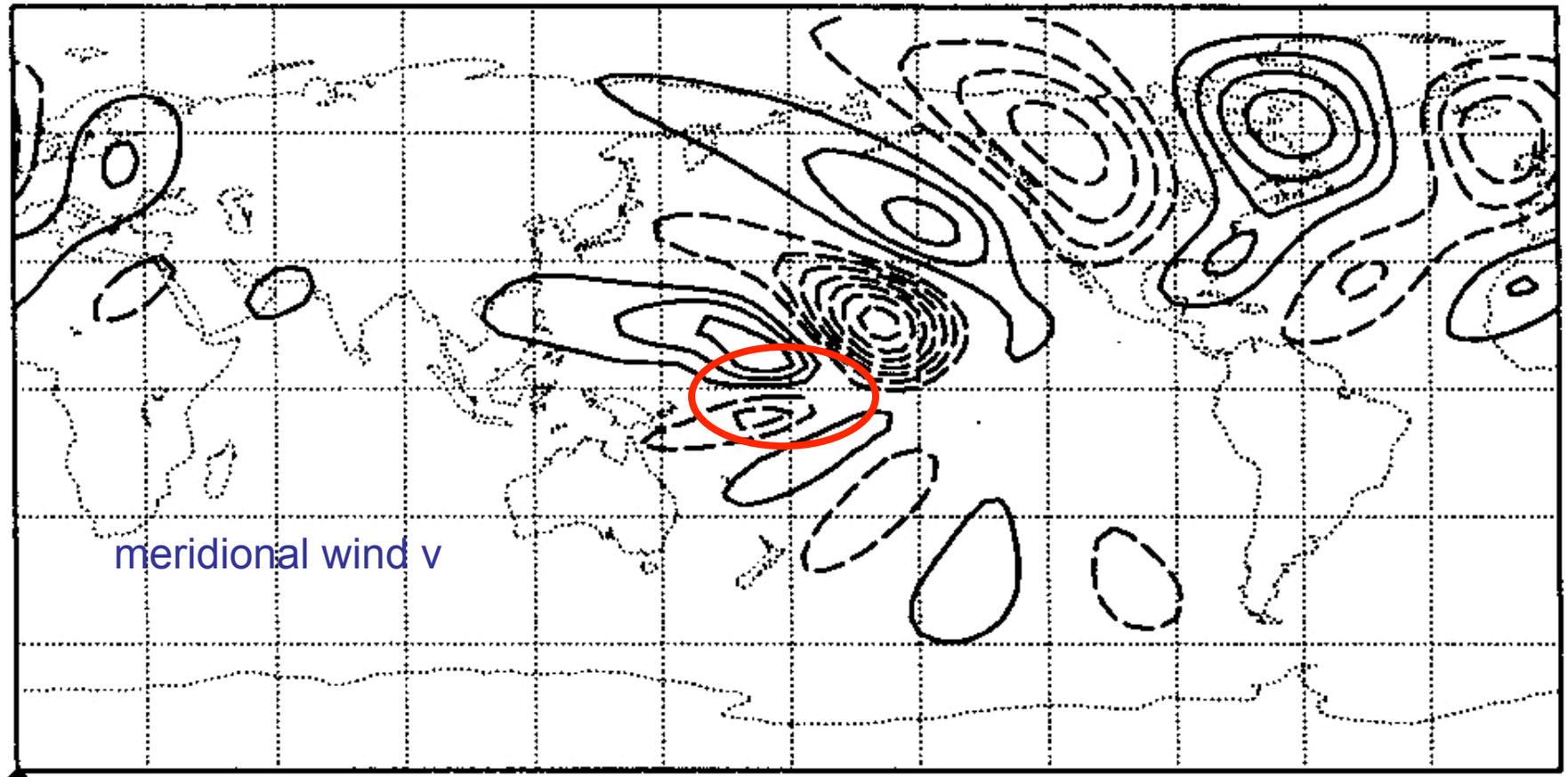


FIG. 8. Longitude–latitude picture of the day 15 $\sigma \approx 0.24$, meridional wind perturbation for the heating on a Dec–Feb zonal flow. The contour interval is 0.5 m s^{-1} . The zero contour is not shown, and the negative contours are dashed.

$$v = \frac{1}{a \cos(\phi)} \frac{\partial \psi}{\partial \lambda}$$

$$\psi = \sum_m (A_m \cos(m\lambda) + B_m \sin(m\lambda))$$

$$v = \sum_m \frac{m}{a \cos(\phi)} (A_m \cos(m\lambda) + B_m \sin(m\lambda))$$

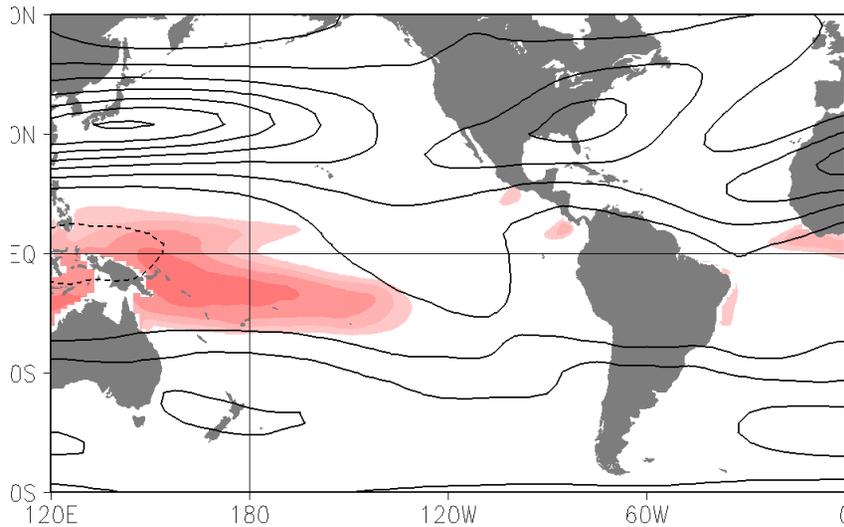
The meridional wind v emphasizes the smaller scale features compared to the streamfunction (or height) field. (Sum above is weighted by wavenumber m)

Part of interest in ENSO warm event is what it does to midlatitudes:

Extension of Pacific Jet

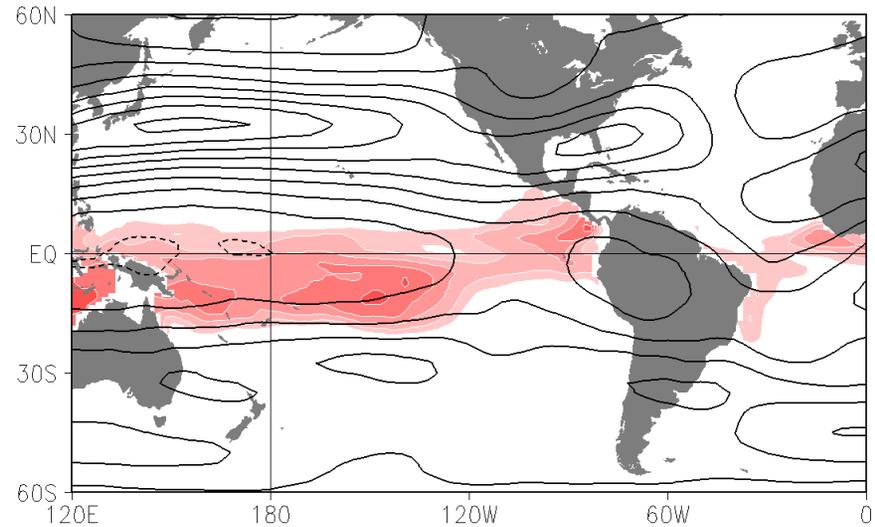
Normal Conditions:

JFM seasonal mean 1982-2002



ENSO Warm event (avg of obs)

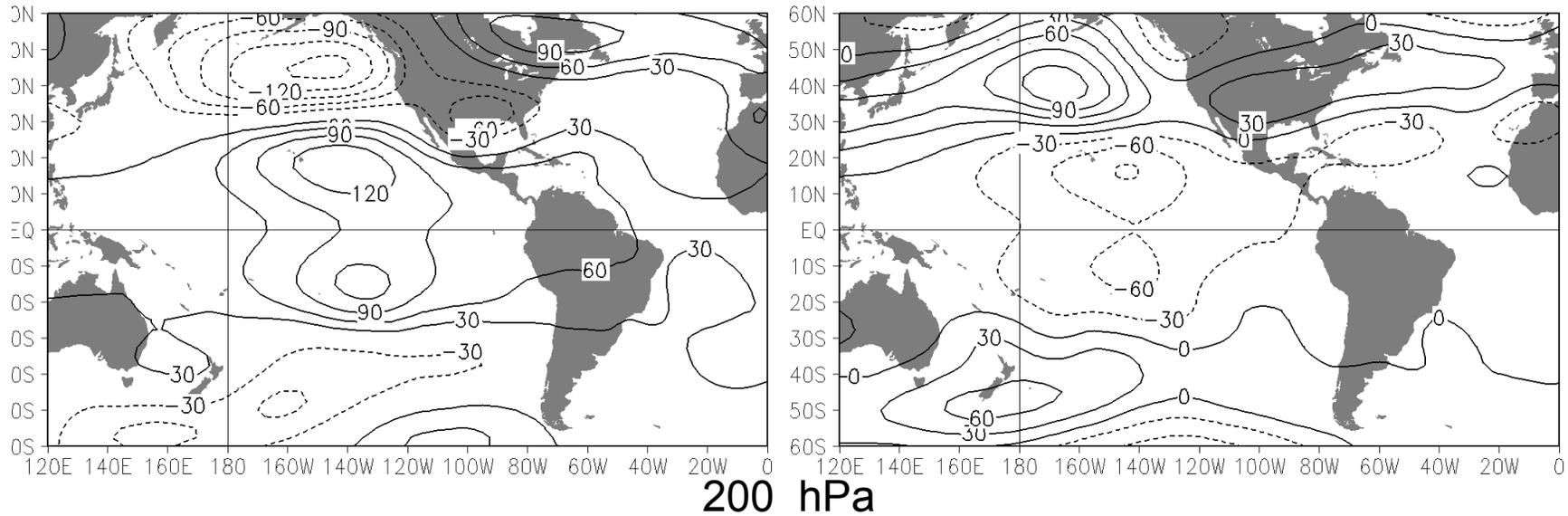
Mean of JFM 1983 and 1998



Contours are jets: Winter mean 200 hPa zonal wind (CI=10 m/s)

Shading is SST in degrees C

Height anomalies = departure of JFM seasonal mean Z from long-term mean

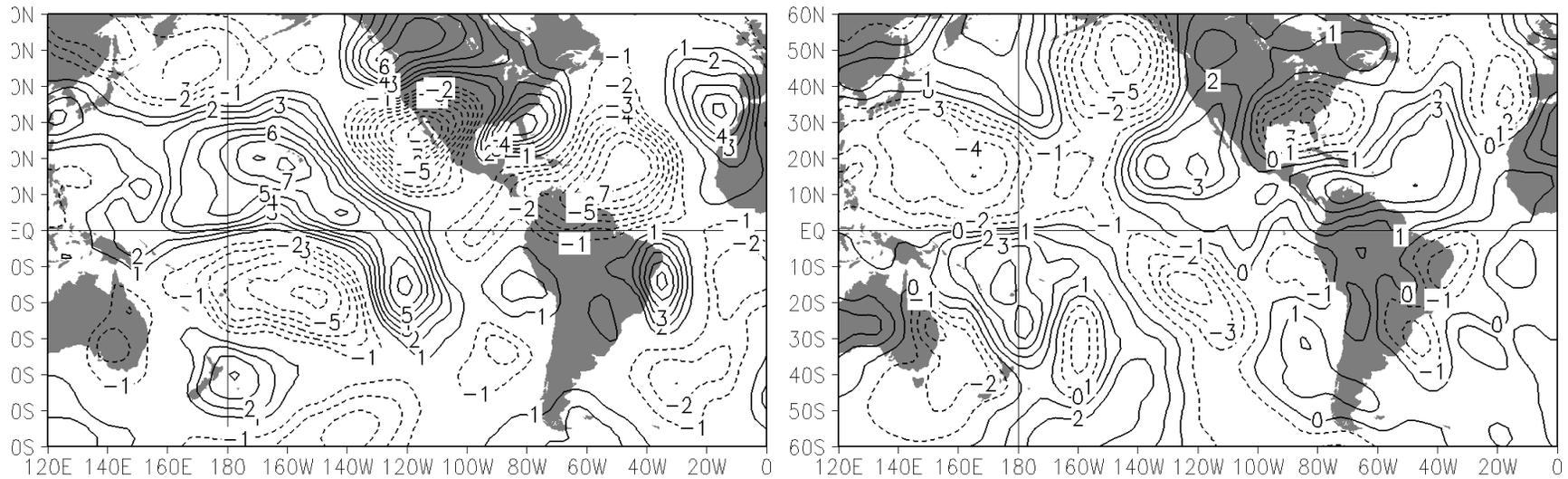


El-Nino (warm SST) events
(avg of 1983, 1998)

La-Nina (cold SST) events
(avg of 1989, 1999)

El-Nino (warm SST) events
(avg of 1983, 1998)

La-Nina (cold SST) events
(avg of 1989, 1999)



Meridional wind v
Anomalies (JFM)

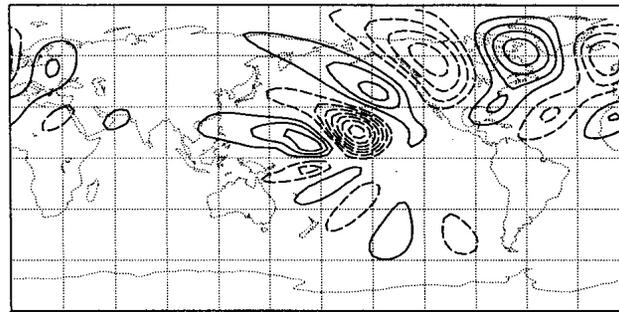


FIG. 8. Longitude–latitude picture of the day 15 $\sigma \approx 0.24$, meridional wind perturbation for the heating on a Dec–Feb zonal flow. The contour interval is 0.5 m s^{-1} . The zero contour is not shown, and the negative contours are dashed.

But chaos does cause problems here too!

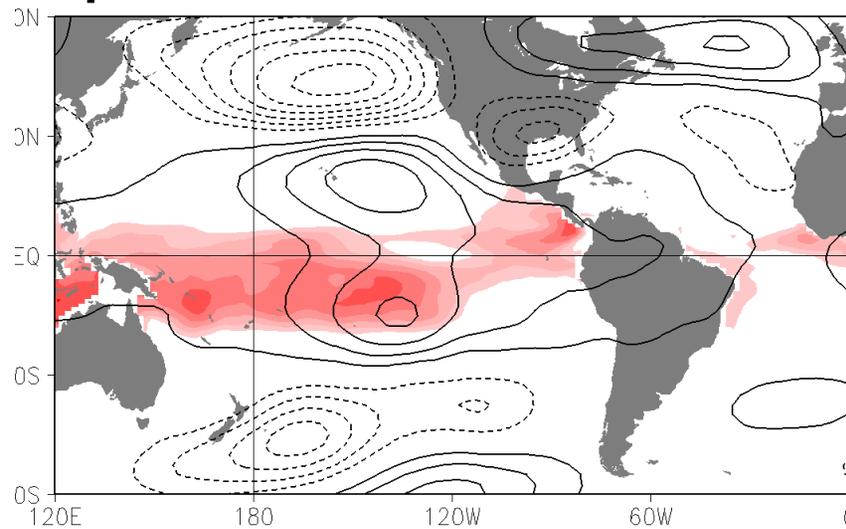
Even in the presence of a strong boundary forcing (SST), predicting the seasonal mean from an initial value problem can be **very difficult due to non-linearity**

Essentially, many atmospheric simulations using the same SST forcing, but slightly different initial conditions may have very different seasonal means!

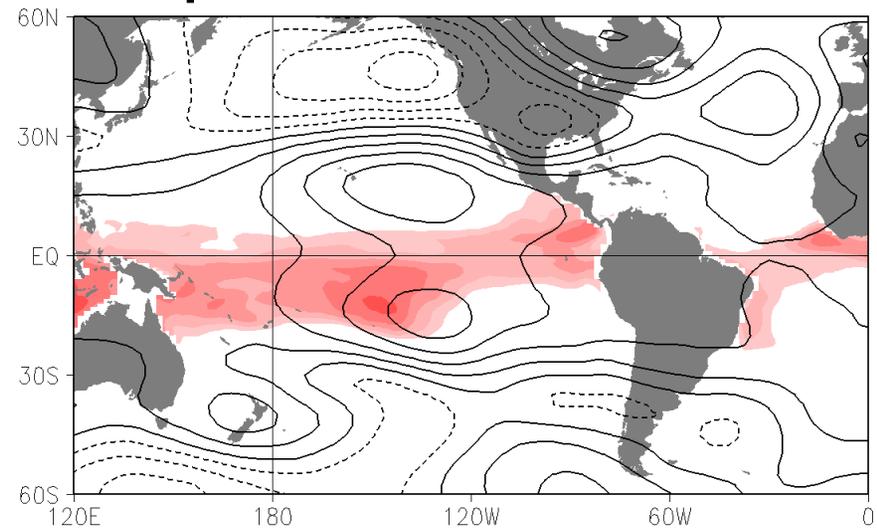
So how can we trust one forecast for the seasonal mean, even in we know the SST conditions?

Different observed warm event SSTs have lead to different observed responses

Response to 1982/83 Warm Event



Response to 1997/98 Warm Event



Contours give the 200 hPa height response (Cl = 30 m)

Shading gives the SST in degrees C

-What part of the difference between two observed events is due to the (modest) difference in SSTs during the two winters?

- What part is due to chaos?

The Ensemble Approach.

For each winter, the **ensemble** consists of a number of simulations made with the same SST but slightly different atmospheric initial conditions

The assumption is that the average seasonal mean of all simulations run for the same winter (same SST) gives a reasonable approximation to the true “forced” response of the atmosphere to that SST

(Note that we cannot observe this theoretical forced response)

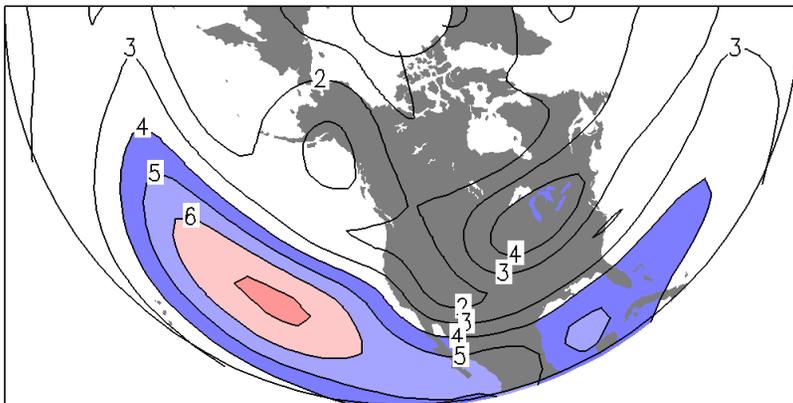
Signal variance:

Variance of 18 winter time series of ensemble mean

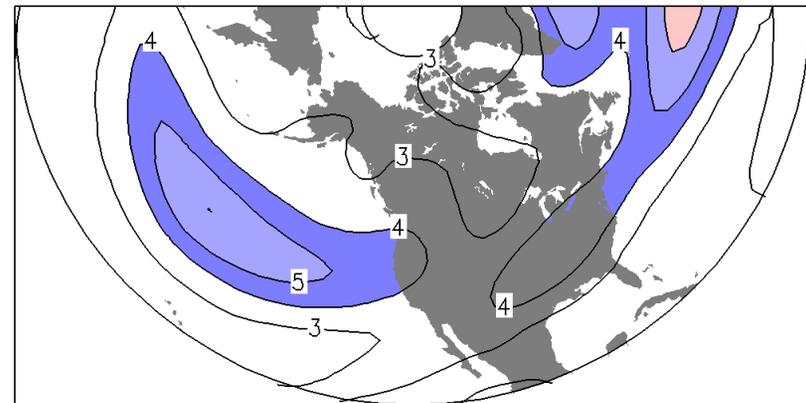
Noise variance:

Variance of each ensemble member about the ensemble mean

200 hPa Seasonal Mean Zonal Wind (jet) variance (units of m^2/s^2)



Signal Variance



Noise Variance

Analysis of Ensemble Simulations

Isolating a few critical patterns in mid-latitudes forced by tropical SST.

The idea is to expand a sequence of (seasonal mean winter) fields using a set of orthogonal patterns. Mathematically this is simply a change of coordinate systems, in which our new coordinates are the coefficients of the patterns. But if the patterns are chosen well, the first few leading patterns (variables) will give us a lot of useful information.

Method A: Principal Component Analysis (EOF analysis)

Here we first take the ensemble mean of (seasonal mean) fields for each year, that is we average over all simulations made with the same SST. If we remove the N-year mean from this sequence of fields, and call the ensemble mean anomaly for grid point i and time t $Z_{i,t}$, the expansion we seek is:

$$Z_{i,t} = \sum_{\alpha} p_t^{(\alpha)} E_i^{(\alpha)}$$

where α is the mode, or pattern index. Note that all the time information is in the components p_t , which are our new coordinates.

In order for the patterns E_i to form an orthogonal set (in the usual vector sense), we must have:

$$\sum_i E_i^{(\alpha)} E_i^{(\beta)} = \delta_{\alpha,\beta}$$

where $\delta_{\alpha,\beta} = 1$ if $\alpha = \beta$ and 0 otherwise.

The new coordinates p_t are also orthogonal in time:

$$\frac{1}{N} \sum_t p_t^{(\alpha)} p_t^{(\beta)} = \delta_{\alpha,\beta} \lambda^{(\alpha)}$$

where λ is an eigenvalue of the covariance matrix $C_{i,j} = \frac{1}{N} \sum_t Z_{i,t} Z_{j,t}$

The corresponding eigenvector is E_i , also known as an Empirical Orthogonal Function.

$$\sum_j C_{i,j} E_j^{(\alpha)} = \lambda^{(\alpha)} E_i^{(\alpha)}$$

The coordinates p_t are called the principal components (PCs).

What makes this expansion very efficient in representing patterns is the property that each mode (α) “explains” a fraction of the total variance V :

$$V = \sum_i \left(\frac{1}{N} \sum_t Z_{i,t}^2 \right) = \sum_{\alpha} \lambda^{(\alpha)}$$

If the (positive definite) eigenvalues λ are arranged in descending order, then the leading mode explains the largest percentage of the variance, the second mode the largest percentage of the variance of any mode orthogonal to the first, and so on. Another way of putting this is that if you synthesize the field with only a few leading modes (say = 1-3 only), the fraction of the

$$Z_{i,t} = \sum_{\alpha=1}^3 p_t^{(\alpha)} E_i^{(\alpha)}$$

the total variance V explained by only these three modes is: $\frac{\lambda^{(1)} + \lambda^{(2)} + \lambda^{(3)}}{\sum_{(\alpha)} \lambda^{(\alpha)}}$

Thus if a few eigenvalues dominate, then only a few modes are needed to explain most of the variance V .

Method B. Optimal Signal-to-Noise Patterns

This approach tries to use *all* the ensemble members (labelled by **m**) for each winter (again labelled by **i**) to determine an SST forced **signal** and the internal dynamics generated **noise**.

We proceed by first subtracting the climate mean, as before, and expand each seasonal mean field in terms of a set of patterns again:

$$Z_{i,t,m} = \sum_{\alpha} p_{t,m}^{(\alpha)} E_i^{(\alpha)}$$

Note that all seasonal means have been used, and that the ensemble information is carried in the coefficients. We also form the ensemble mean coefficients:

$$\hat{p}_t^{(\alpha)} = \frac{1}{M} \sum_m p_t^{(\alpha)}$$

where M is the ensemble size. Then the signal S and noise N are defined *for each individual mode* α as follows:

$$S = \frac{1}{N} \sum_t \left(\hat{p}_t^{(\alpha)} \right)^2$$

$$N = \frac{1}{N} \sum_t \left(\frac{1}{M} \sum_m \left(p_{t,m}^{(\alpha)} - \hat{p}_t^{(\alpha)} \right)^2 \right)$$

The patterns \mathbf{E}_i and coefficients $p_{t,m}$ are determined (via a generalized eigenvalue problem) by the requirement that:

- The leading mode maximizes the signal-to-noise ratio S/N
- The second leading mode maximizes S/N subject to the constraint that the associated pattern $\mathbf{E}_i^{(2)}$ is orthogonal to $\mathbf{E}_i^{(1)}$.
- ...

The patterns in Principal Component Analysis are orthogonal, while those in the Optimal Signal-to-Noise expansion are bi-orthogonal. **In both cases they form a complete set that can be used to expand any set of fields.**

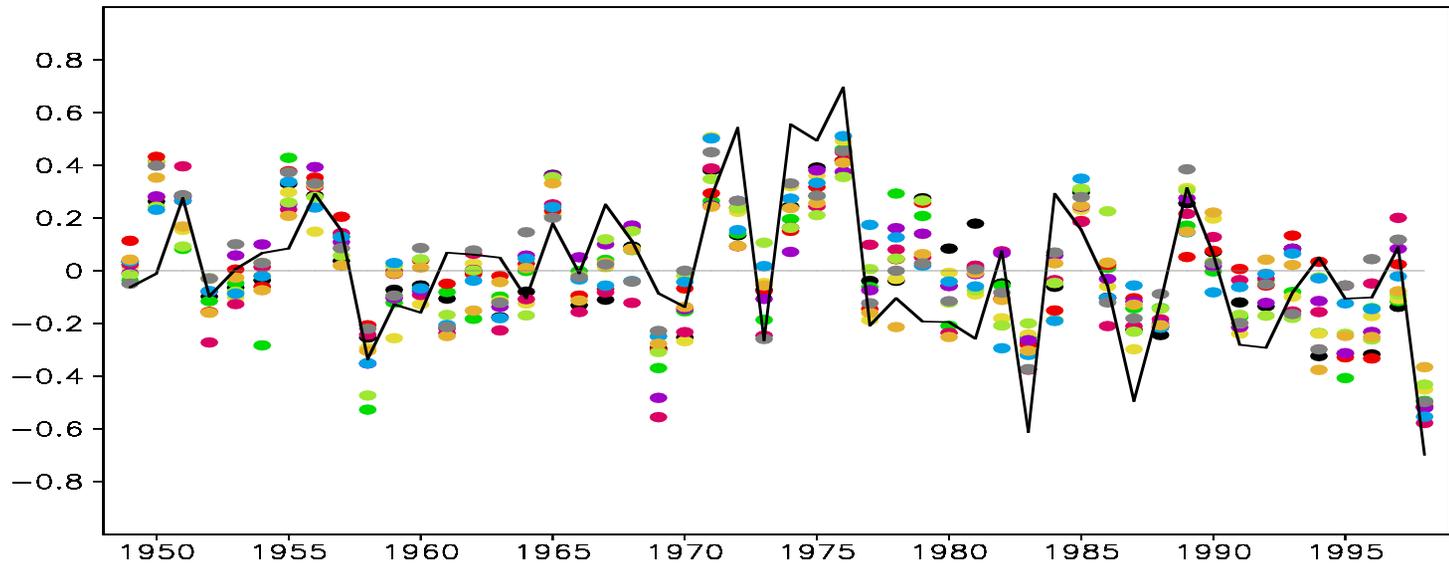
Application to C20C Experiments with COLA AGCM

10 member ensembles, 50 winters used

Each colored dot represents one of the coefficients $p_{t,m}$ for the leading Optimal mode.

The solid line is the (single) NCEP reanalysis winter seasonal mean projected onto the pattern of the leading mode

GCM Optimal PC (circle) NCEP(line)

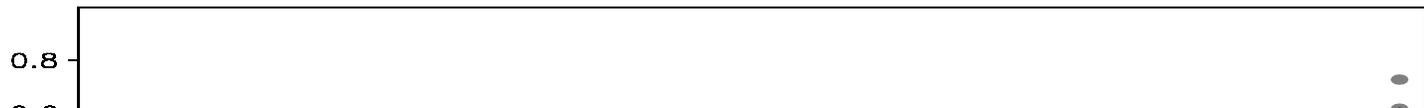


all 50 winters together

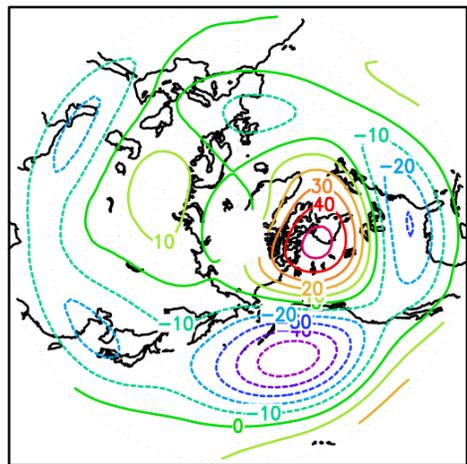
ICTP Climate Predictability II

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GCM Optimal PC (circle) NCEP(line)

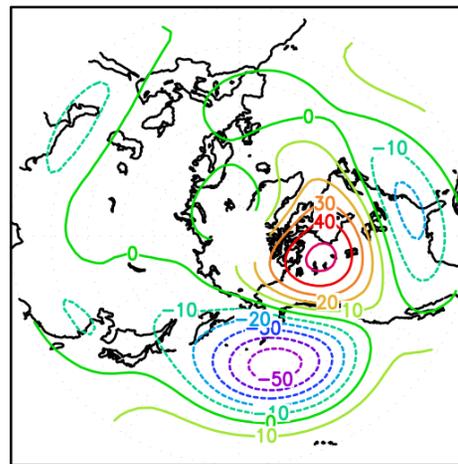


1981–1998 EEOF-1



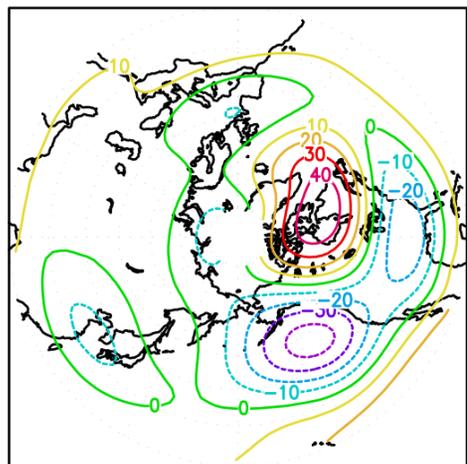
38% variance explained

1949–1980 EEOF-1



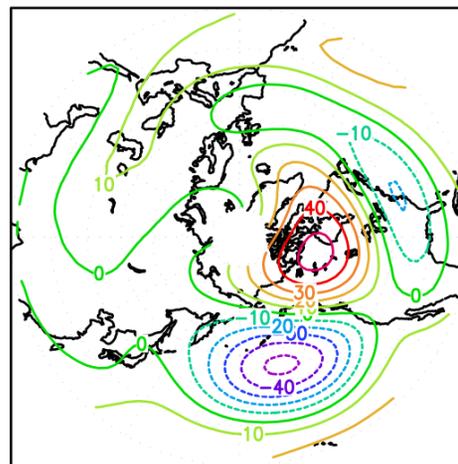
42% variance explained

1981–1998 OPT-1



Ratio = 8.2

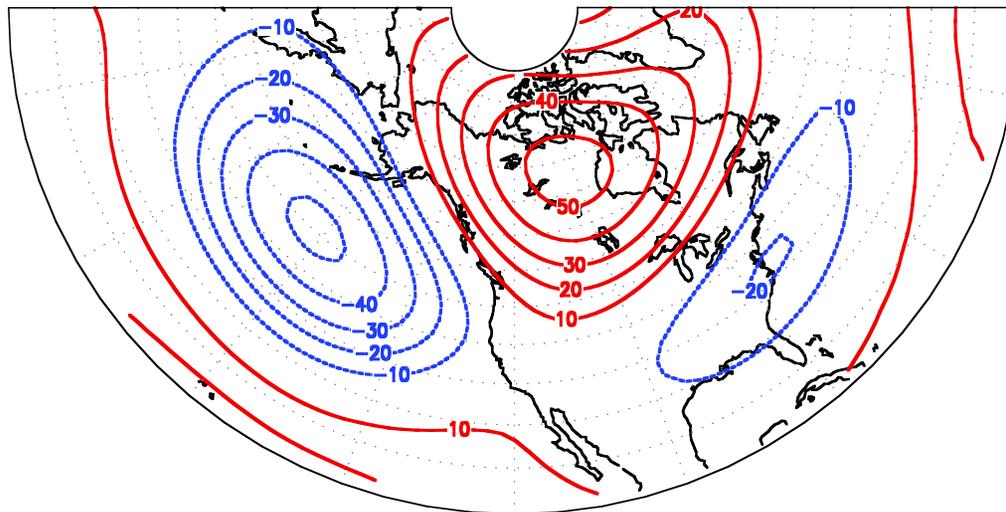
1949–1980 OPT-1



Ratio = 5.7

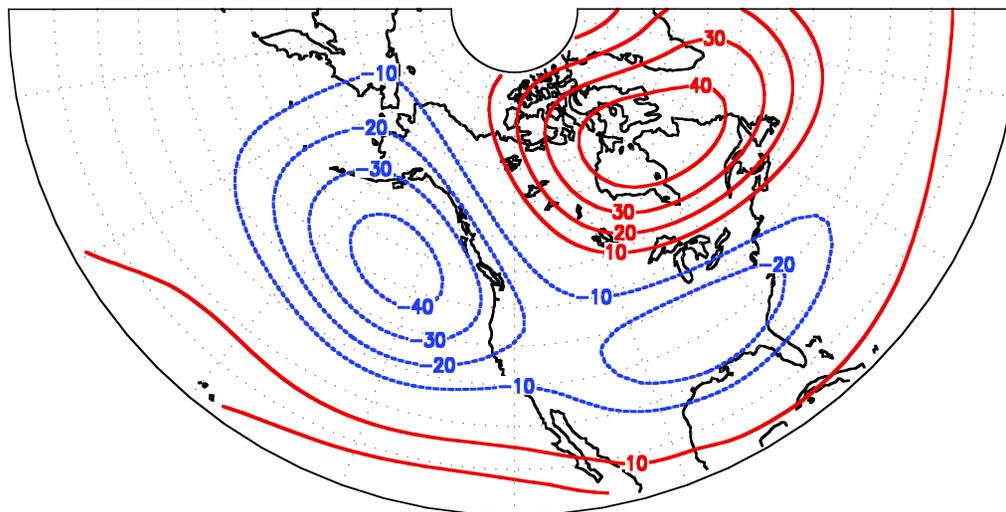
Pattern that optimizes the signal-to-noise ratio for seasonal mean 200 hPa height for winter, from observations.

(a) Early Period (1949–1980)



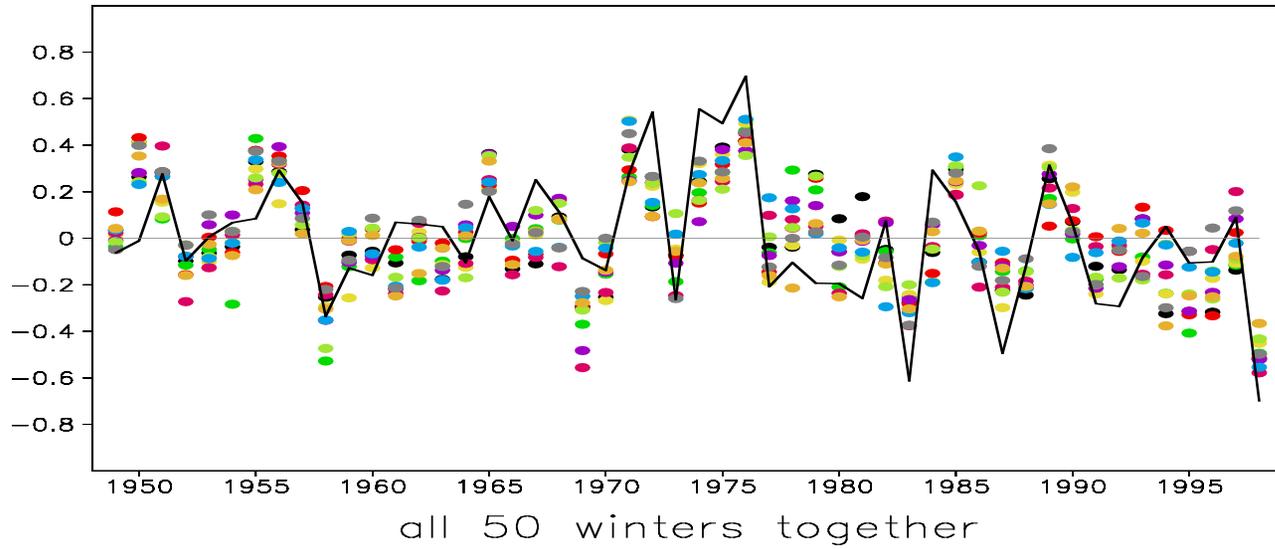
JFM Geo 200 Optimal Pattern (m)

(b) Later Period (1981–1998)



JFM Geo 200 Optimal Pattern (m)

GCM Optimal PC (circle) NCEP(line)



GCM Optimal PC (circle) NCEP(line)

