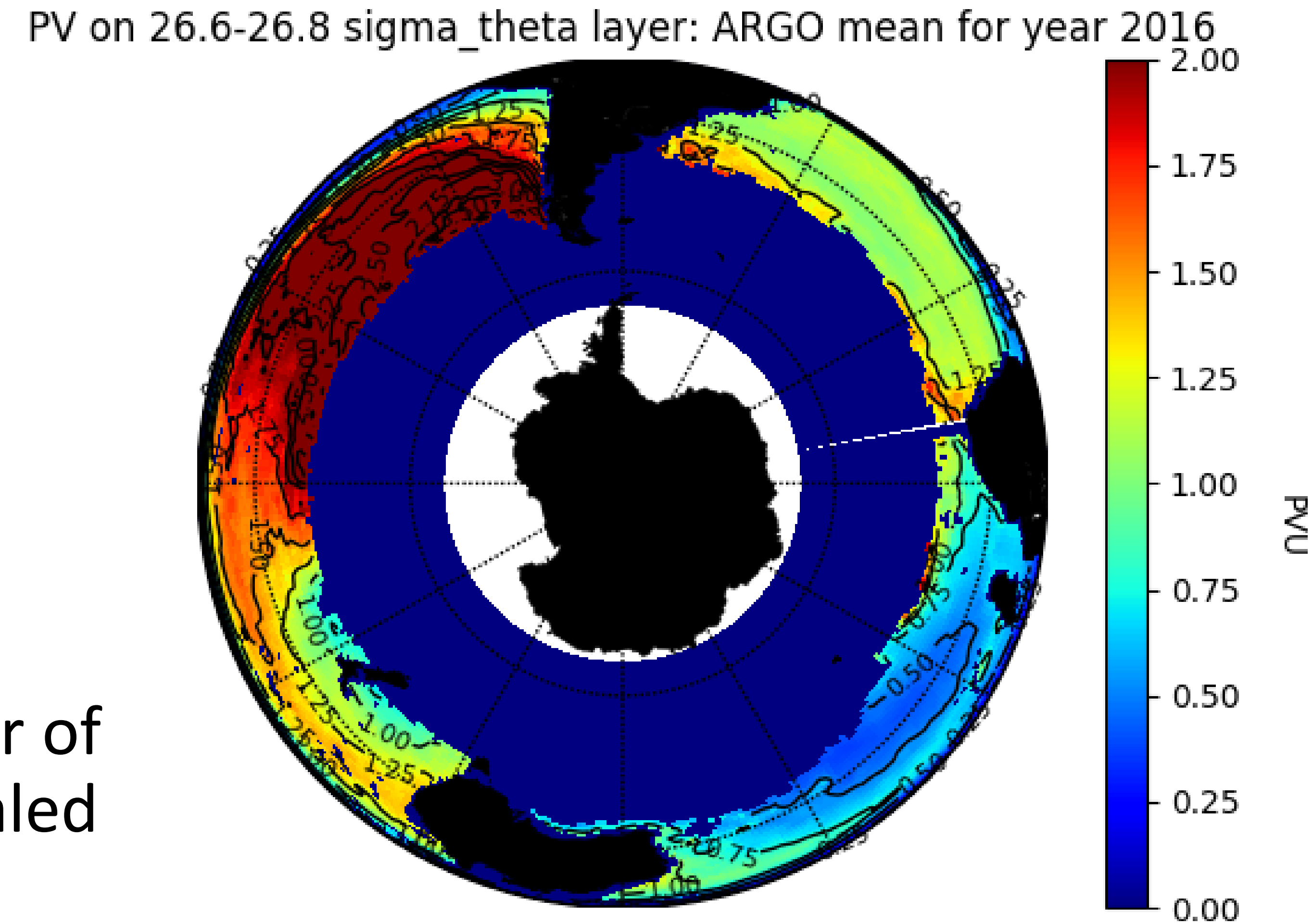
Ocean-atmosphere coupling: theory and “observations”

Arnaud Czaja

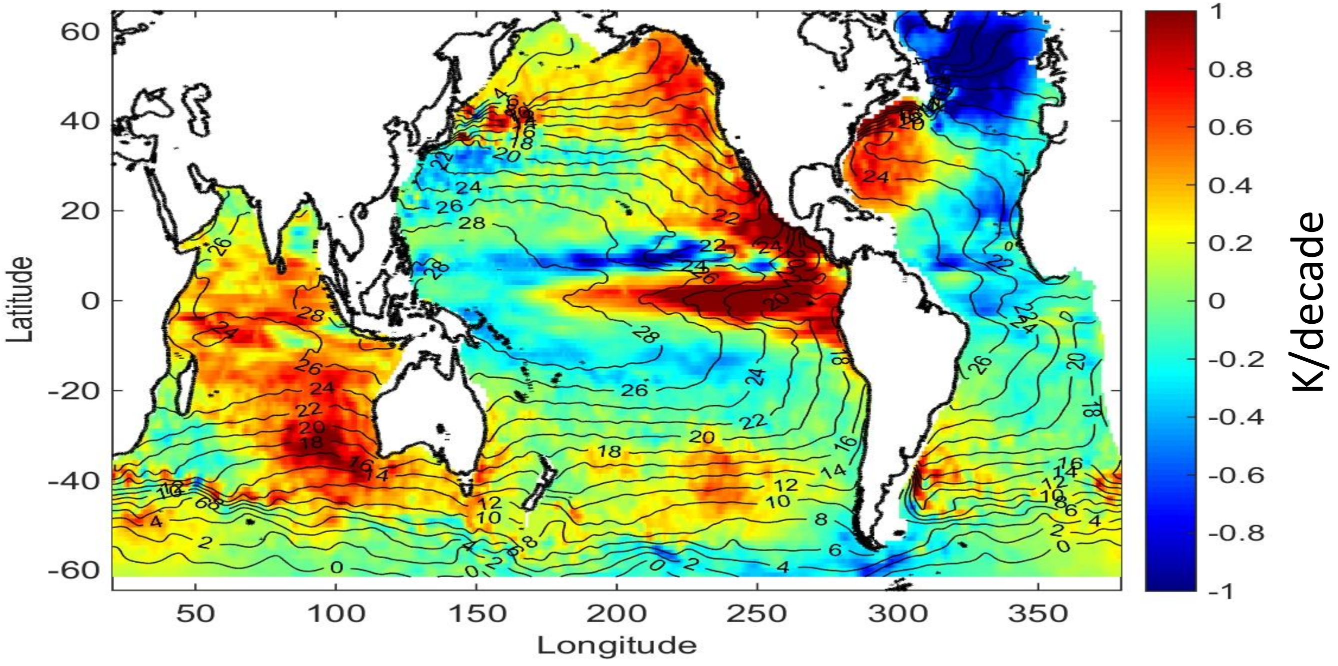
Imperial College London



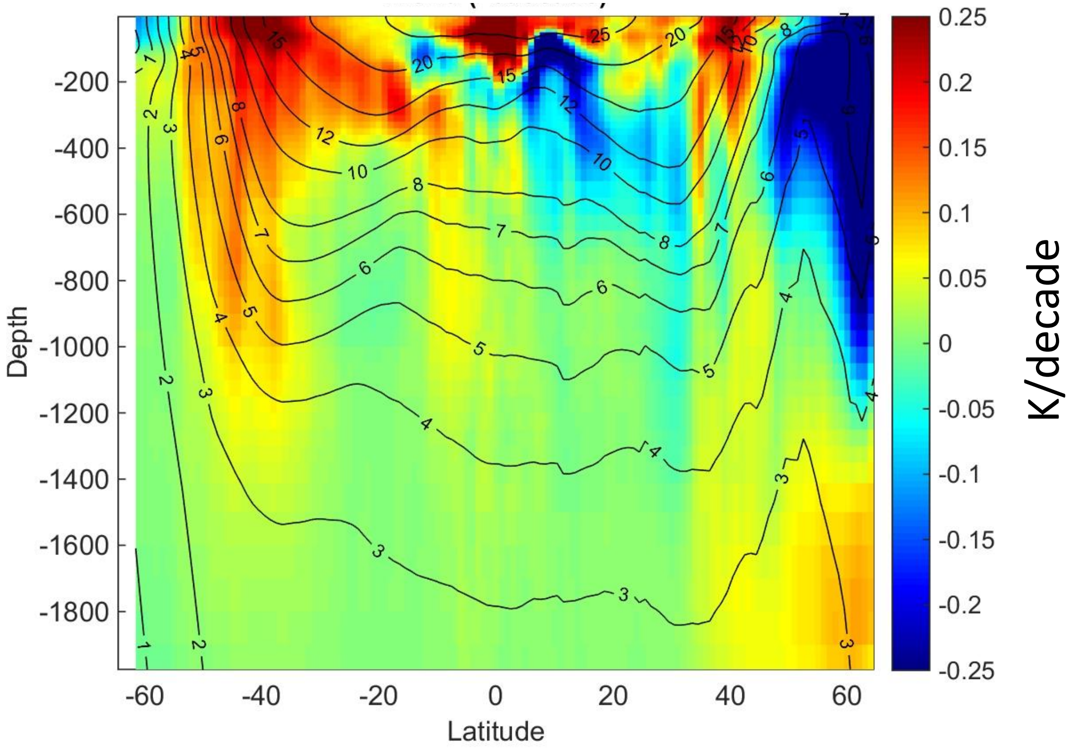
- The rich dynamical behaviour of the ocean begins to be revealed by global observations

Observed changes in ocean heat content from Argo floats (2004-2015)

Upper ocean (0-100m) heat content trend



Zonal mean heat content trend



Global average ~0.5 W/m² driving sea level rise ~1mm/yr…
but also changes in global weather patterns?

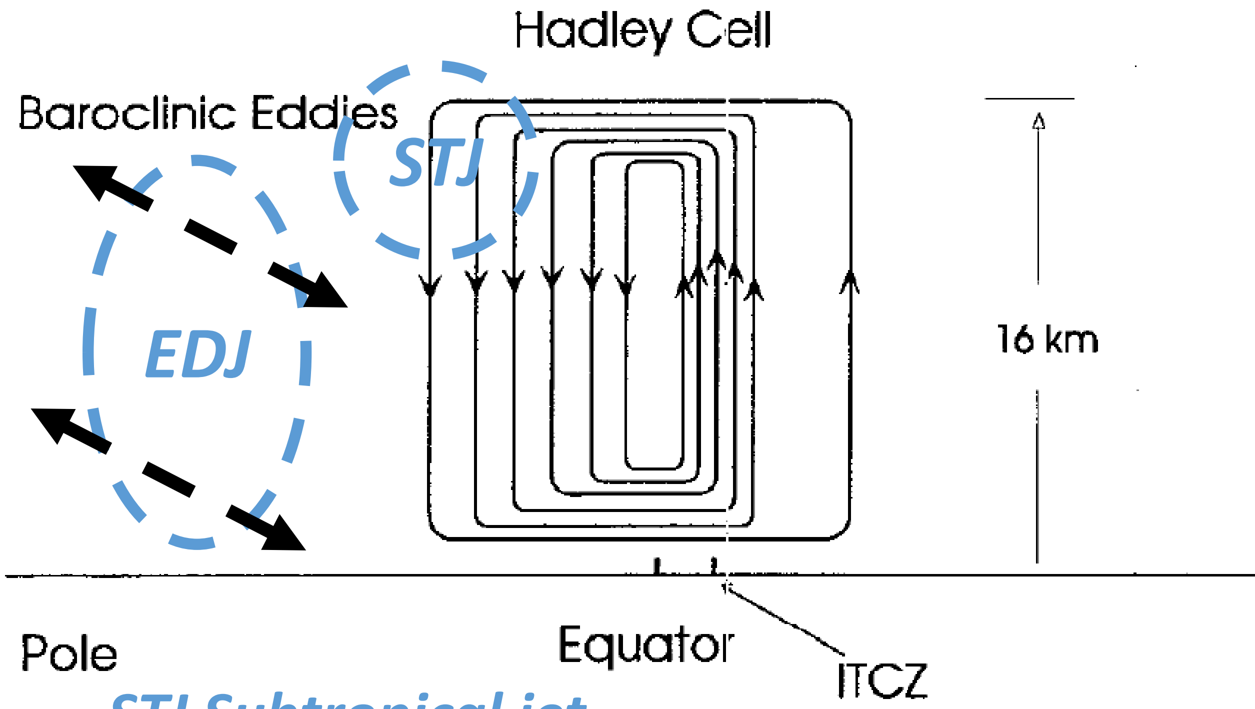
Key questions of this lecture:

- Are changes in upper ocean temperature only affecting the surface of the atmosphere (~1km)…?
- …or do they extend upward beyond the first km or so?
- If so, how does it work?

Key question of this lecture:

- Are changes in upper ocean temperature only affecting the surface of the atmosphere (~1km)…?
- …or do they extend upward beyond the first km or so?
- If so, how does it work?

Winter Hemisphere



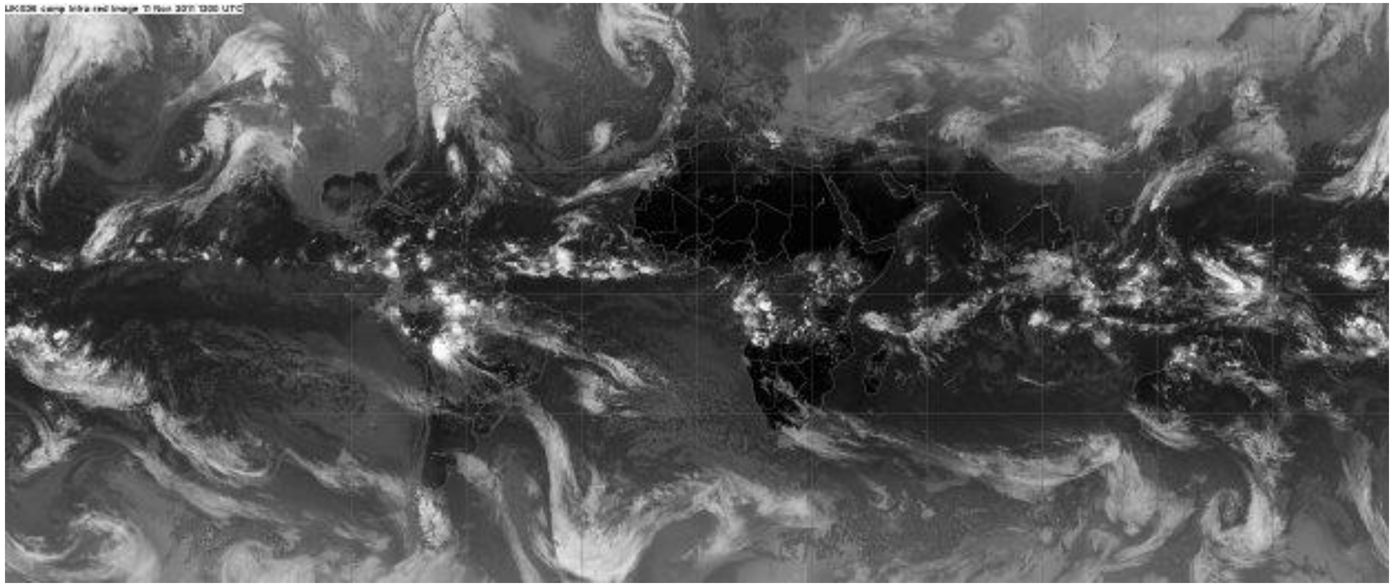
STJ Subtropical jet
EDJ Eddy-driven jet

Picture adapted from Lindzen (1994)

Outline

- An overview of observed "coupled variability" between oceans and atmosphere

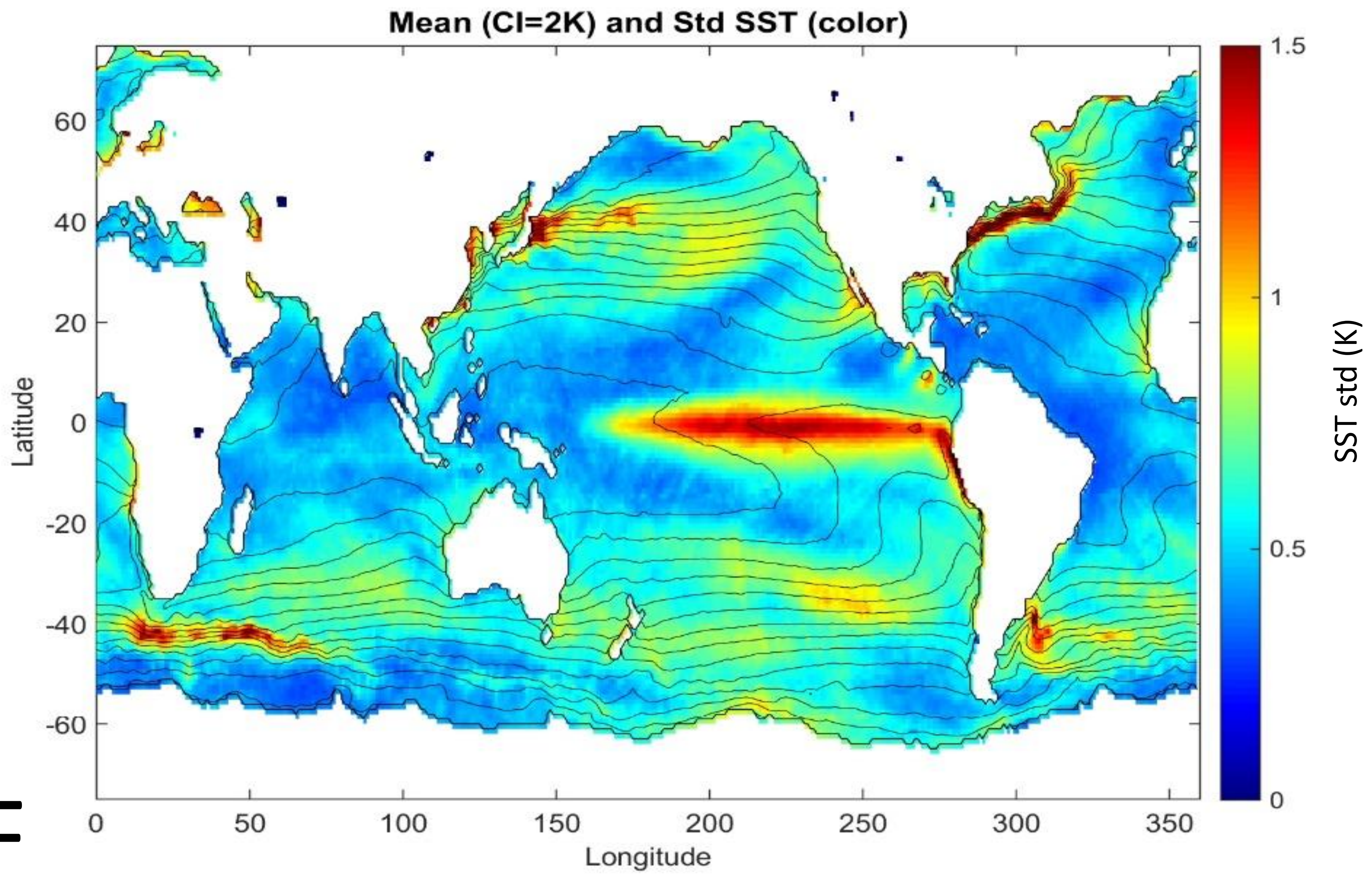
- key issue: How do SST patterns affect the atmosphere above the boundary layer?



Infrared snapshot: white=cold=cloud top

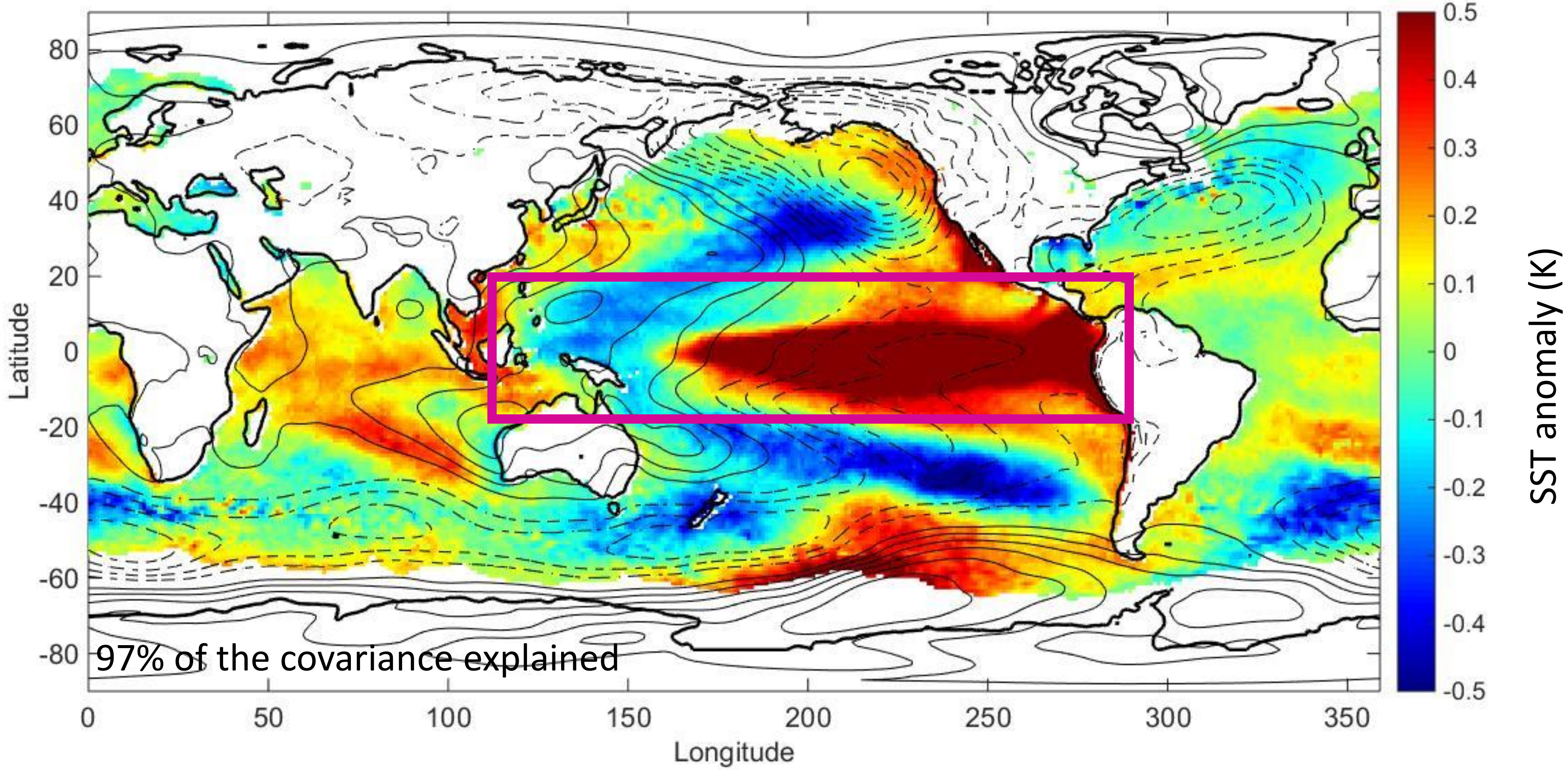
1. An overview of “coupled” ocean-atmosphere variability

- Use de-seasonalised monthly anomalies in sea surface temperature (SST) and sea level pressure (SLP) to compare the observed variability in the North Atlantic, North Pacific, Equatorial Pacific and the Southern Ocean
- Apply a maximum covariance analysis (MCA) to SST and SLP from the ERA20C (1960-2010) reanalysis to do so
- Discuss predictability of ENSO and NAO


Mean (CI=2K) and Std SST (color)
Latitude
Longitude
SST std (K)
60
40
20
0
-20
-40
-60
0
50
100
150
200
250
300
350
1.5
1
0.5
0

DJF

SST (color) / MSLP (ci=0.25hPa) Equatorial Pacific
(Max Cov. Ana., ERA20C, D-J-F 1960-2010)



NB Analysis extended globally via linear regression

SST / MSLP statistics for the Equatorial Pacific
(Max Cov. Ana., ERA20C, D-J-F 1960-2010)

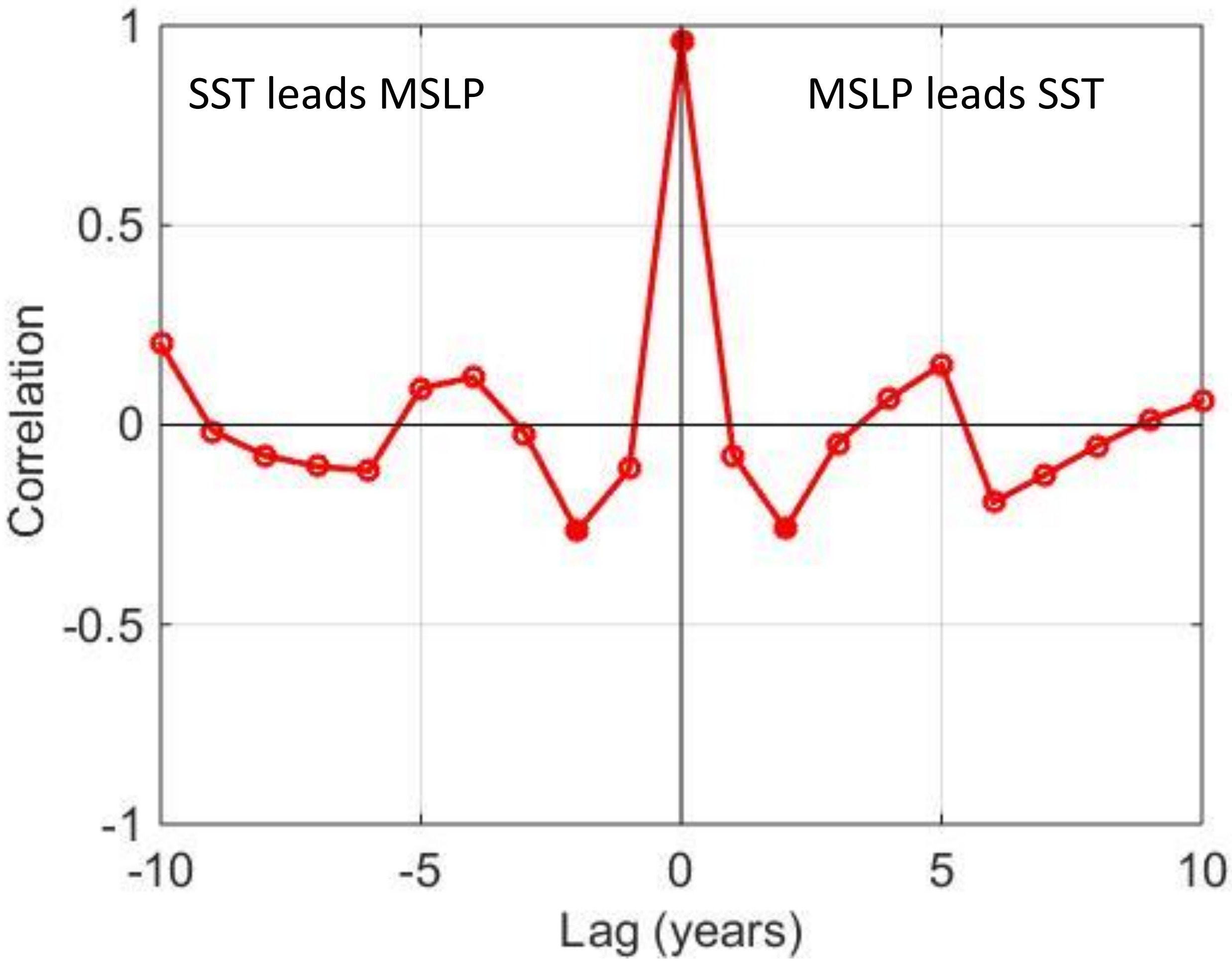
- Monthly statistics:

SST 1-month autocor = 0.97

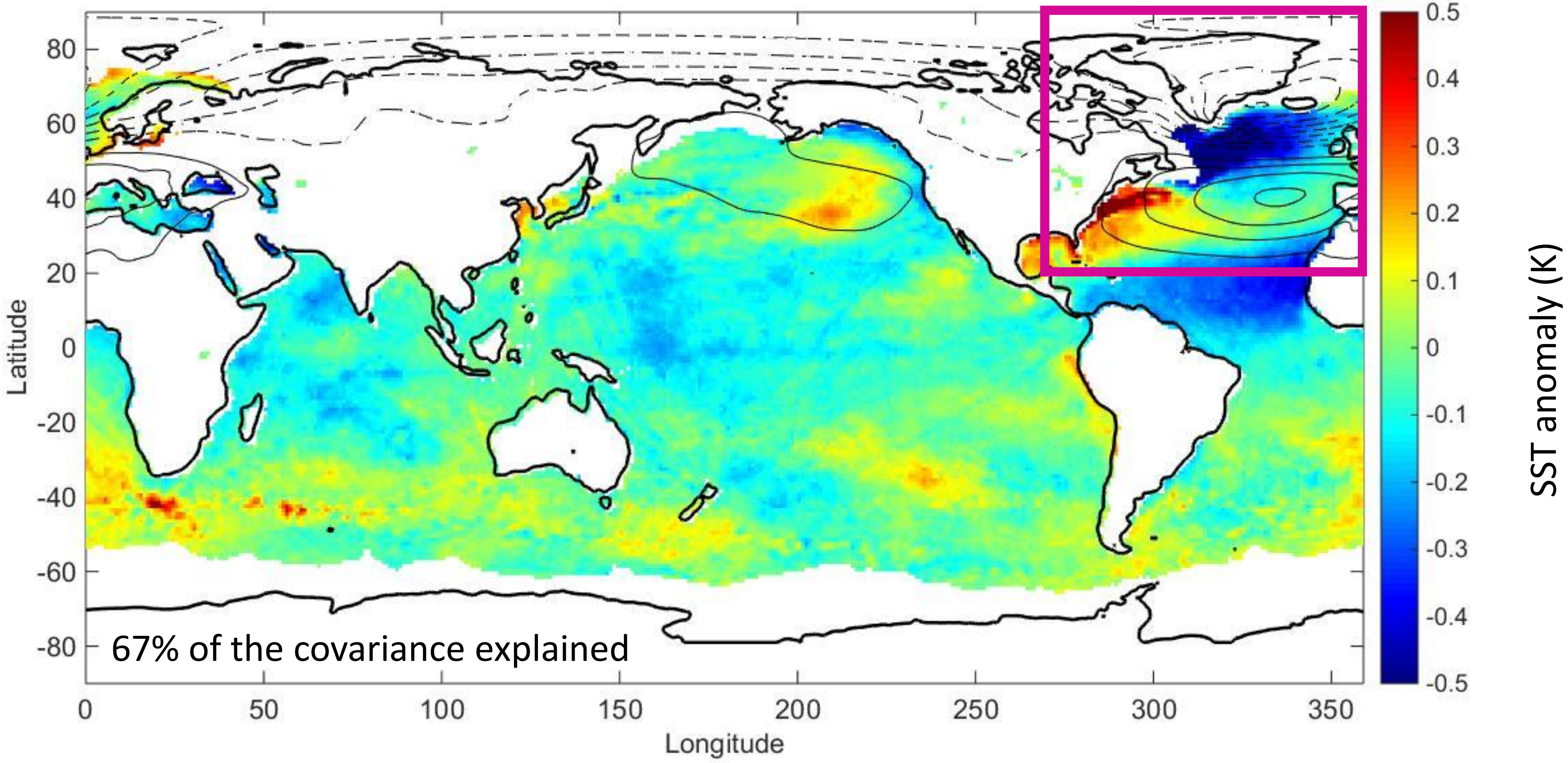
MSLP 1-month autocor = 0.84

Cross-corr: 0.89, 0.9, 0.89

- Longer timescale statistics: strong covariability, clear evidence of interannual oscillation



SST (color) / MSLP (ci=1hPa) North Atlantic
(Max Cov. Ana., ERA20C, D-J-F 1960-2010)



NB Analysis extended globally via linear regression

SST / MSLP statistics for the North Atlantic
(Max Cov. Ana., ERA20C, D-J-F 1960-2010)

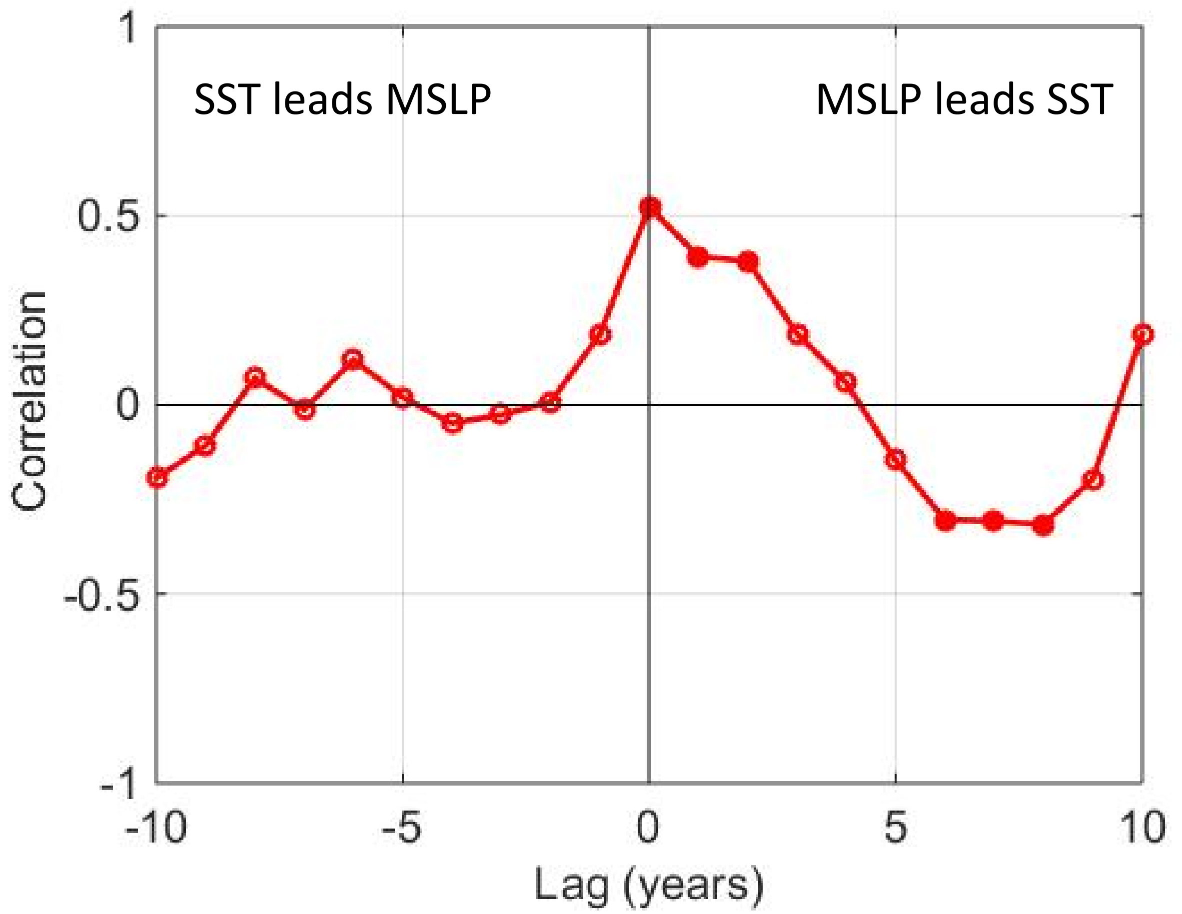
- Monthly statistics:

SST 1-month autocor = 0.86

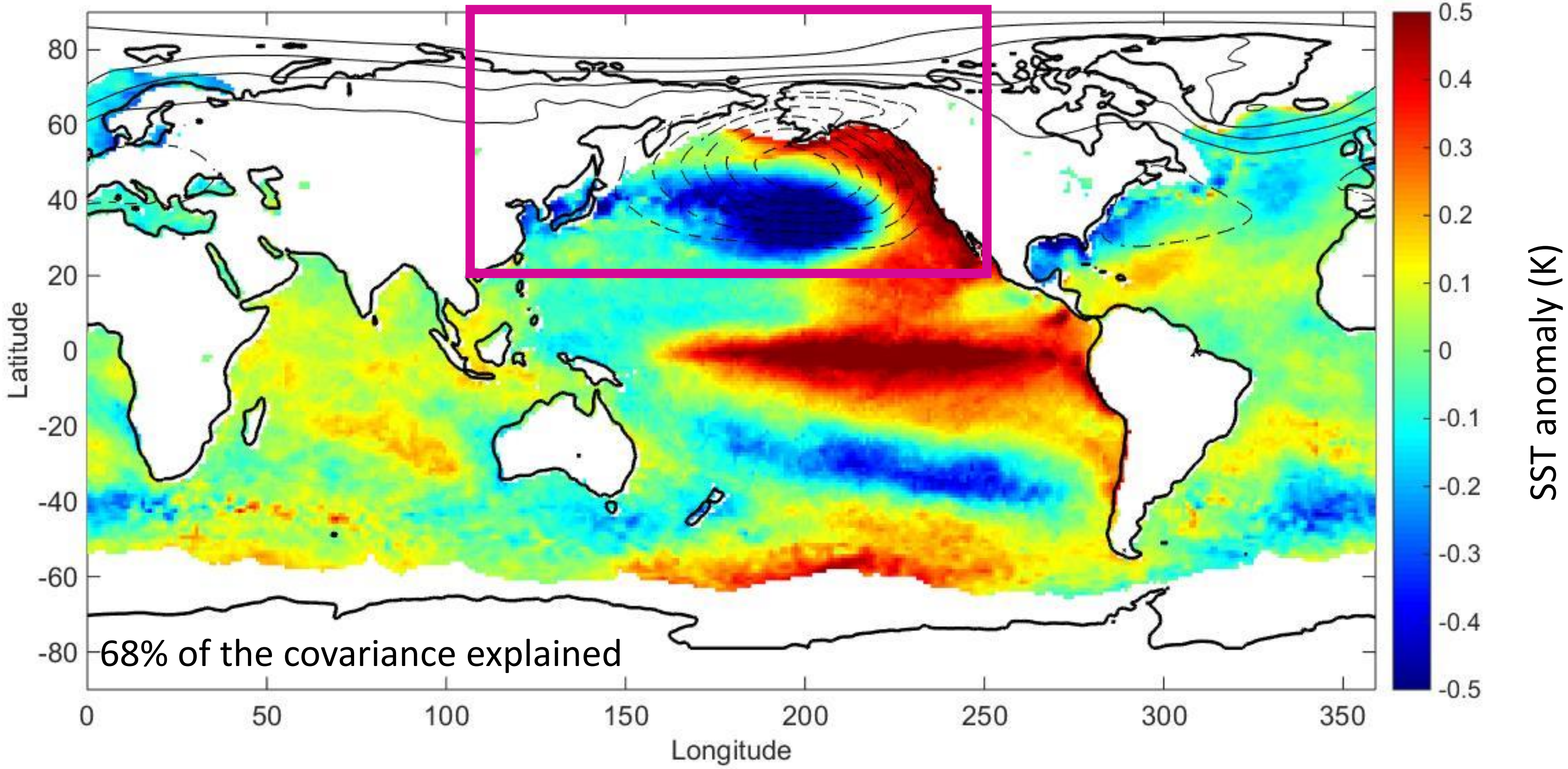
MSLP 1-month autocor = 0.24

Cross-corr: 0.15, 0.47, 0.53

- Longer timescale statistics: indication of decadal oscillation, with SST pattern reversing sign 6-8 years after MSLP



SST (color) / MSLP (ci=1hPa) North Pacific
(Max Cov. Ana., ERA20C, D-J-F 1960-2010)



NB Analysis extended globally via linear regression

SST / MSLP statistics for the North Pacific
(Max Cov. Ana., ERA20C, D-J-F 1960-2010)

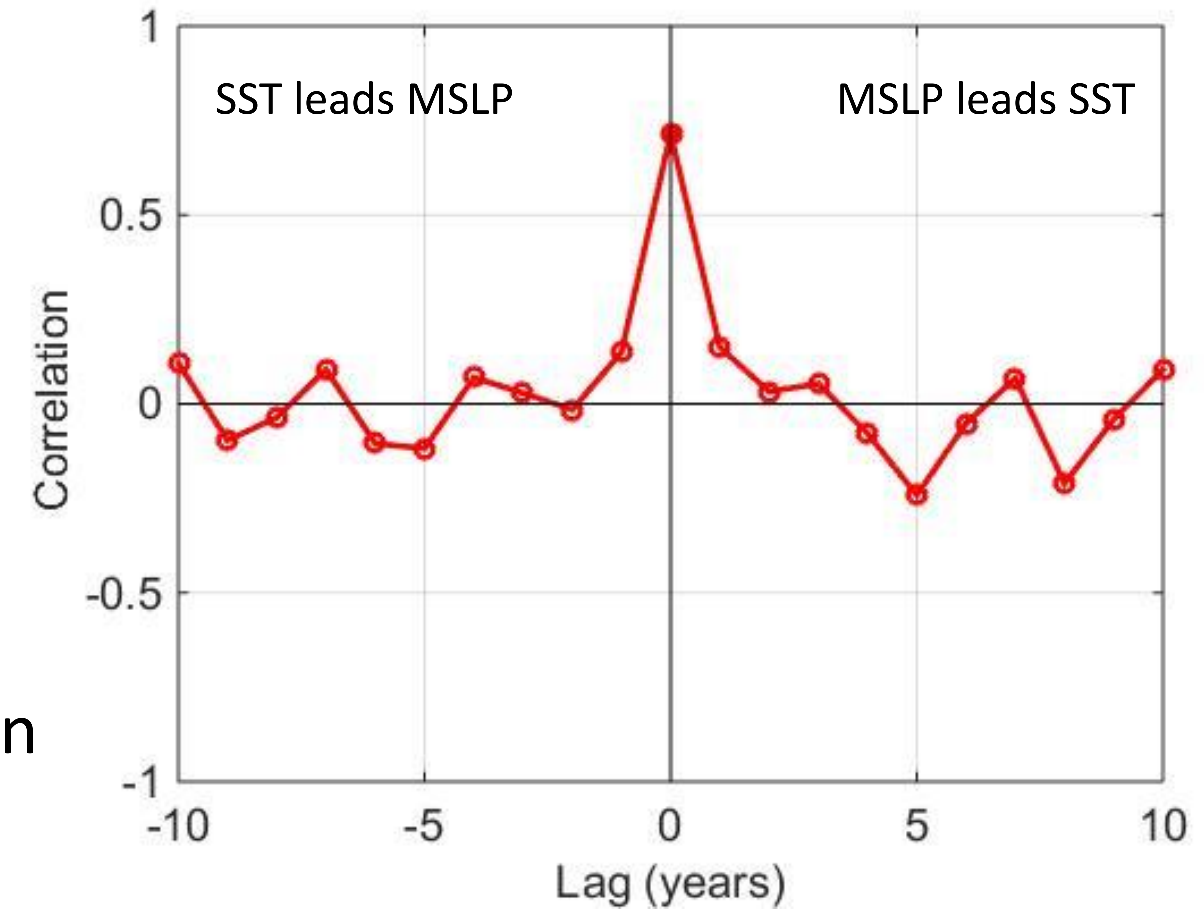
- Monthly statistics:

SST 1-month autocor = 0.92

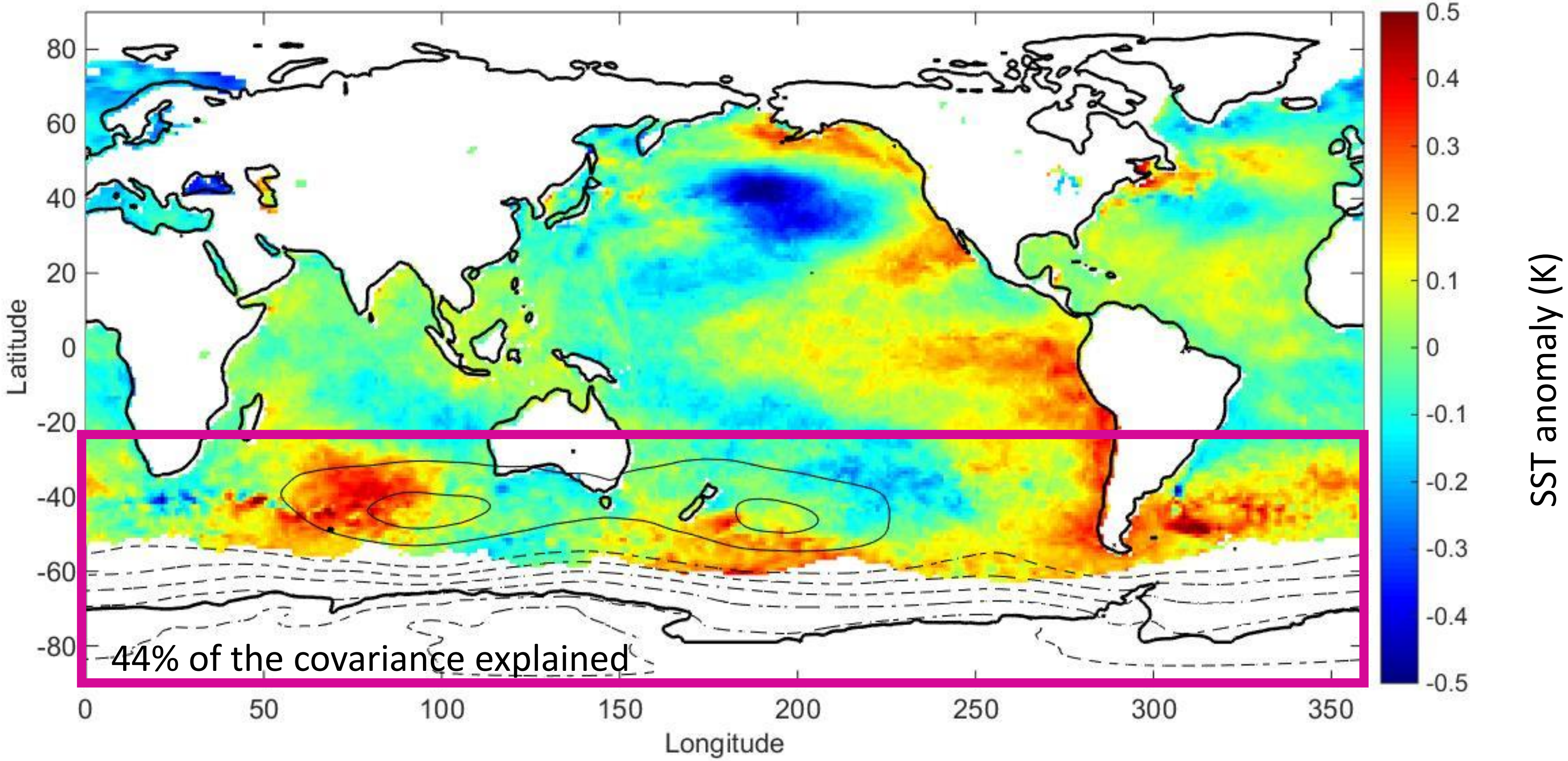
MSLP 1-month autocor = 0.41

Cross-corr: 0.44, 0.54, 0.64

- Longer timescale statistics: strong covariability; no stat. significant delay but decadal oscillation seen in each field separately (not shown) → "stationary oscillation"



SST (color) / MSLP (ci=1hPa) Southern Ocean
(Max Cov. Ana., ERA20C, J-J-A 1960-2010)



NB Analysis extended globally via linear regression

SST / MSLP statistics for the Southern Ocean
(Max Cov. Ana., ERA20C, J-J-A 1960-2010)

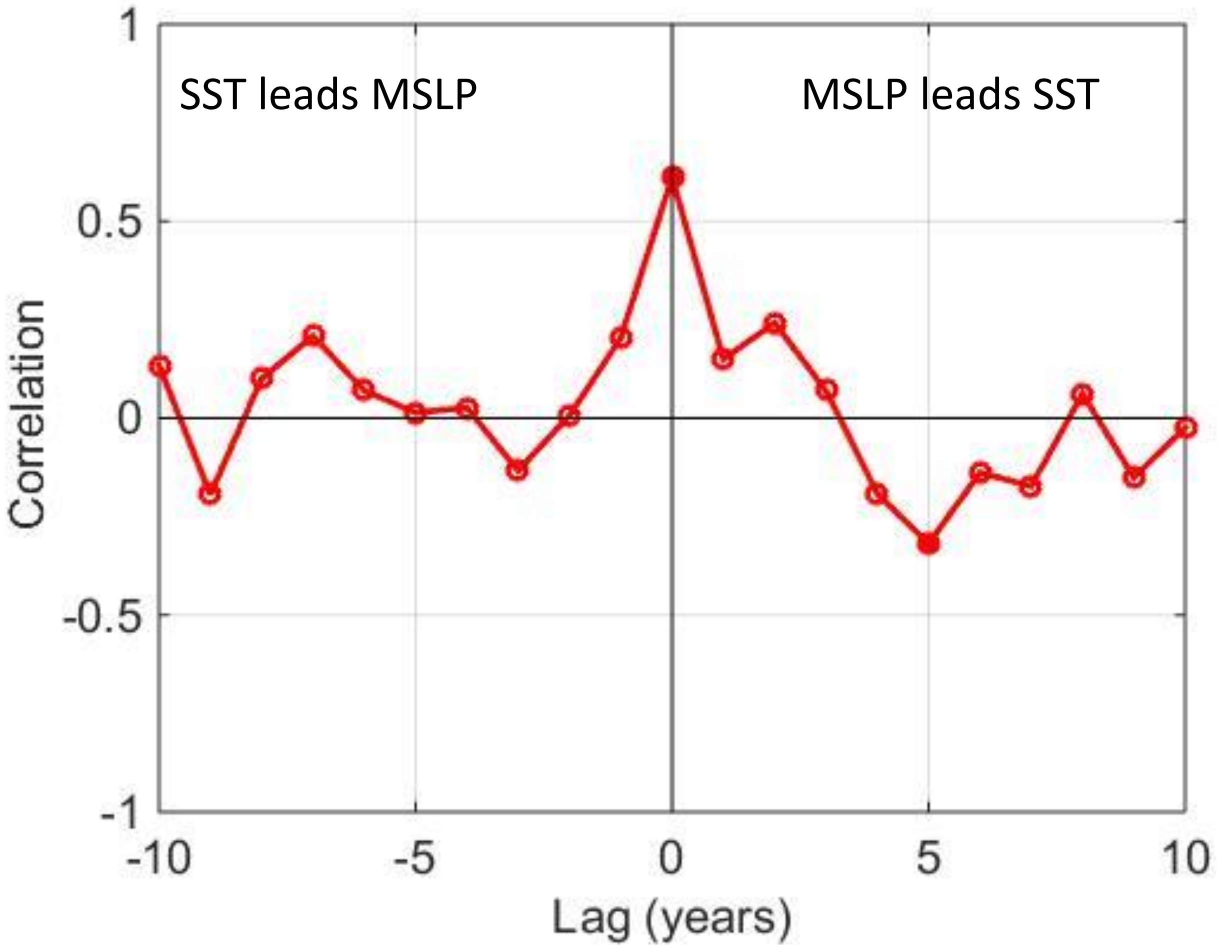
- Monthly statistics:

SST 1-month autocor = 0.89

MSLP 1-month autocor = 0.18

Cross-corr: 0.34, 0.43, 0.45

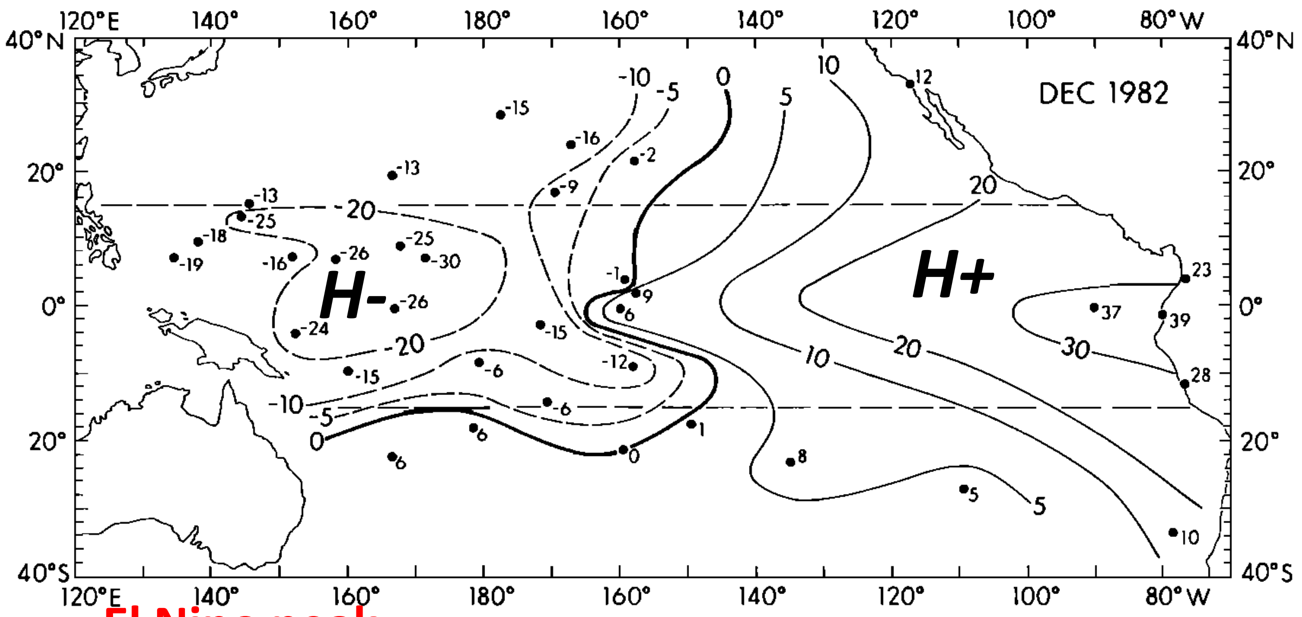
- Longer timescale statistics: indication of decadal oscillation, with a reversal of SST pattern 5 years after MSL



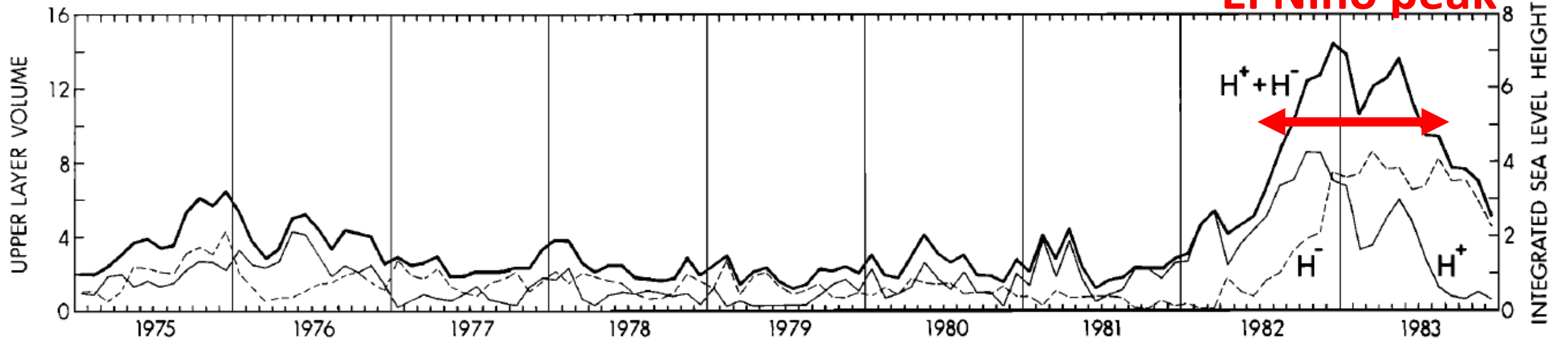
Wyrti’s (1985) view of El Nino

- El Nino = burst event to release heat stored in the equatorial Pacific

Sea level anomalies in December 1982 (cm)



El Nino peak



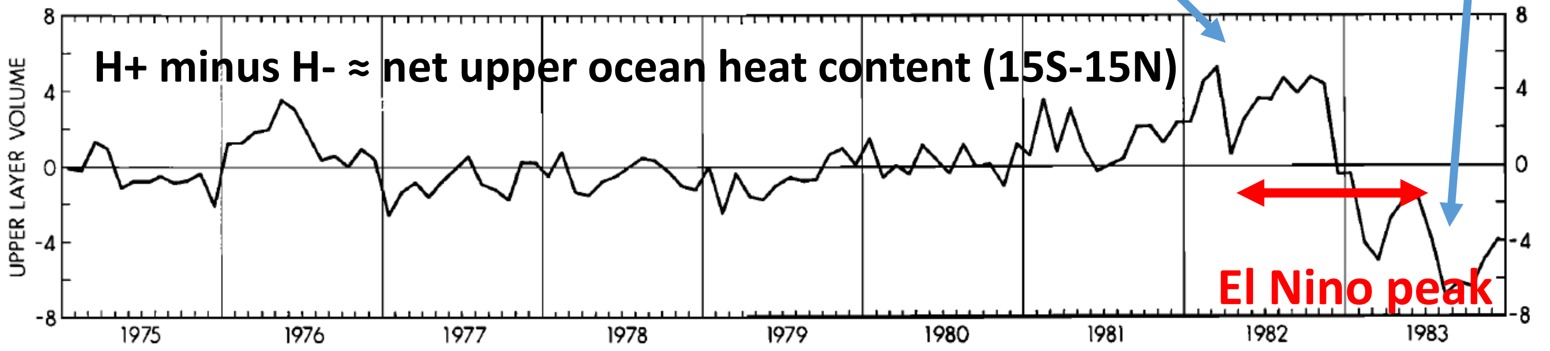
Heat builds up in the Tropical Pacific

Heat is exported away from the Tropical Pacific

NB: mean volume is ~70 in this unit

NB2: See Mayer et al. (2016) for the full energy budget view

H+ minus H- ≈ net upper ocean heat content (15S-15N)

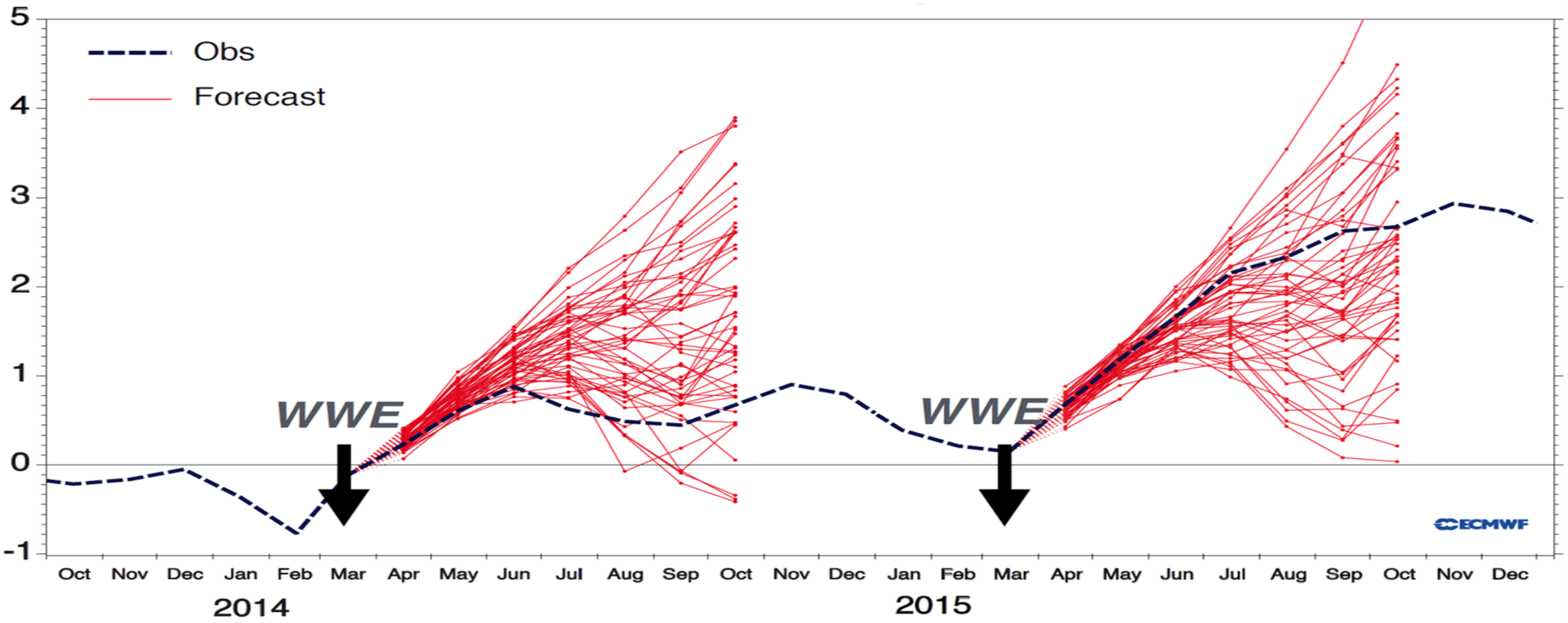


El Nino peak

Wyrkti's (1985) view of El Nino

- "The explanation of an El Nino cycle as a combination of atmospheric randomness and a deterministic ocean might be unpleasing to many scientists, but it probably characterizes correctly the interaction between the two media"

Predictive skill in the Tropics: “La Nada 2014”

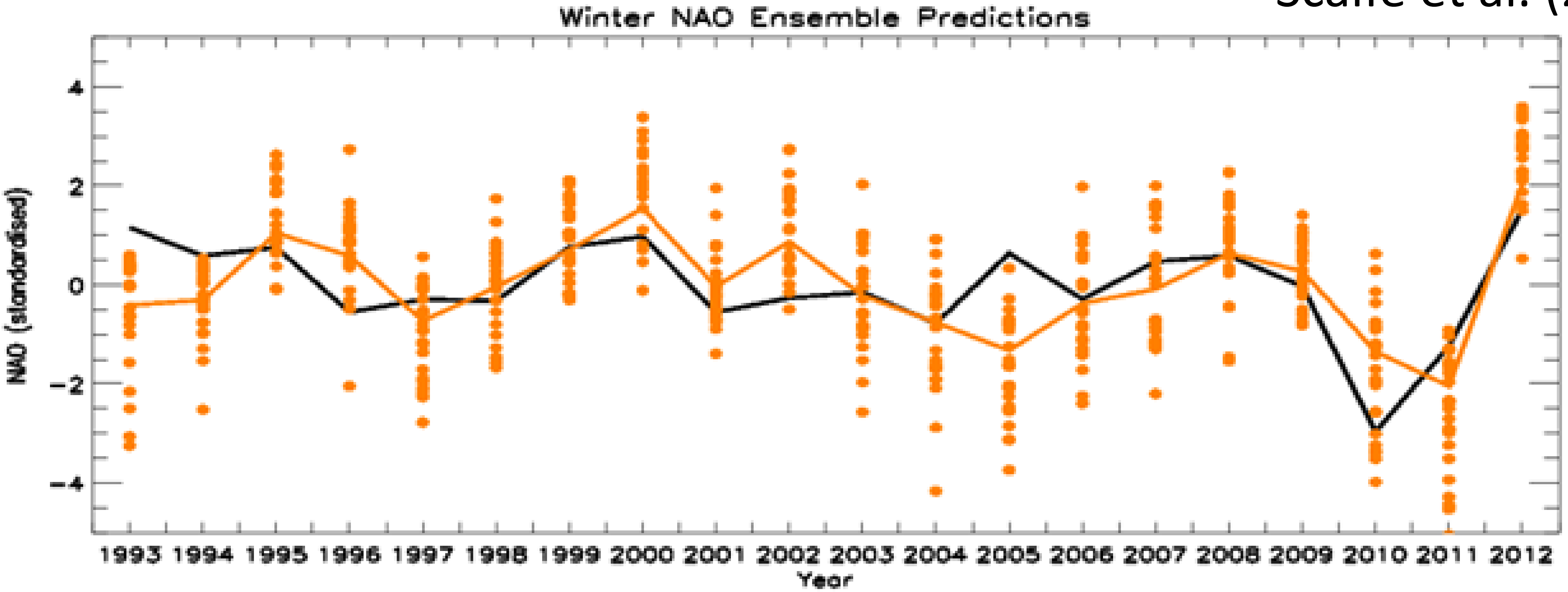


(slide courtesy of J. Vialard)

See also McPhaden (2015)

Predictive skills in the extra-tropics

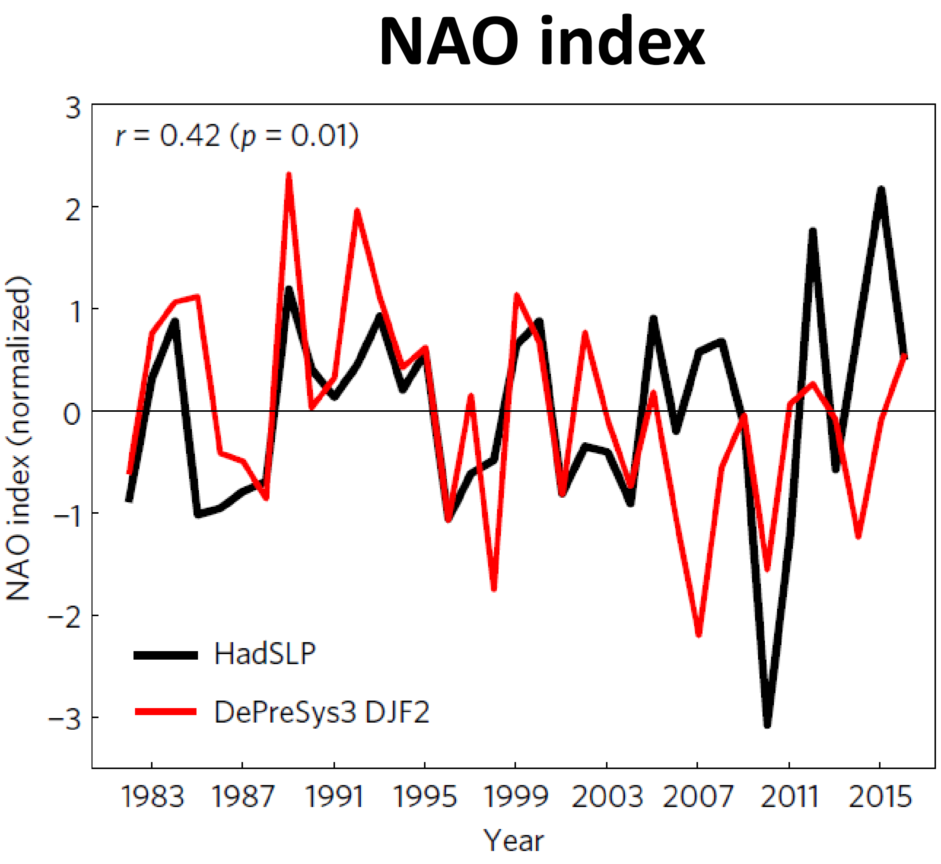
Scaife et al. (2014)

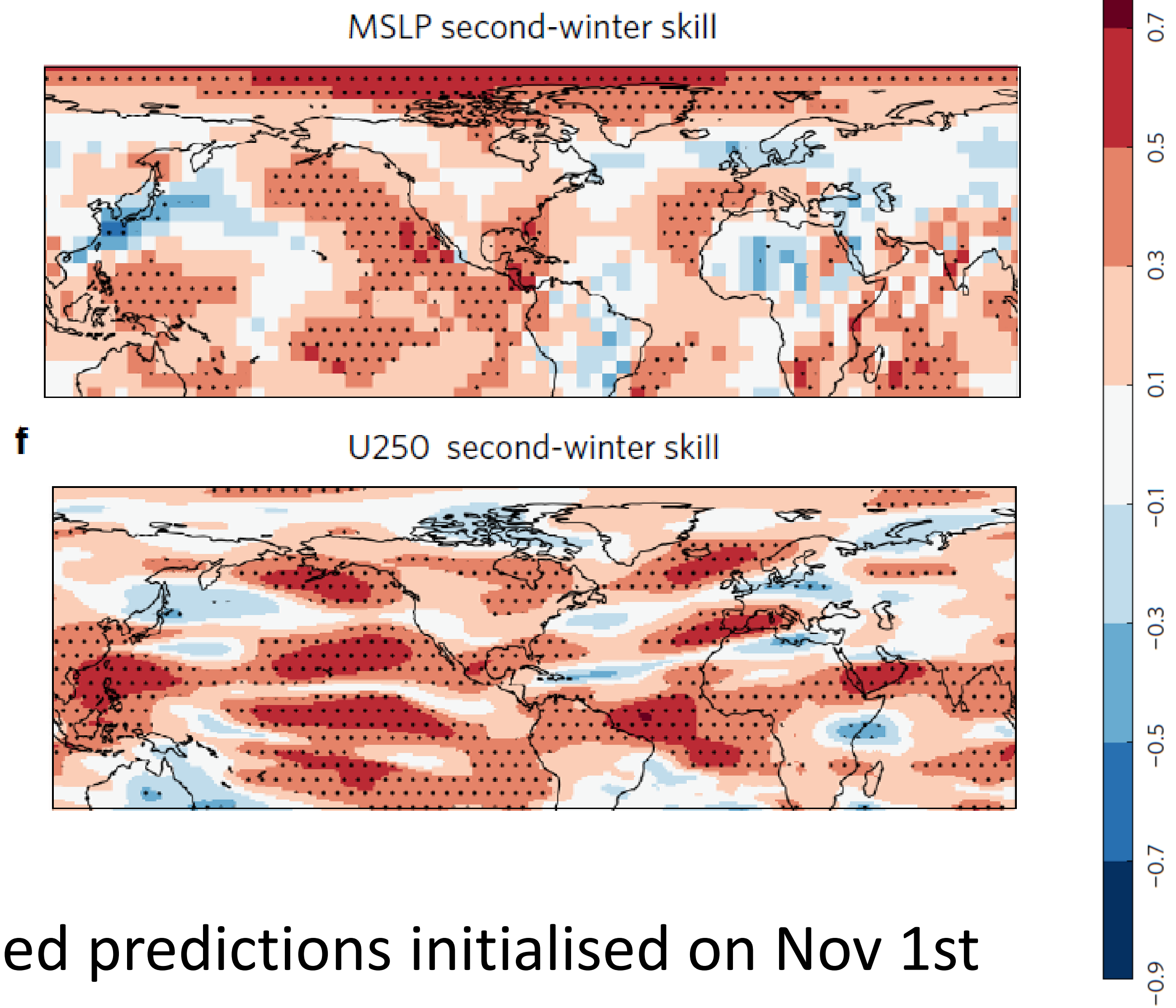


DJF NAO predicted from coupled predictions (GLOsea5) initialised on Nov 1st

Predictive skills in the extra-tropics

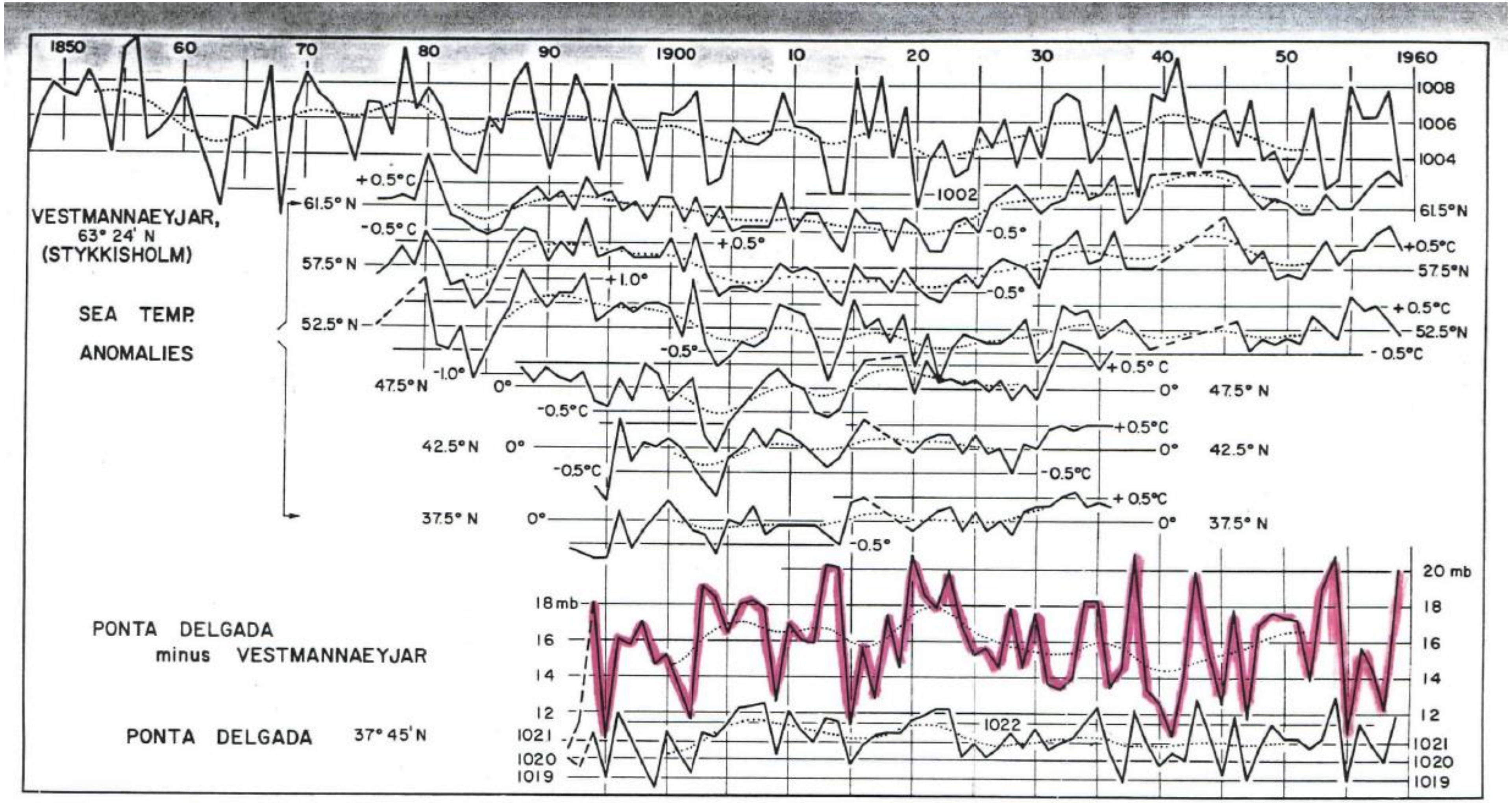
Dunstone et al. (2016)





“2nd” DJF NAO predicted from coupled predictions initialised on Nov 1st

Bjerknes (1964): Atlantic air-sea interactions



The role of ocean advection in altering SST

- The high latitude (>50N) correlation between strength of the westerlies and SST remains <0 for “short” and “long trends” (19-yr smoothing used)

- This contrasts with the 40-50N belt where positive correlations appear on long timescales

TABLE III. Correlation of short-period residuals of sea temperature versus pressure difference Ponta Delgada minus Vestmannaeyar.

“Short trend”

Period	Test field	Corr. coeff.
1900–1928	61.5°N	−0.48
1900–1928	57.5°N	−0.64
1900–1928	52.5°N	−0.72
1900–1928	47.5°N	−0.65
1900–1928	42.5°N	−0.45
1900–1928	37.5°N	−0.45

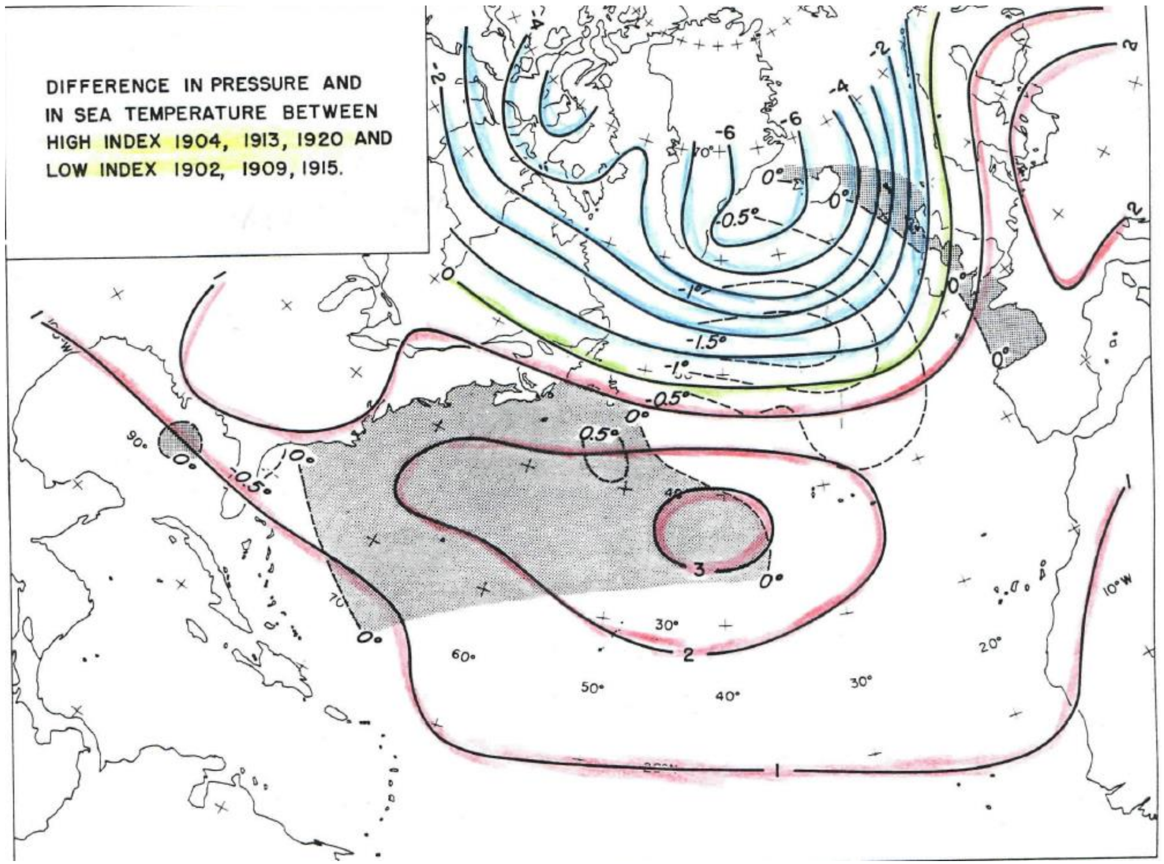
TABLE II. Correlation of sea temperature versus pressure difference Ponta Delgada minus Vestmannaeyar, all data identically smoothed.

“Long trend”

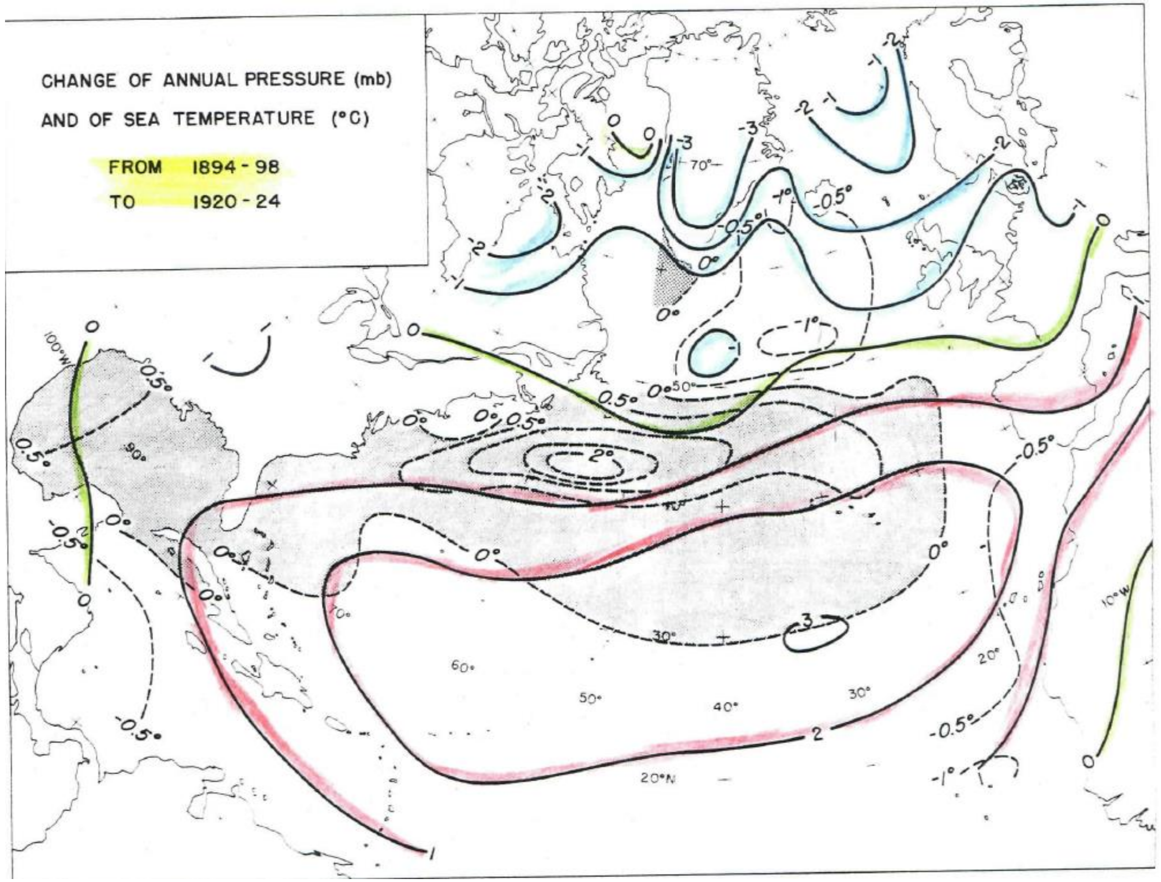
Period	Test field	Corr. coeff.
1900–1928	61.5°N	−0.34
1900–1928	57.5°N	−0.82
1900–1928	52.5°N	−0.82
1900–1928	47.5°N	0.18
1900–1928	42.5°N	0.33
1900–1928	37.5°N	−0.37

Spatial patterns of MSLP and SST

"Short trends of change"

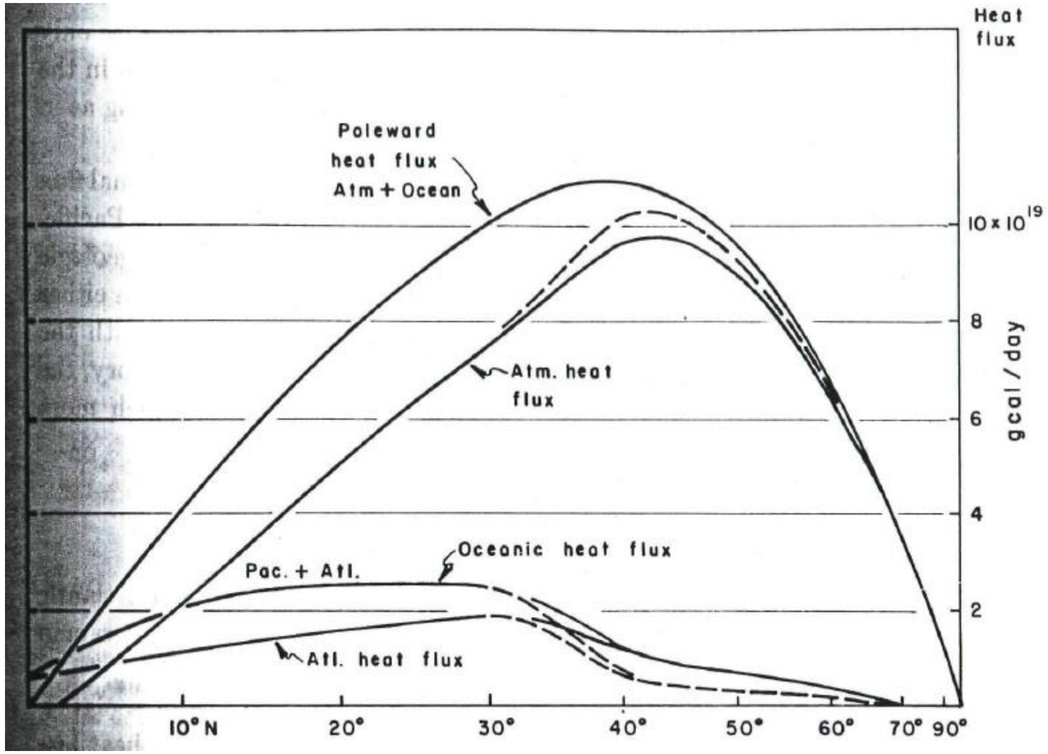


"Long trends of change"



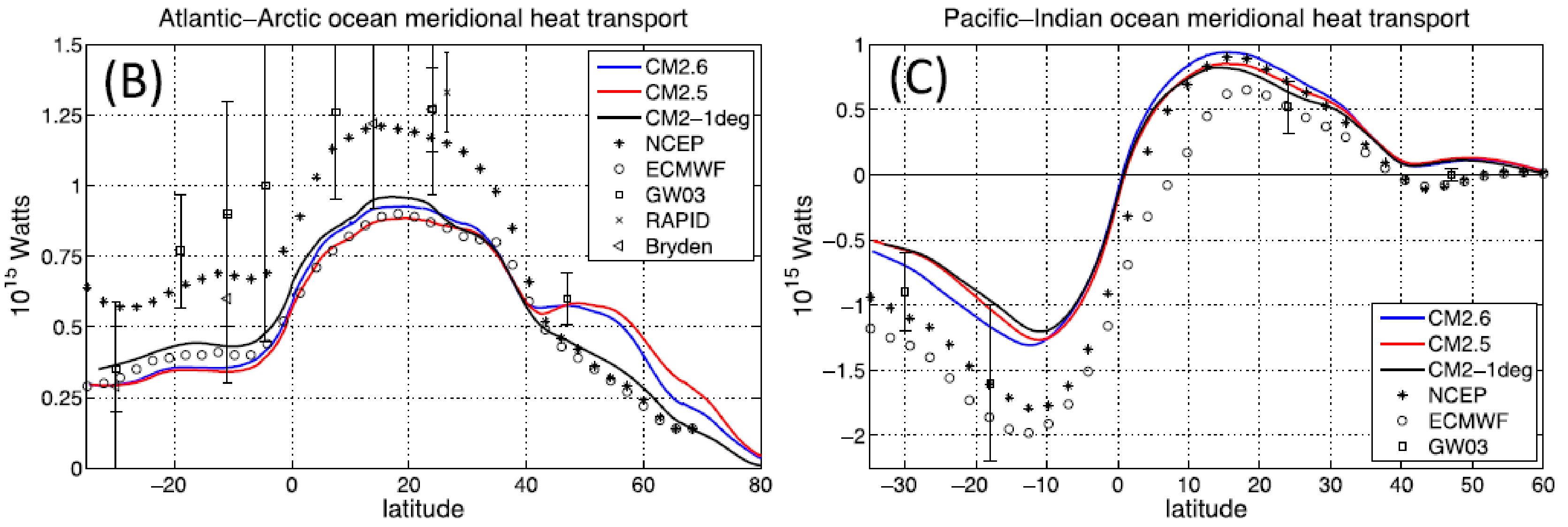
NB The pre / post 1950 records do not agree on spatial pattern (Deser and Blackmon, 1993)

Bjerknes' compensation hypothesis



- Mechanistically, periods of weak westerlies have higher than normal Ha, but also lead to a weakening of the Gulf Stream and a weaker Ho, thus re-establishing a constant total (Ho+Ha) poleward heat transport
- Climate variability reflects compensating fluctuations in oceanic (Ho) and atmospheric (Ha) poleward heat transports, without significant changes in the top-of-the-atmosphere radiative budget

Bjerknes' compensation hypothesis



Griffies et al. (2015)

- Because the Atlantic is the main contributor to oceanic poleward heat transport (>40N), it plays a leading role in "natural" climate fluctuations.
- Bjerknes' monograph is unclear as to causality and can be read as reflecting a passive response of the ocean to the atmosphere. But it has opened the way to much emphasis on driving of climate variability by the midlatitude oceans.

2. How do SST changes affect the atmosphere above the boundary layer?

- Diabatic heating processes
- Theory: prescribe a heat source and predict the atmospheric circulation response
- Challenge: apply the theory to observations and realistic numerical experiments

Diabatic heating processes

- Any process adding or removing heat to/from air parcels:
- Exchange with Space: radiation (absorption or shortwave, absorption/emission of longwave radiation)
- Exchange with the lower boundary (ocean, land, ice): radiative and turbulent heat fluxes ("sensible" heat flux)
- Phase change: depending whether one works with dry or moist potential temperature, this is taken as an exchange with the lower boundary (surface evaporation) or an internal heat source (latent heat release) –see Emmanuel, 2000; Pauluis et al., 2010.

Diabatic heating in models (GCMs or diagnostics)

$$\frac{\partial \theta}{\partial t} + \boldsymbol{v}.\boldsymbol{\nabla}\theta = Q_{diab}$$

- Need distinguishing between resolved (<>) and unresolved (*) processes, or slow (<>) and fast (*) motions:

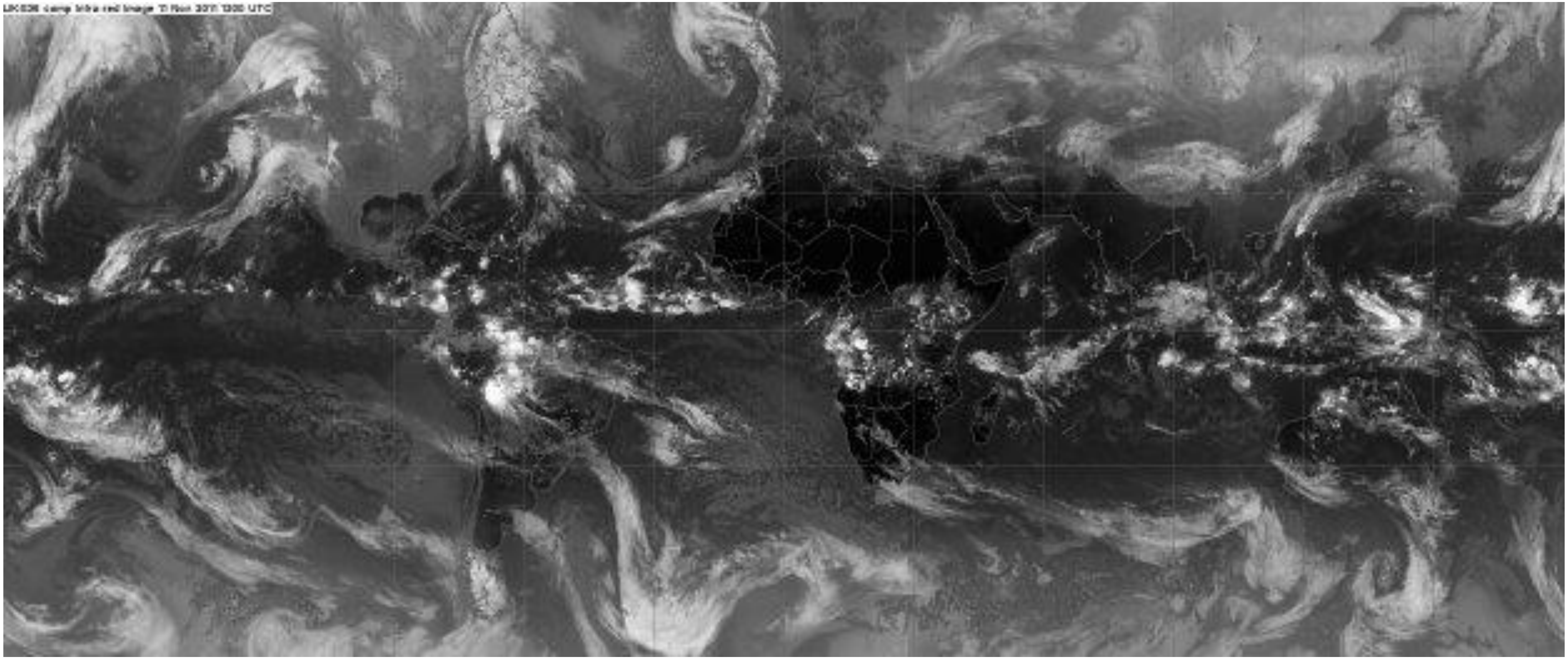
$$\frac{\partial < \theta >}{\partial t} + < \boldsymbol{v} > .\boldsymbol{\nabla} < \theta > = < Q_{diab} > - \boldsymbol{\nabla}. < \boldsymbol{v}^{\star}\theta^{\star} > \qquad (1)$$

with $$< Q_{diab} > = < Q_{rad} > + < Q_{phase} > + < Q_{sen} >$$

- The terms on the r.h.s of (1) represent an *apparent heating* for the resolved flow which can be quite different from the assumed diabatic heating taking place:

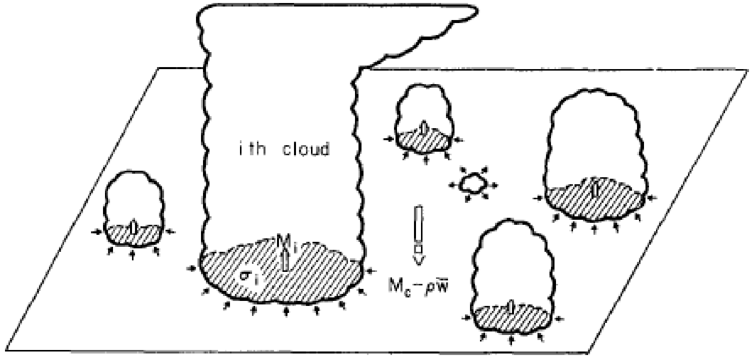
$$< Q_{app} > = < Q_{diab} > - \boldsymbol{\nabla}. < \boldsymbol{v}^{\star}\theta^{\star} >$$

Diabatic heating: two examples



Global Infrared image (snapshot): white = cold = cloud tops

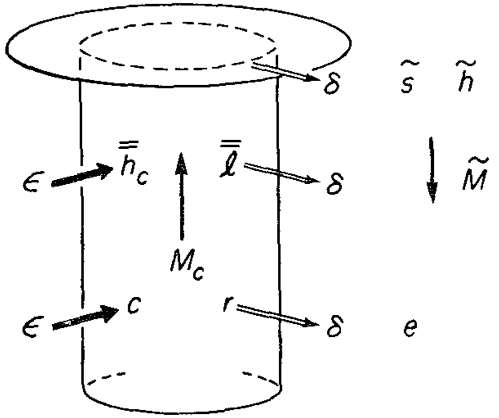
Diabatic heating example 1: cumulus parameterisation (e.g., Yanai et al. 1973)



- Neglecting horizontal effects:

$$<Q_{app}> = <Q_{diab}> - \nabla . <\boldsymbol{v}^{\star}\theta^{\star}> \approx <Q_{diab}> - \frac{<w^{\star}\theta^{\star}>}{\partial z}$$

- …and using an idealised entrainment/detrainment model for the cloud ensemble, one can obtain an expression for the apparent heating:

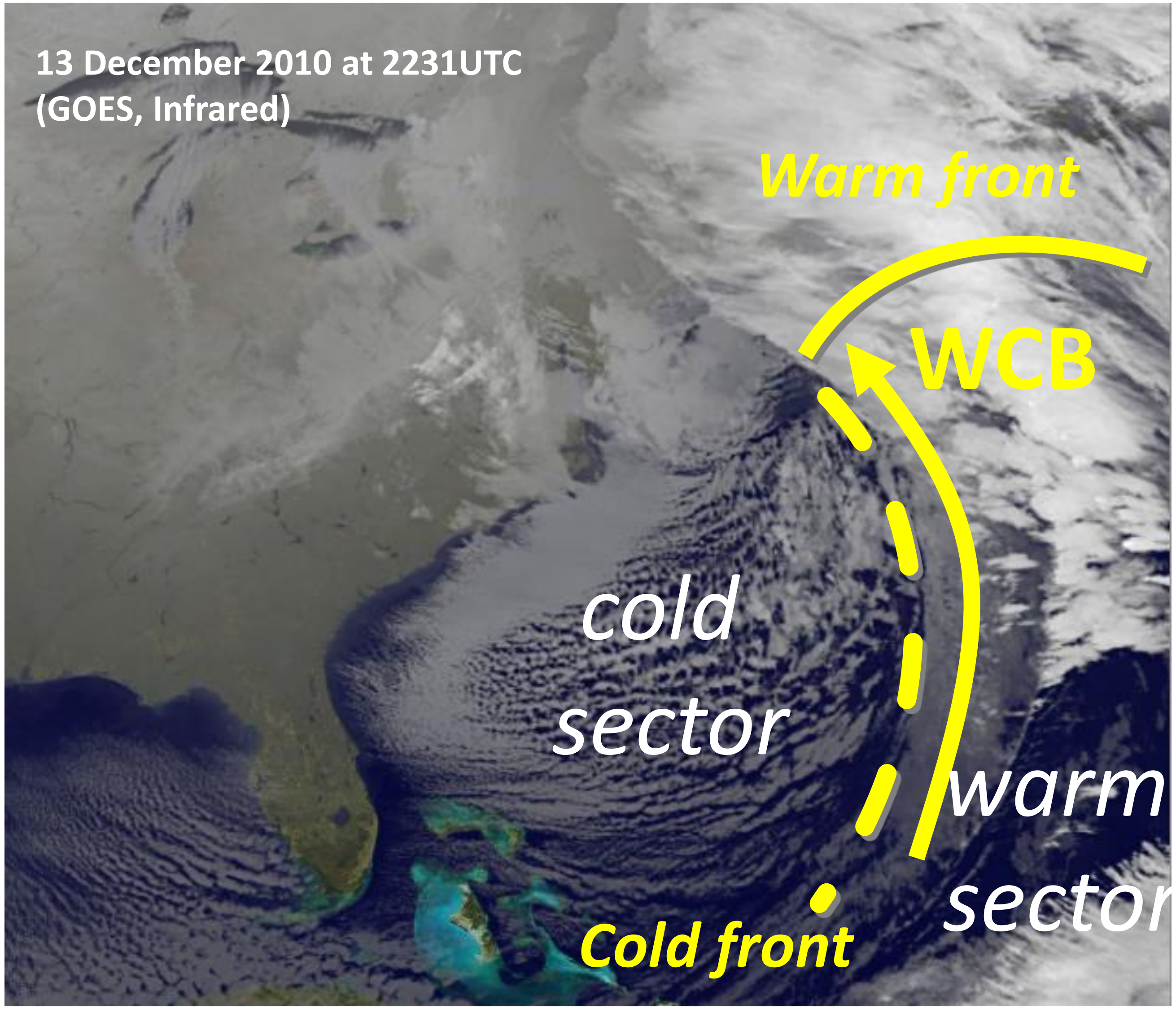


$$<Q_{app}> \approx <Q_{rad}> + <Q_{sen}> + M_c \frac{\partial <\theta>}{\partial z} - l_v e$$

Warming due to compensating subsidence (Mc=cloud mass flux)

Cooling due to re-evaporation of cloud droplets

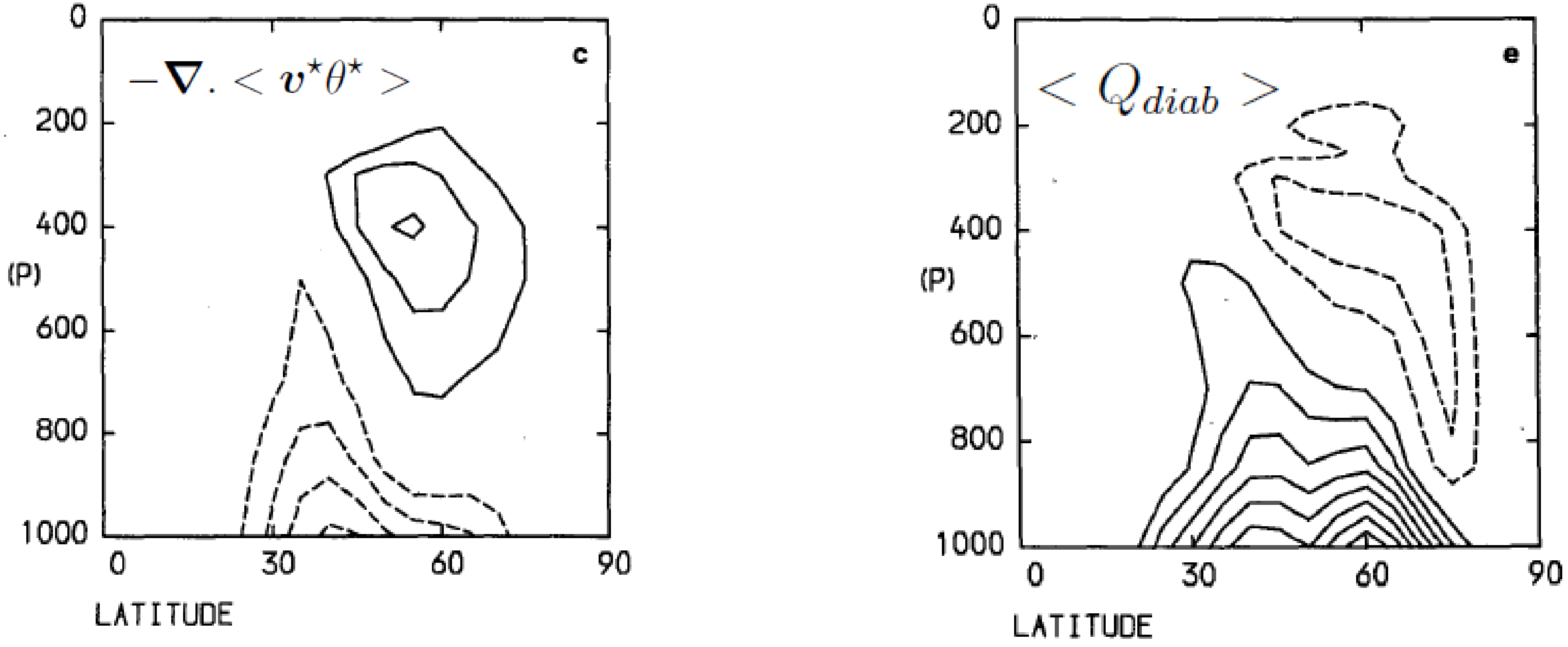
Diabatic heating example 2: Extra-tropical cyclones



Diabatic heating example 2: extra-tropical cyclones or "storm-track" (Hoskins & Valdes, 1990)

Averaged heating over the Northwest Atlantic (CI=0.25K/day)

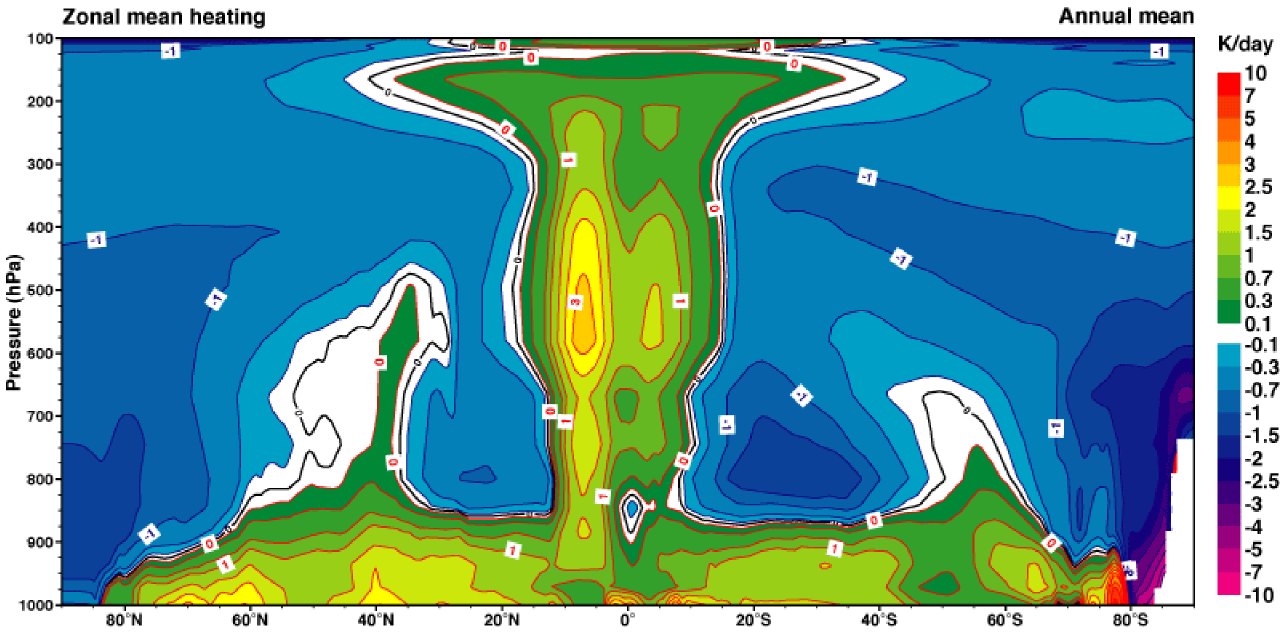
NB 2.5-6 day high-pass filter applied



- Qapp is a small residual between diabatic heating and thermal forcing by eddies

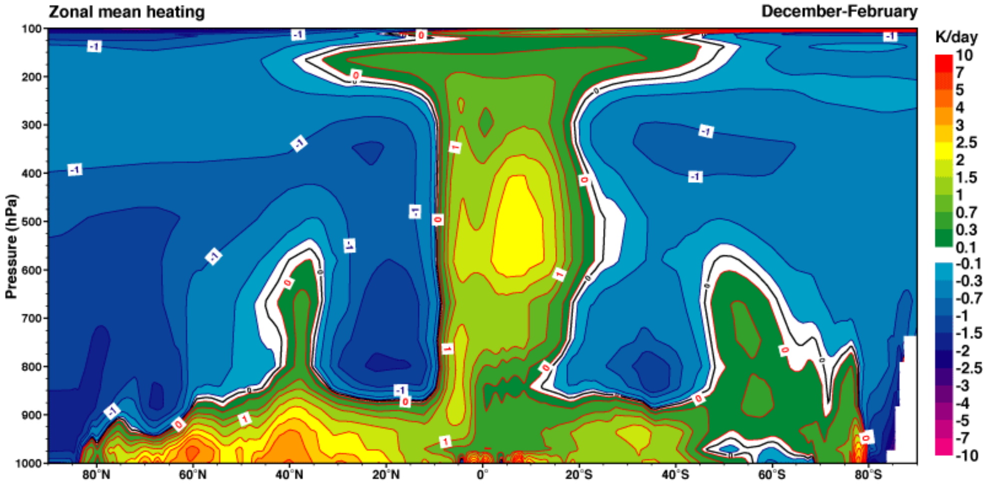
Qdiab climatology from ERA40 atlas (K/day)

Annual mean

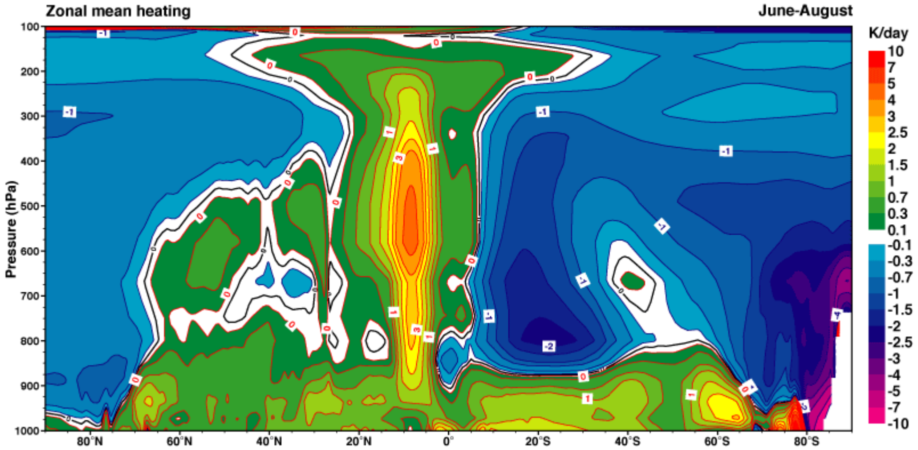


NB Computed from the net forecast tendency as a residual in the thermodynamic equation

DJF



JJA



Response of the atmosphere (temp., winds) to a prescribed heating anomaly

- Goal is to gain understanding as to what maintains a climatic anomaly or perturbation (e.g., a positive phase of the NAO or El Nino conditions averaged over many such events)

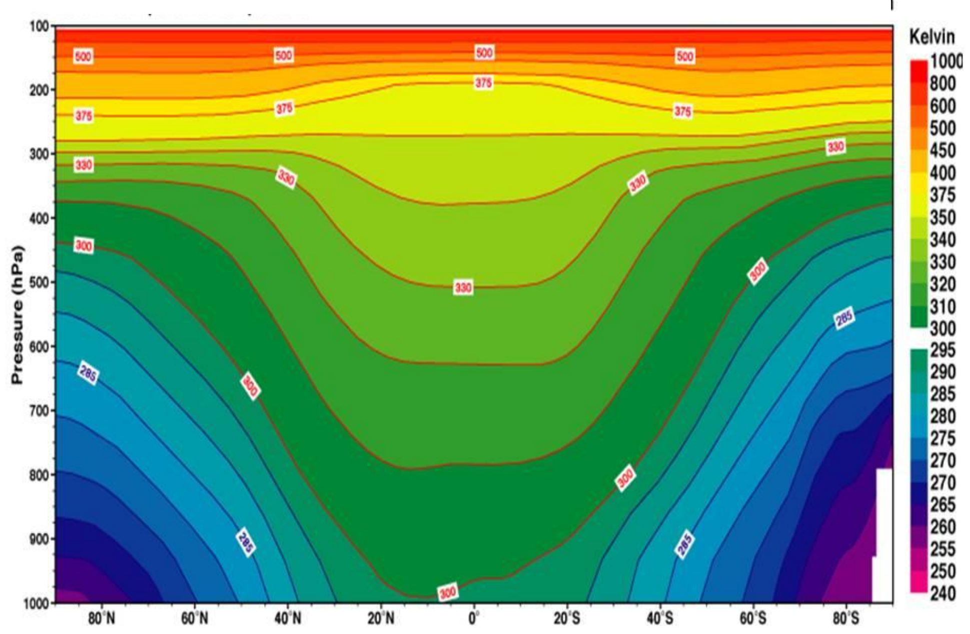
- The overbar represent the “normal” state while primes denote the “anomalous” state

- Need to do this to increase our confidence that numerical models do the right thing

Basic theory: Hoskins and Karoly (1981)

NB Notes provide derivation of this model

Annual and zonal mean of θ



- Basic state (overbar) = zonal flow in thermal wind balance
- Forcing: prescribed heat source
- Frictional effects entirely neglected
- Perturbations (primes) are geostrophic and obey linear conservation of vorticity and heat (entropy):

$$\bar{u}\xi_x' + \beta v' = f w_z',$$

$$\bar{u}\theta_x' + v'\bar{\theta}_y + w'\bar{\theta}_z = (\theta_0/g)Q.$$

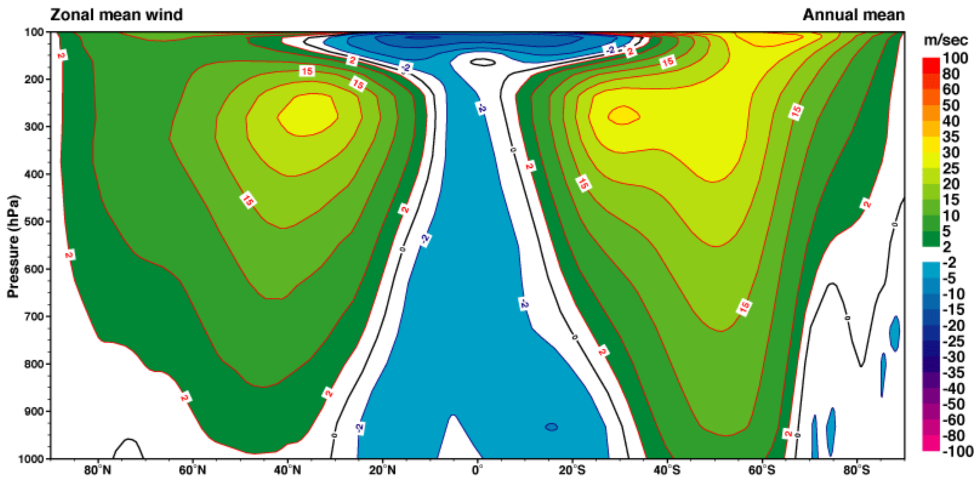
Prescribed heat source

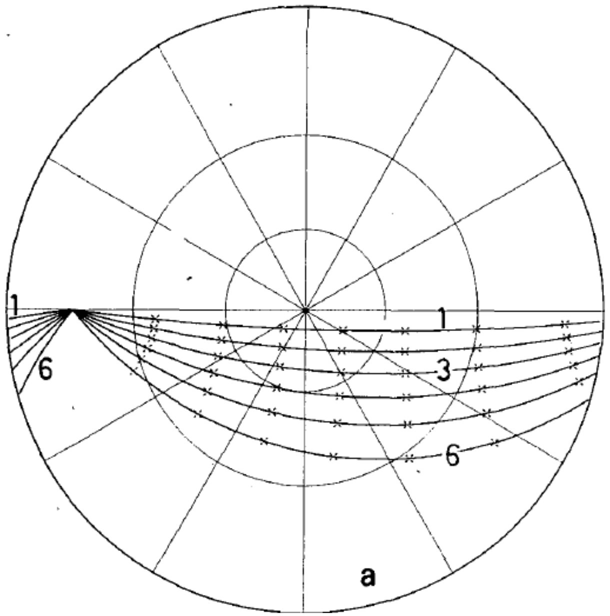
Basic theory: Hoskins and Karoly (1981)

- Because the zonal winds increase with height in the troposphere, the zonal advection term is more important at upper than at lower levels.

- At upper levels, vorticity conservation is a stationary Rossby wave equation:

Annual and zonal mean of **U**



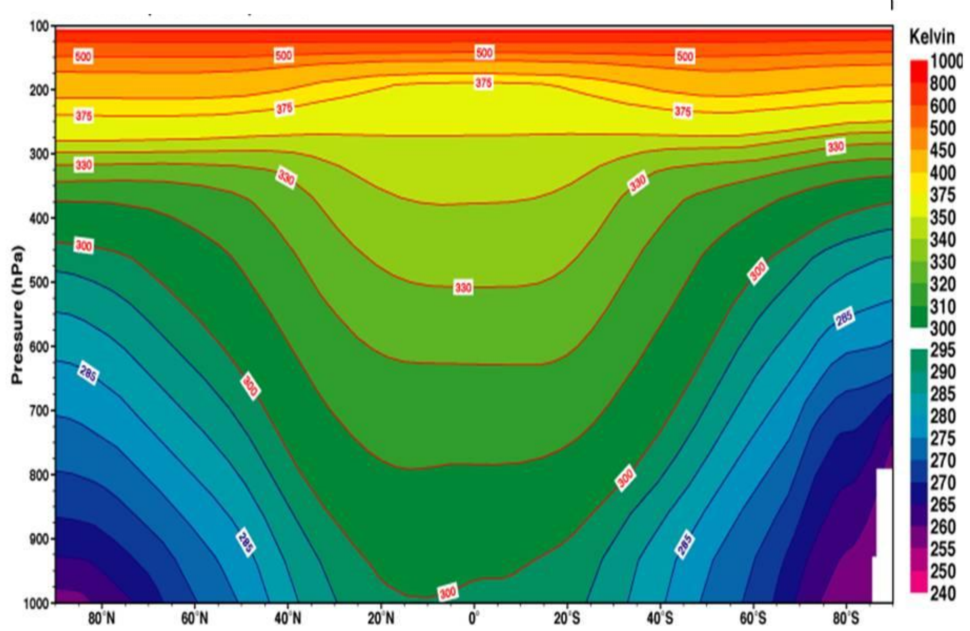


$$\bar{u}\xi_x' + \beta v' = f w_z',$$

Localized Rossby wave source → remote forcing of wind anomalies (anticyclones generated above ascending regions)

Basic theory: Hoskins and Karoly (1981)

Annual and zonal mean of θ



- Simplification: we restrict ourselves to lower levels and long waves (>1000 km) for which vorticity conservation reduces to Sverdrup balance:

$$\bar{u}\xi_x' + \beta v' = f w_z',$$

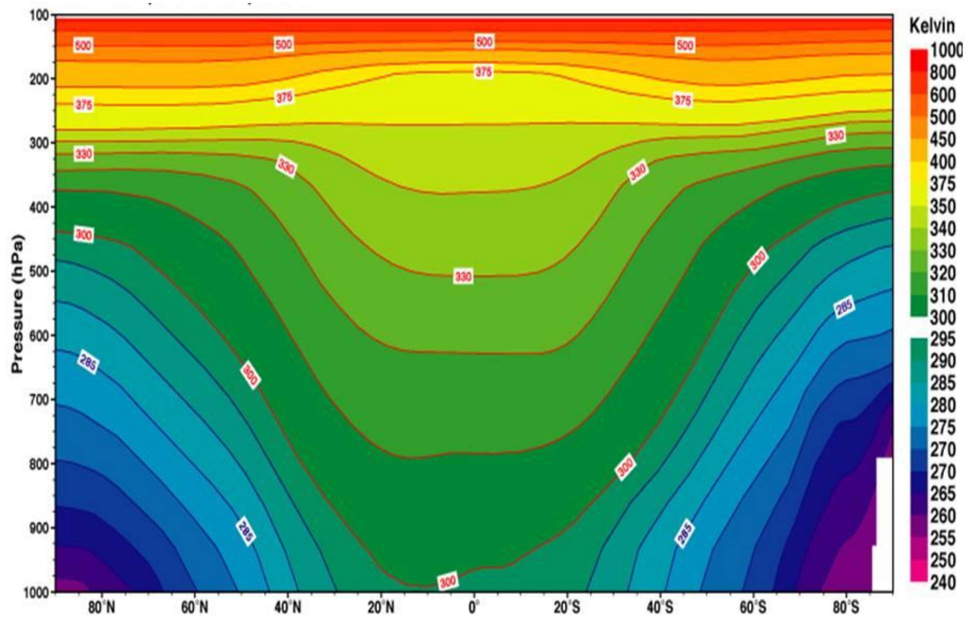
$$\bar{u}\theta_x' + v'\bar{\theta}_y + w'\bar{\theta}_z = (\theta_0/g)Q.$$

Remember David's lecture yesterday

Basic theory: Hoskins and Karoly (1981)

- Qualitative understanding by looking under which conditions we can simply consider only one term on the l.h.s of the heat equation:

Annual and zonal mean of θ



$$\bar{u}\xi_x' + \beta v' = f w_z',$$

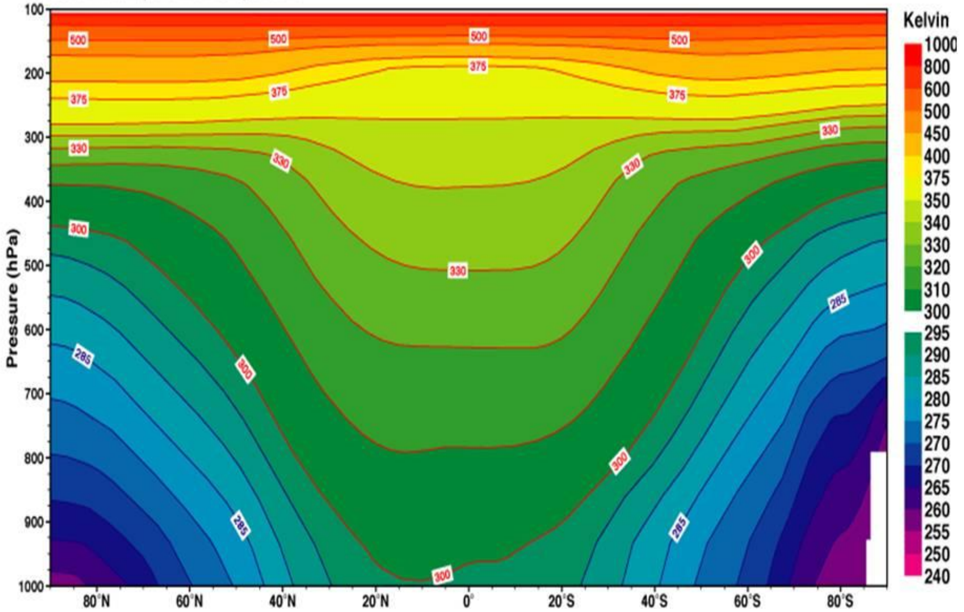
$$\bar{u}\theta_x' + v'\bar{\theta}_y + w'\bar{\theta}_z = (\theta_0/g)Q.$$

?

Basic theory: Hoskins and Karoly (1981)

- Qualitative understanding by looking under which conditions we can simply consider only one term on the l.h.s of the heat equation:

Annual and zonal mean of θ



$$\bar{u}\xi_x' + \beta v' = f w_z',$$

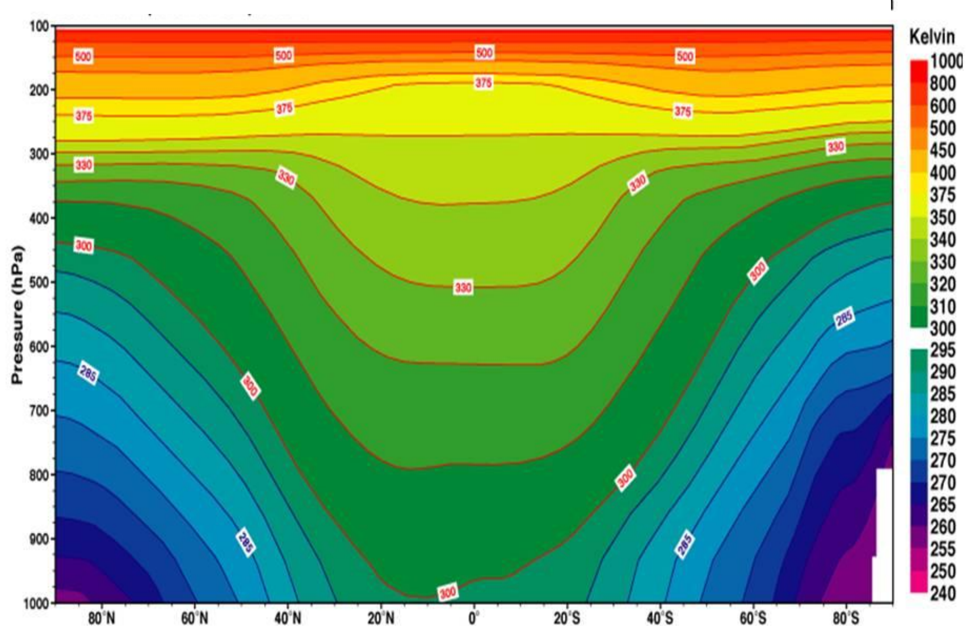
$$\bar{u}\theta_x' + v'\bar{\theta}_y + w'\bar{\theta}_z = (\theta_0/g)Q.$$

?

Basic theory: Hoskins and Karoly (1981)

- Qualitative understanding by looking under which conditions we can simply consider only one term on the l.h.s of the heat equation:

Annual and zonal mean of θ



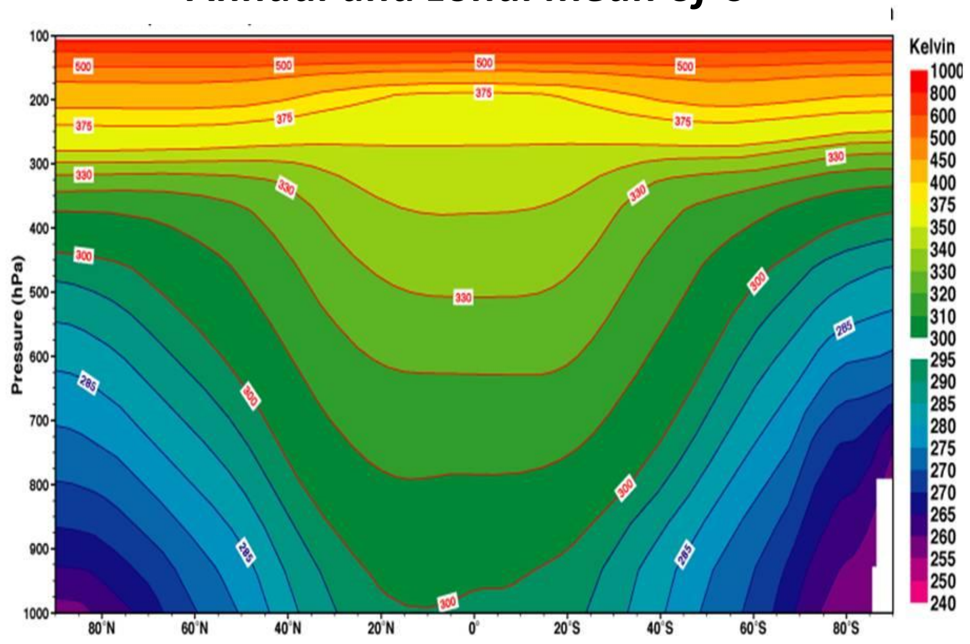
$$\bar{u}\xi_x' + \beta v' = f w_z',$$

$$\bar{u}\theta_x' + v'\bar{\theta}_y + w'\bar{\theta}_z = (\theta_0/g)Q. \quad ?$$

Basic theory: Hoskins and Karoly (1981)

- Use the thermal wind to rewrite the heat equation as:

Annual and zonal mean of θ



Buoyancy frequency

$$N^2 = \frac{g}{\theta o}\frac{d\theta}{dz}$$

$$f\bar{u}v_z{}' - f\bar{u}_z v' + w' N^2 = Q.$$

Zonal advection of heat

Meridional advection of heat

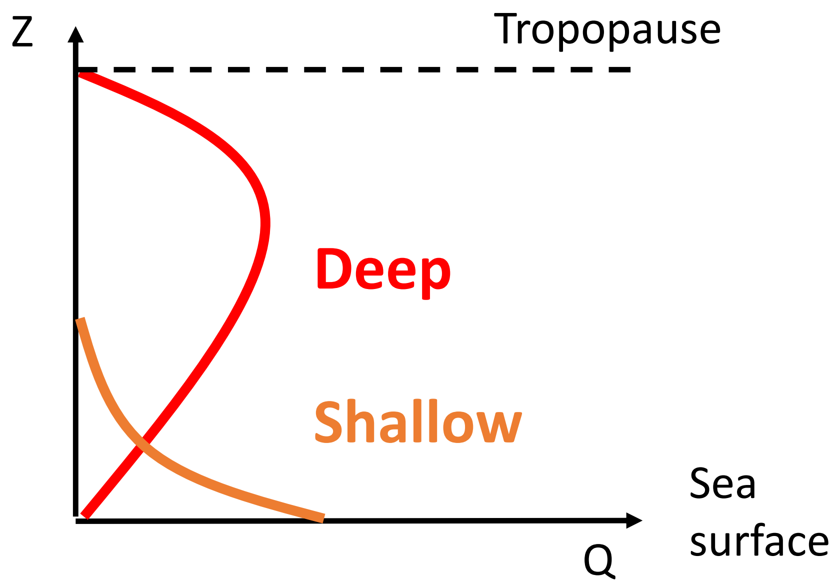
Vertical advection of heat

Basic theory: Hoskins and Karoly (1981)

- If zonal advection of heat dominates, then:

$$v' \simeq QH_Q/f\overline{u} \quad \text{with} \quad H_Q \equiv Q/Q_z$$

(a few kms for shallow source, 5-10km for a deep source)



$$f\bar{u}v_z' - f\bar{u}_z v' + w'N^2 = Q.$$

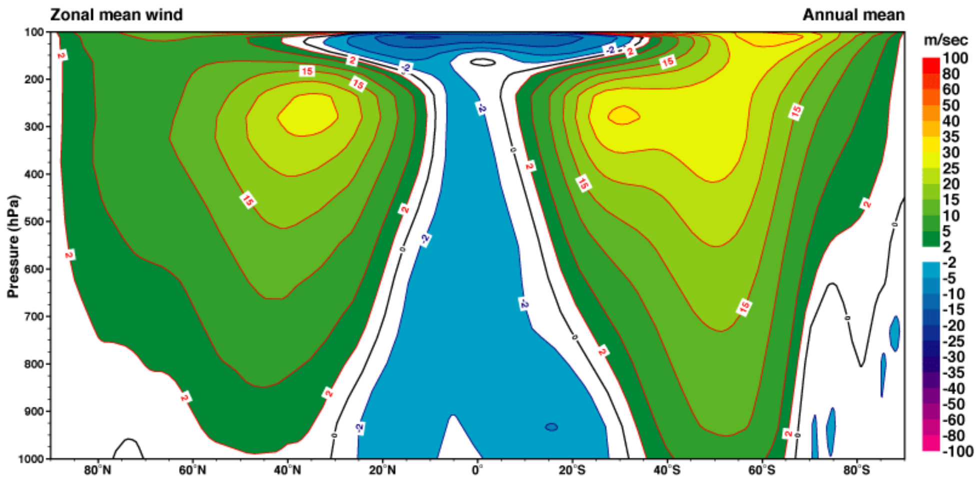
Zonal advection of heat

Meridional advection of heat

Vertical advection of heat

Basic theory: Hoskins and Karoly (1981)

***Annual and zonal mean of* U**



- If meridional advection of heat dominates, then:

$$v' \simeq Q/f\overline{u}_z = QH_u/f\overline{u} \quad \text{with} \quad H_u \equiv \overline{u}/\overline{u}_z$$

~10ms-1/(30ms-1/10km) ~3km

$$f\bar{u}v_z' - f\bar{u}_z v' + w'N^2 = Q.$$

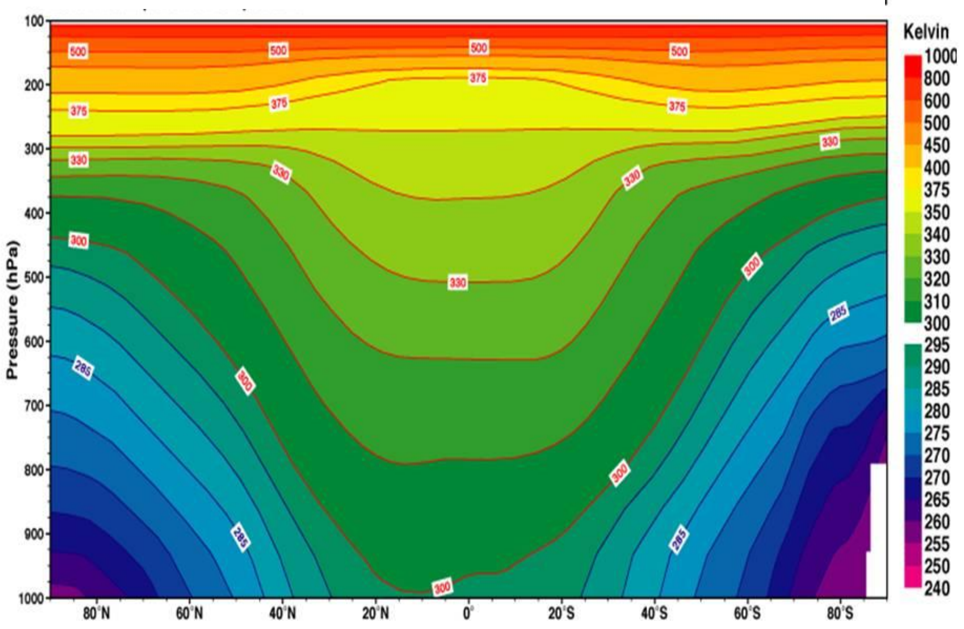
Zonal advection of heat

Meridional advection of heat

Vertical advection of heat

Basic theory: Hoskins and Karoly (1981)

Annual and zonal mean of θ



- If vertical advection of heat dominates, then:

$$v' \simeq f w' / \beta H_Q = f Q / \beta N^2 H_Q$$

$$f\bar{u}v_z' - f\bar{u}_z v' + w'N^2 = Q.$$

Zonal advection of heat — Meridional advection of heat — Vertical advection of heat

Basic theory: Hoskins and Karoly (1981)

- Key assumption: mechanism with smallest v' will dominate (this is a thermodynamic argument as the energy source for the motion is heating and this must somehow be converted to kinetic energy)

(1) $$v' \simeq QH_Q/f\overline{u}$$ (if zonal adv. dominates)

(2) $$v' \simeq Q/f\overline{u}_z = QH_u/f\overline{u}$$ (if meridional adv. dominates)

(3) $$v' \simeq fw'/\beta H_Q = fQ/\beta N^2 H_Q$$ (if vertical adv. dominates)

Basic theory: Hoskins and Karoly (1981)

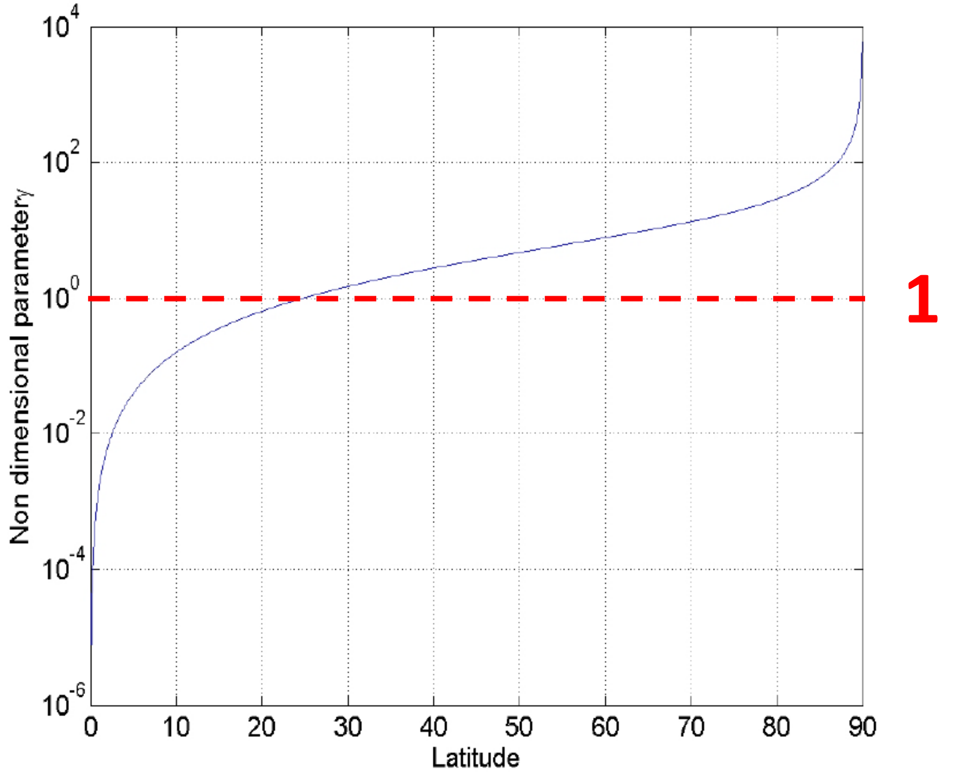
- The ratio of (3) to (1) or (2) is a non dimensional number:

$$\gamma = f^2\bar{u}/(\beta N^2 H_Q H)$$

where H = min(Hu,HQ)

- γ >>1 in midlatitudes (hor. adv. wins)

- γ <<1 in Tropics (vert. adv. wins)



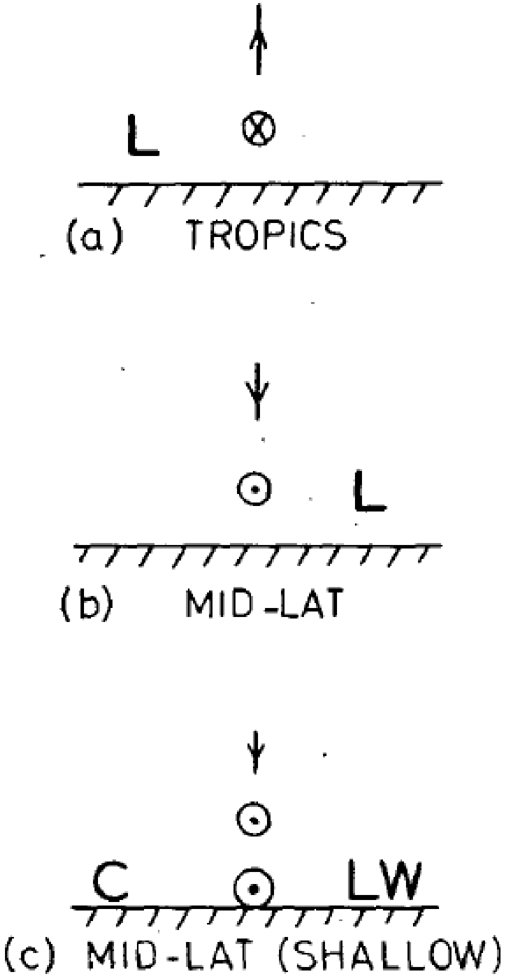
$H_Q = 6km, H_u = 3km, N = 10^{-2}s^{-1}$ and $\overline{u} = 10ms^{-1}$

Basic theory: Hoskins and Karoly (1981)

- γ <<1 in Tropics (vert. adv. wins): upward motion is in phase with the deep heat source. Sverdrup balance requires this to have poleward motion below and thus a low pressure to the west at low levels

- γ >>1 in midlatitudes: if deep heat source then meridional adv. wins at low levels and a low pressure is found to the east of the heat source. <u>*This implies sinking motion*</u> to accommodate Sverdrup balance.

- γ >>1 in midlatitudes: if shallow heat source then zonal adv. wins at low levels and a cold/warm dipole is found across the source. Sinking motion is again produced.



"Gill's response" to heating (1980)

Surface wind and vertical motion at midlevel

Surface pressure and winds

Walker cell (longitude-height plane)

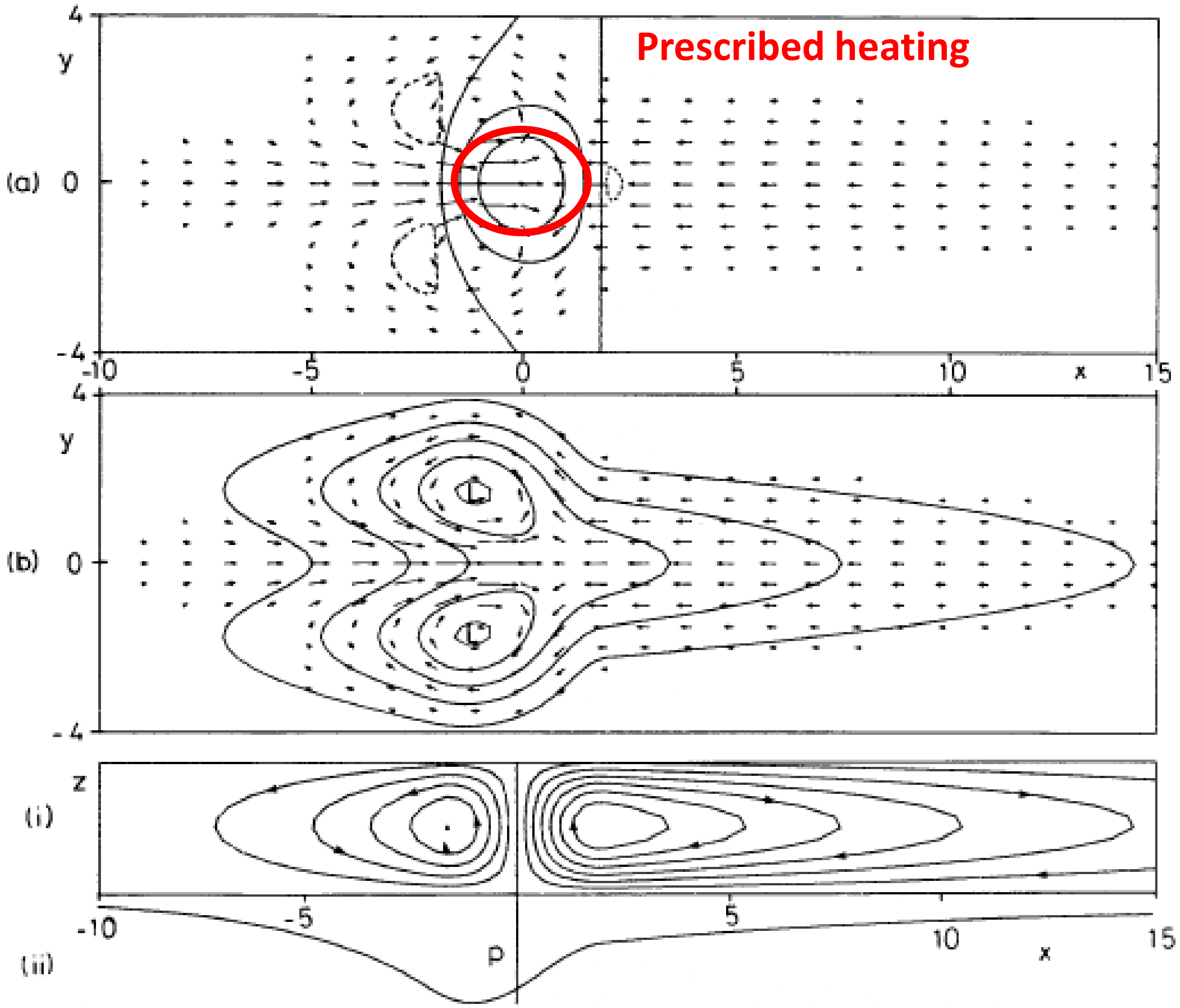


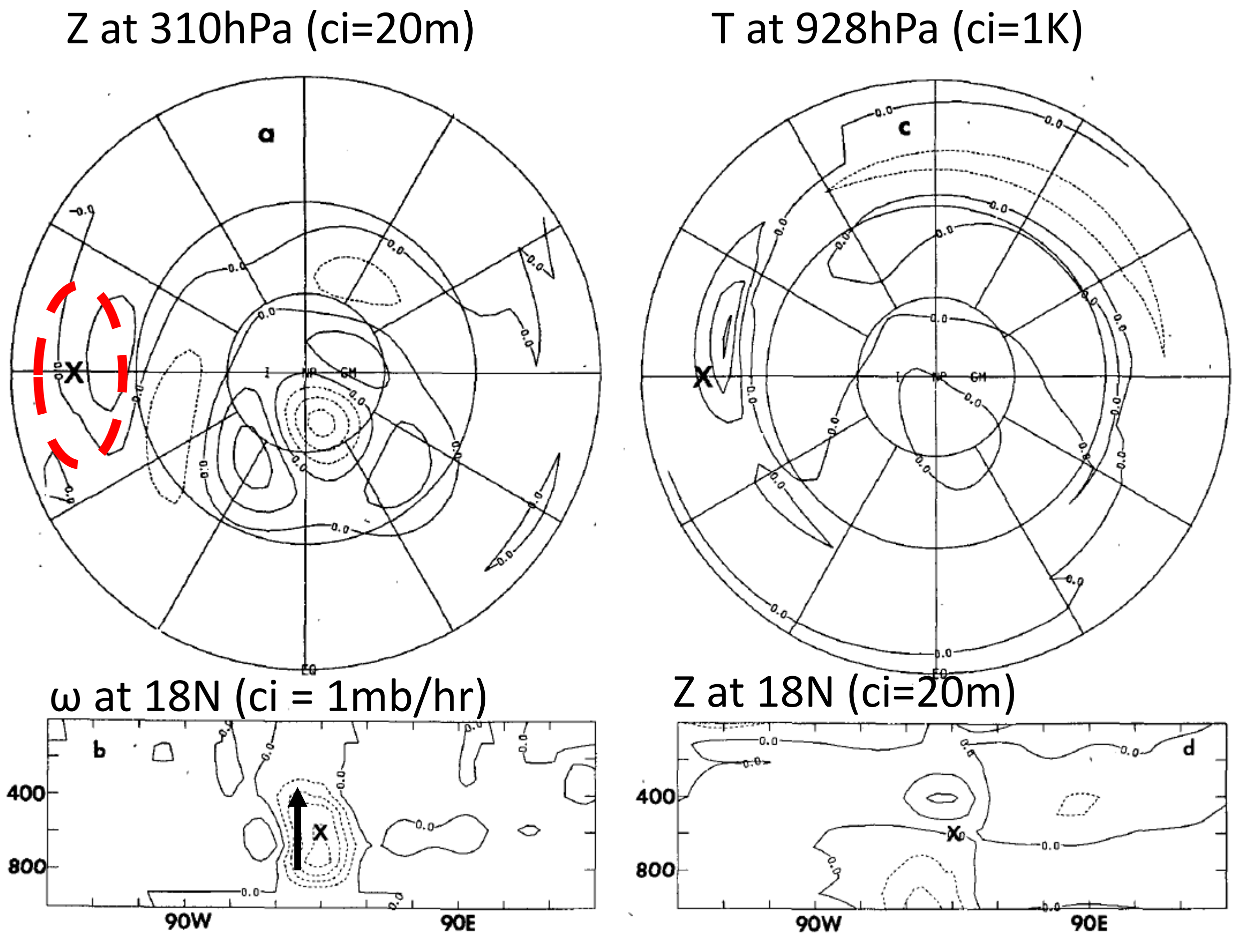
Figure 1. Solution for heating symmetric about the equator in the region $|x| < 2$ for decay factor $\epsilon = 0{\cdot}1$.
(a) Contours of vertical velocity w (solid contours are 0, 0·3, 0·6, broken contour is −0·1) superimposed on the velocity field for the lower layer. The field is dominated by the 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).

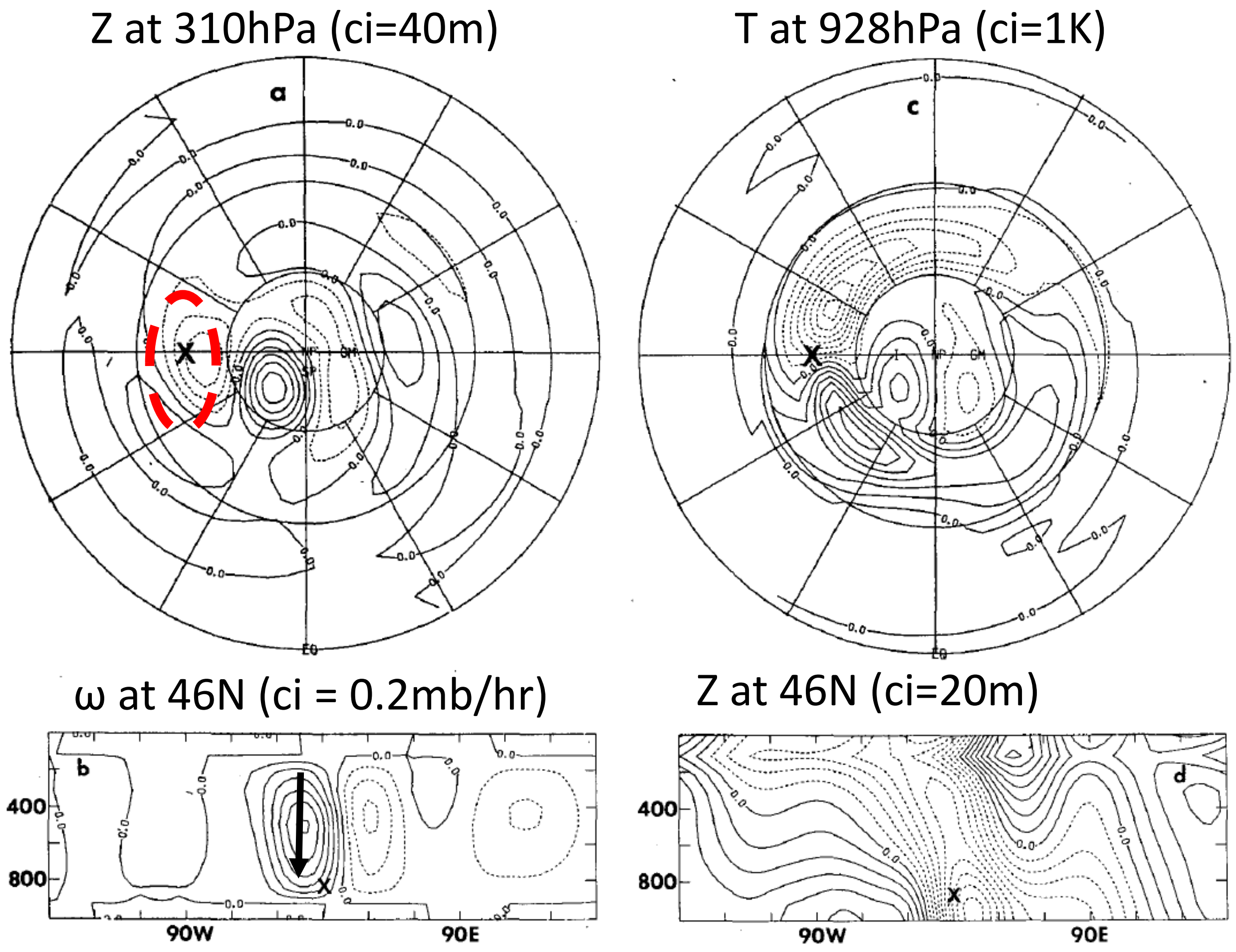
Hendon & Hartmann's experiments (1982)

- Response to a deep heat source (60 deg lon X 30 deg lat, vertically averaged heating of 350Wm-2) centered at 15N
- Upward motion balances the heating
- Low pressure west of the source at low levels
- Wave train propagating northeastward is generated at upper levels



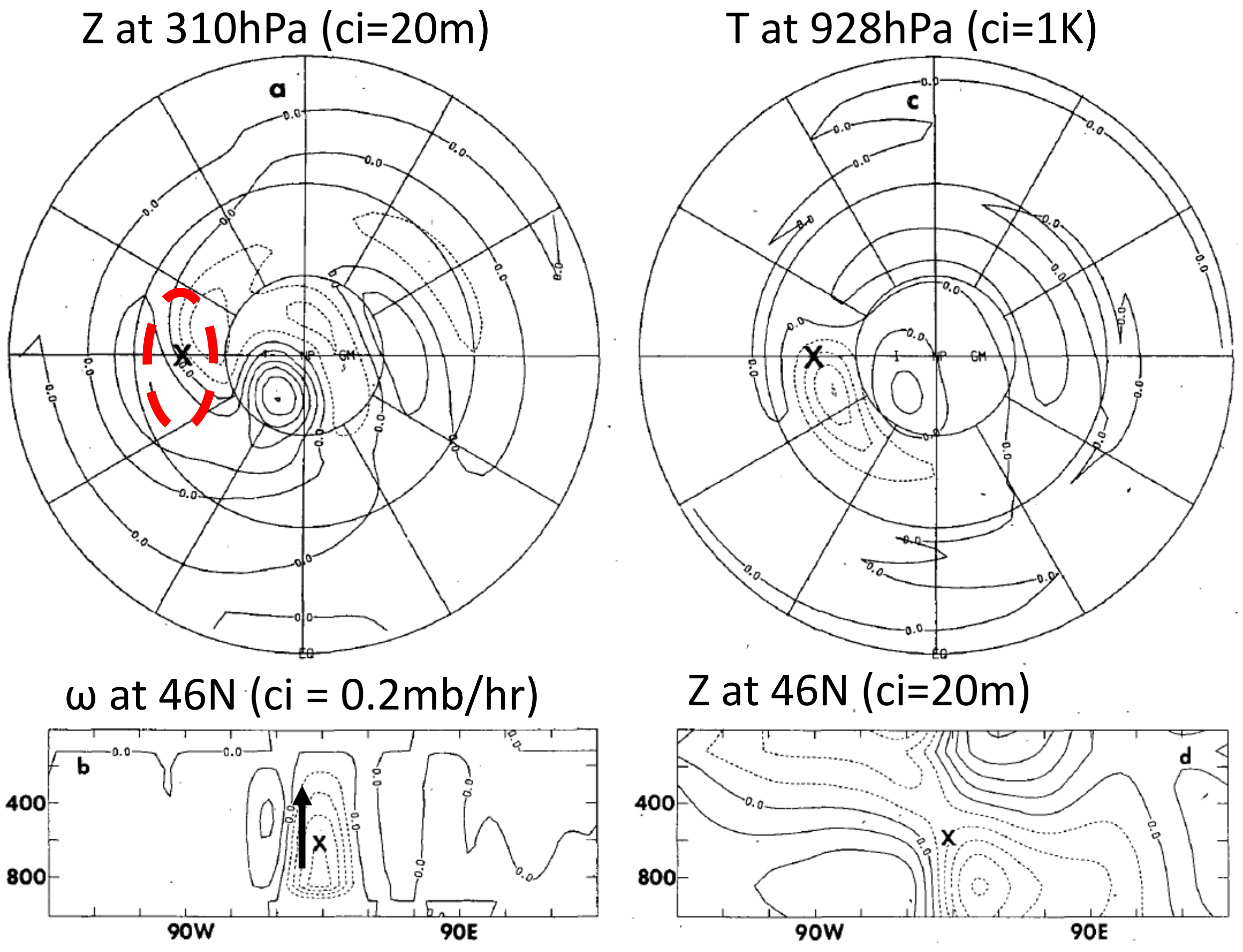
Hendon & Hartmann's experiments (1982)

- Response to a shallow heat source (60 deg lon X 30 deg lat, vertically averaged heating of 350Wm-2) centered at 46N
- Downward motion over the source & horizontal advection balances the heating
- Strong low pressure east of the source at low levels
- Strong remote response (anticyclone) generated at upper levels



Hendon & Hartmann’s experiments (1982)

- Response to a deep heat source (60 deg lon X 30 deg lat, vertically averaged heating of 350Wm-2) centered at 46N
- Upward and equatorward motion balances the heating & a cold anomaly is produced(!)
- Low pressure east of the source at low levels
- Split wave train is generated at upper levels



Challenge…

- Use the previous knowledge to compare to more realistic experiments and observations

- Reminder:

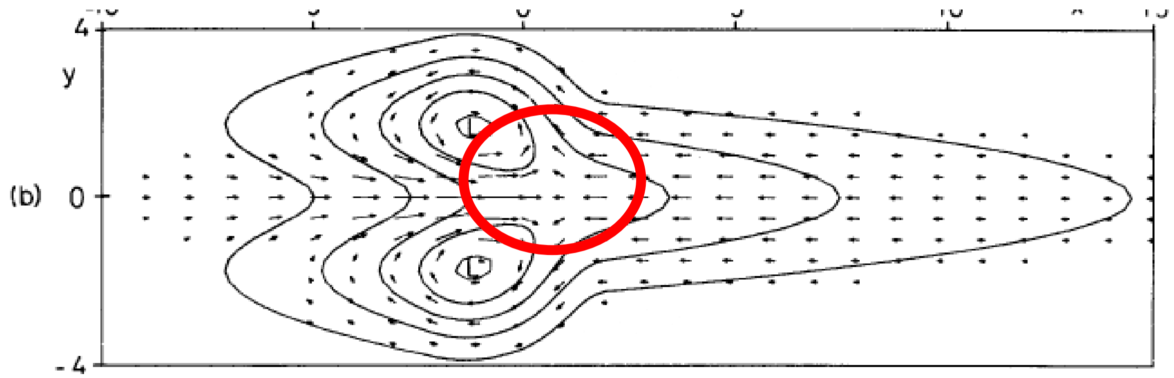
$$< Q_{app} > = < Q_{diab} > - \boldsymbol{\nabla} . < v^{\star} \theta^{\star} >$$

Call this <Qfast> (either subgrid scale or high frequency)

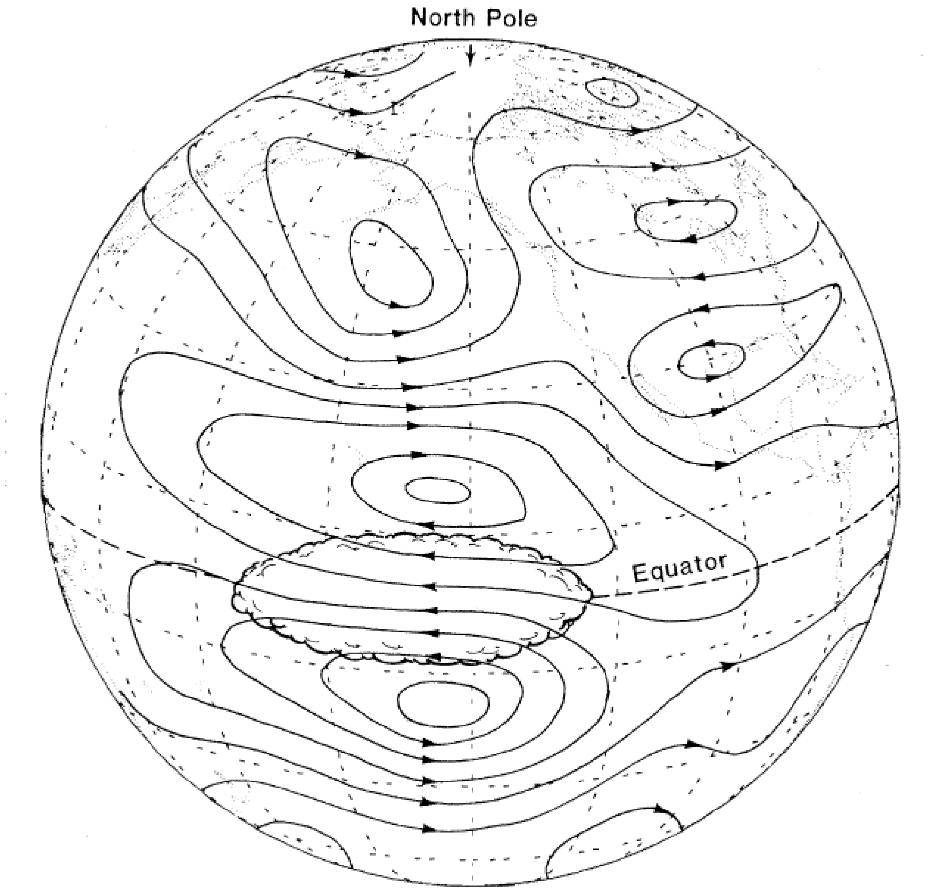
Gill's response in observations

Upper level circulation at the peak of the 1982-83 El Nino

- Upper level pressure field is the negative of the surface one in Gill's solution



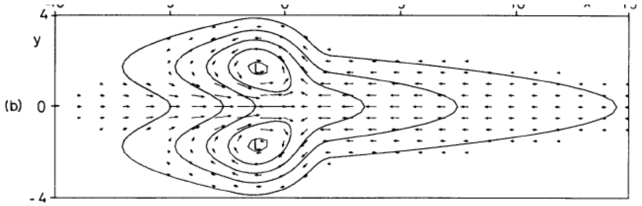
Prescribed heating anomaly

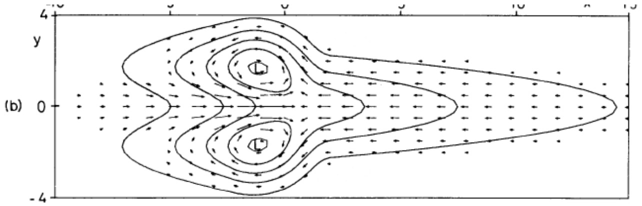


Rasmusson and Wallace (1983)

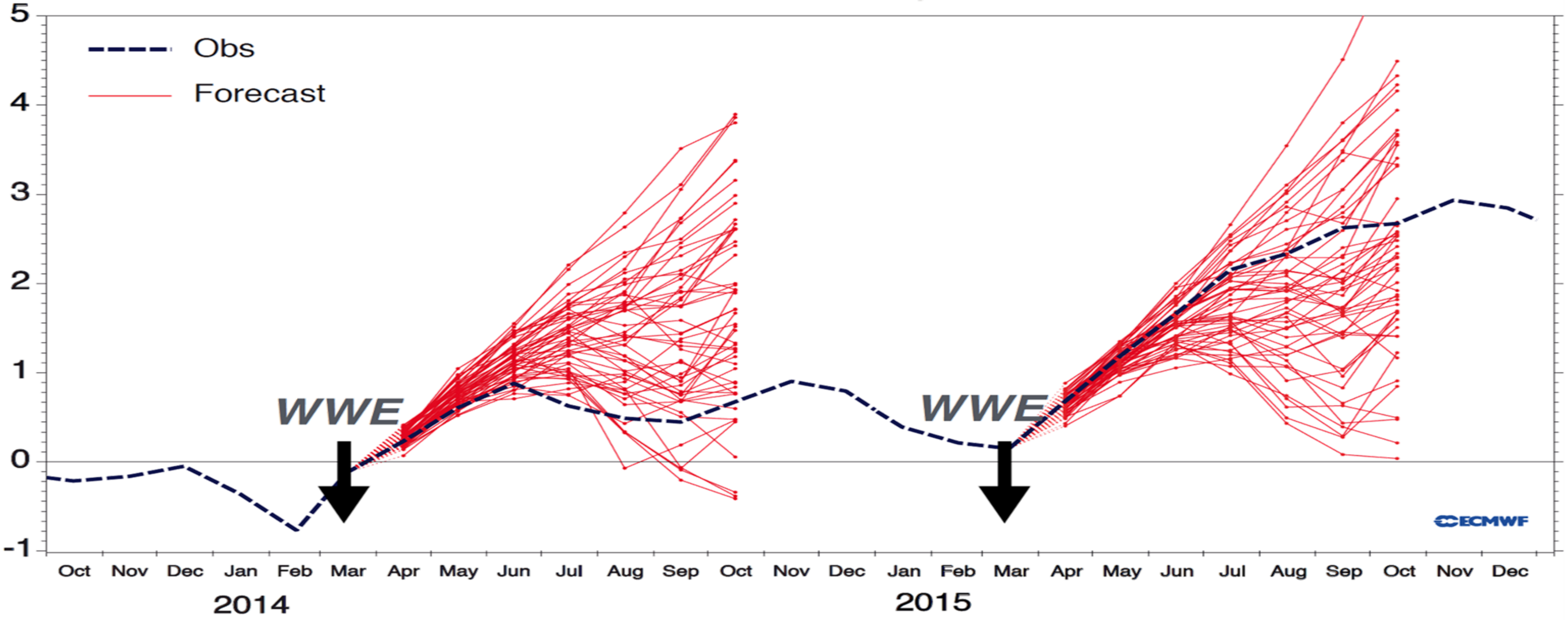
Predictive skill in the Tropics: "La Nada 2014"

OFF!





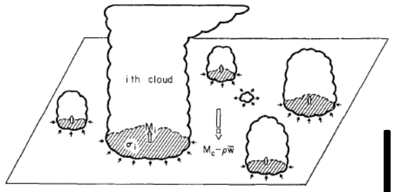
ON!

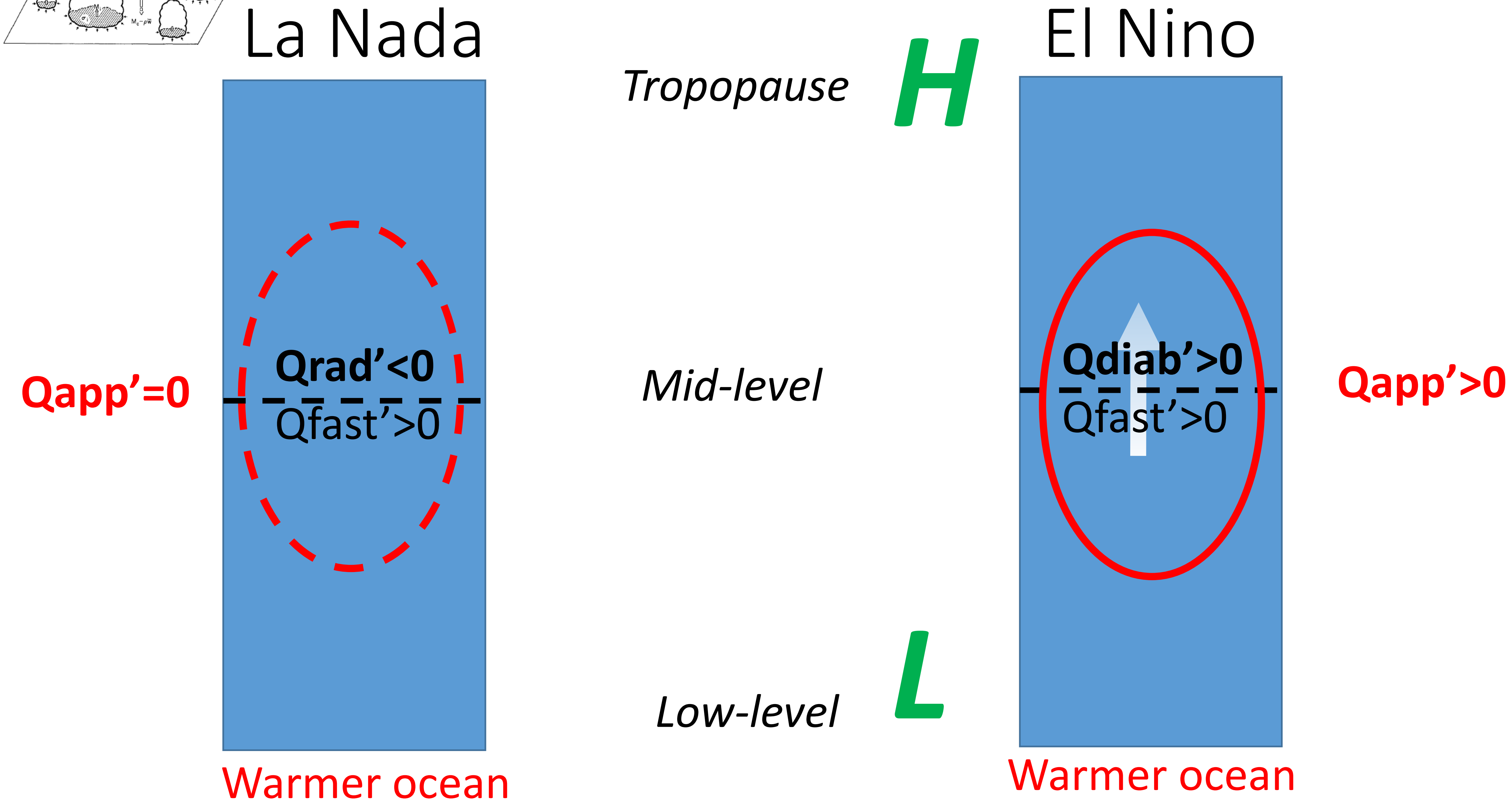


(slide courtesy of J. Vialard)

See McPhaden (2015)

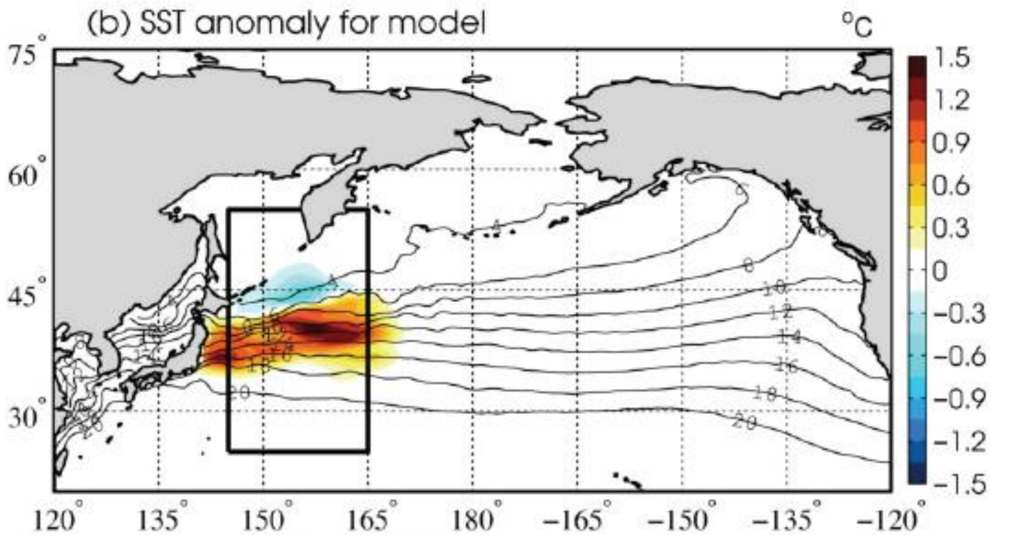
$$< Q_{app} > \approx < Q_{rad} > + < Q_{sen} > + M_c \frac{\partial < \theta >}{\partial z} - l_v e$$





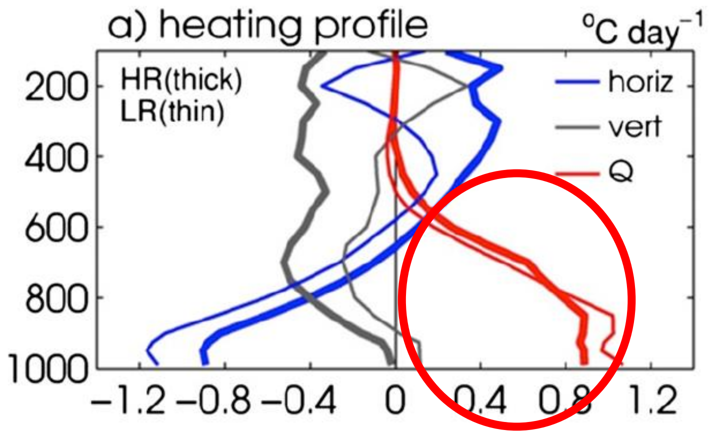
The Smirnov et al. (2015) experiments

HR = ¼ deg   LR = 1 deg

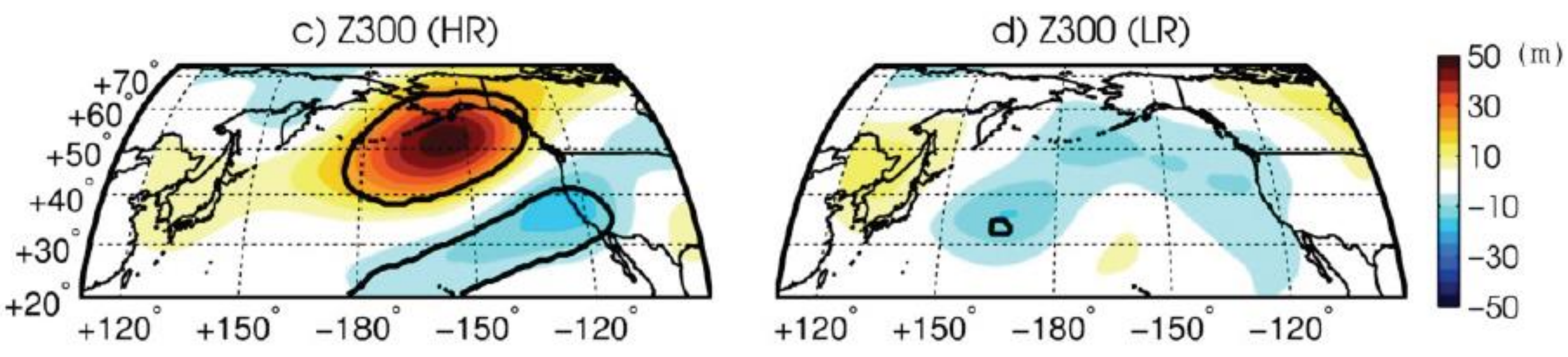


- Comparison of the response of an AGCM to an SST anomaly at two different resolution (~high and low res)
- A very similar diabatic heating is produced with two completely different circulation responses!

Heat budget components averaged over the box (35-43N)



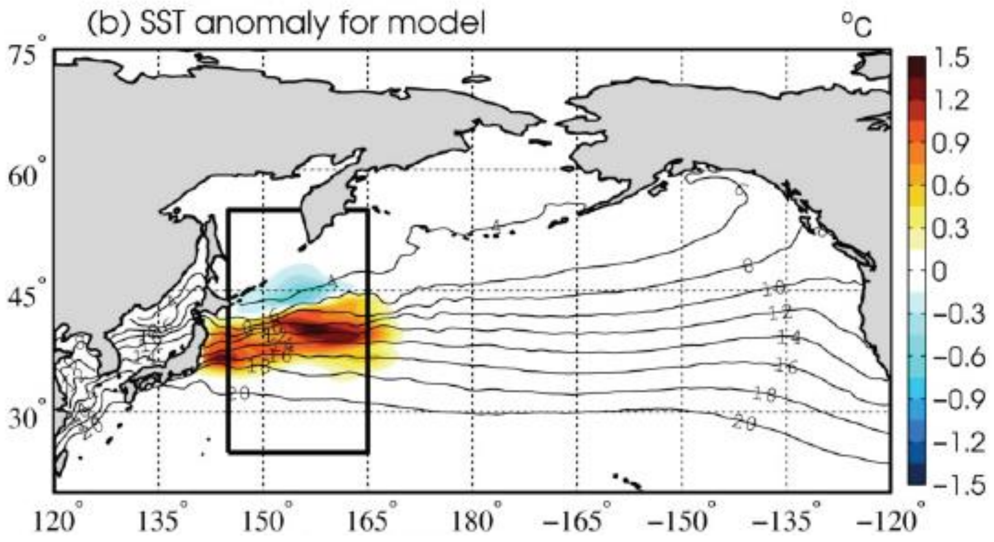
Upper level height anomalies (300hPa)



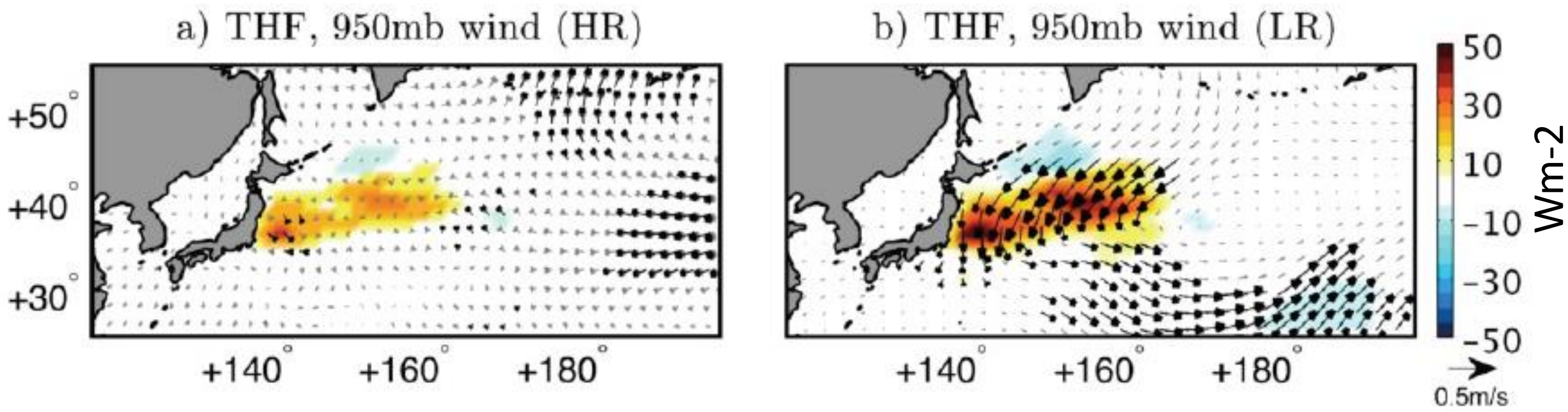
…the "Quantum Café"

HR = ¼ deg   LR = 1 deg

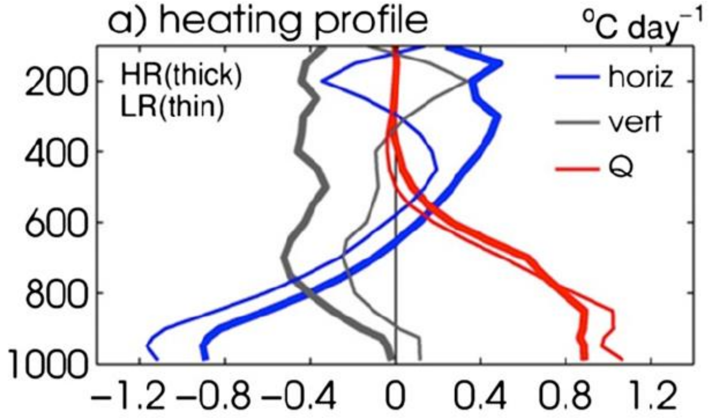
- Comparison of the response of an AGCM to an SST anomaly at two different resolution (~high and low res)
- Completely different responses despite the same Qdiab



Surface winds and turbulent heat flux anomalies



Heat budget components averaged over the box (35-43N)

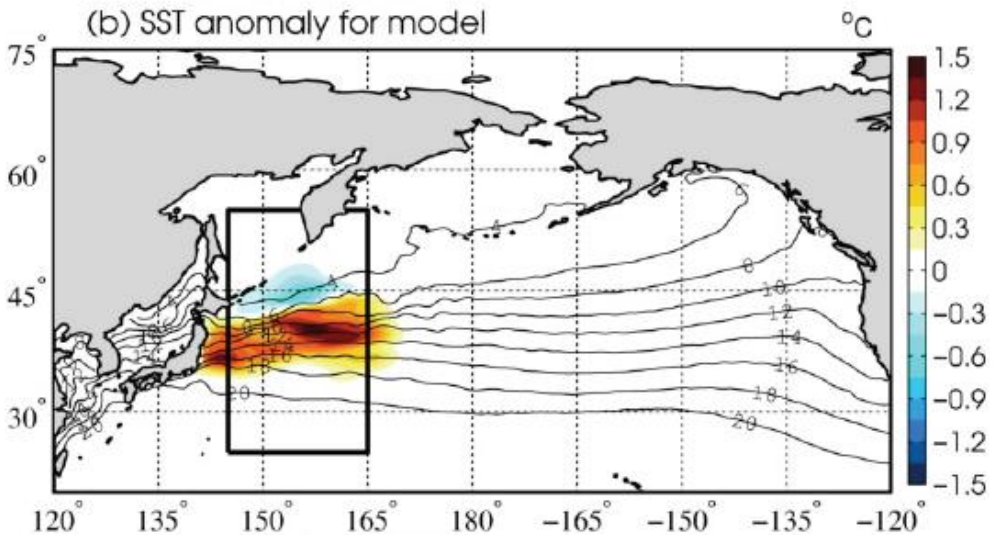


Smirnov et al. (2015)

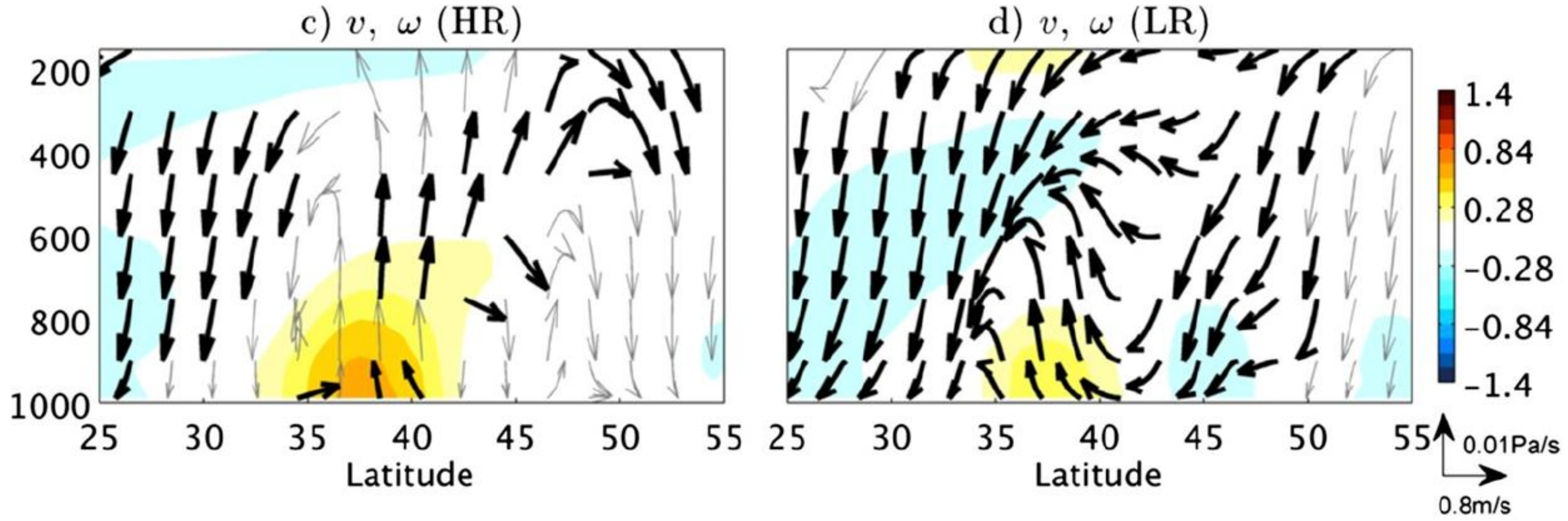
HR = ¼ deg LR = 1 deg

…the “Quantum Café”

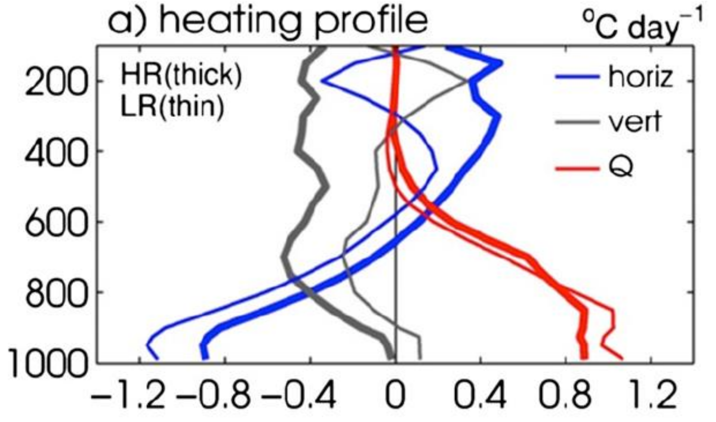
- Comparison of the response of an AGCM to an SST anomaly at two different resolution (~high and low res)
- Completely different responses despite the same Qdiab



Zonally averaged meridional circulation in the box (color = moist θ perturbation)



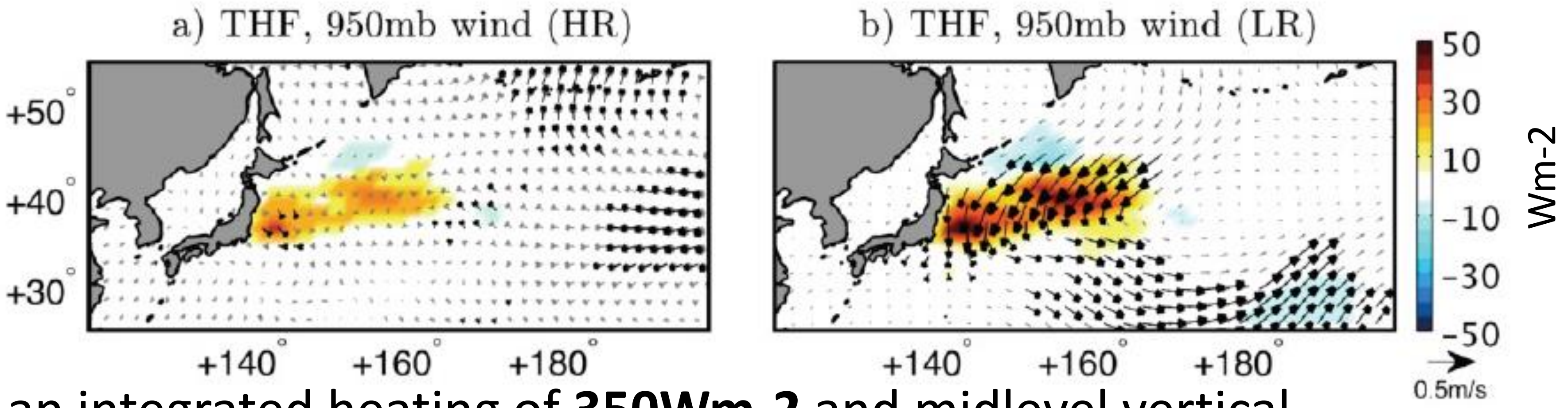
Heat budget components averaged over the box (35-43N)



Smirnov et al. (2015)

Numbers

Surface turbulent heat flux anomaly (color) and surface wind vectors



- Hendon & Hartmann had an integrated heating of **350Wm-2** and midlevel vertical velocities of:

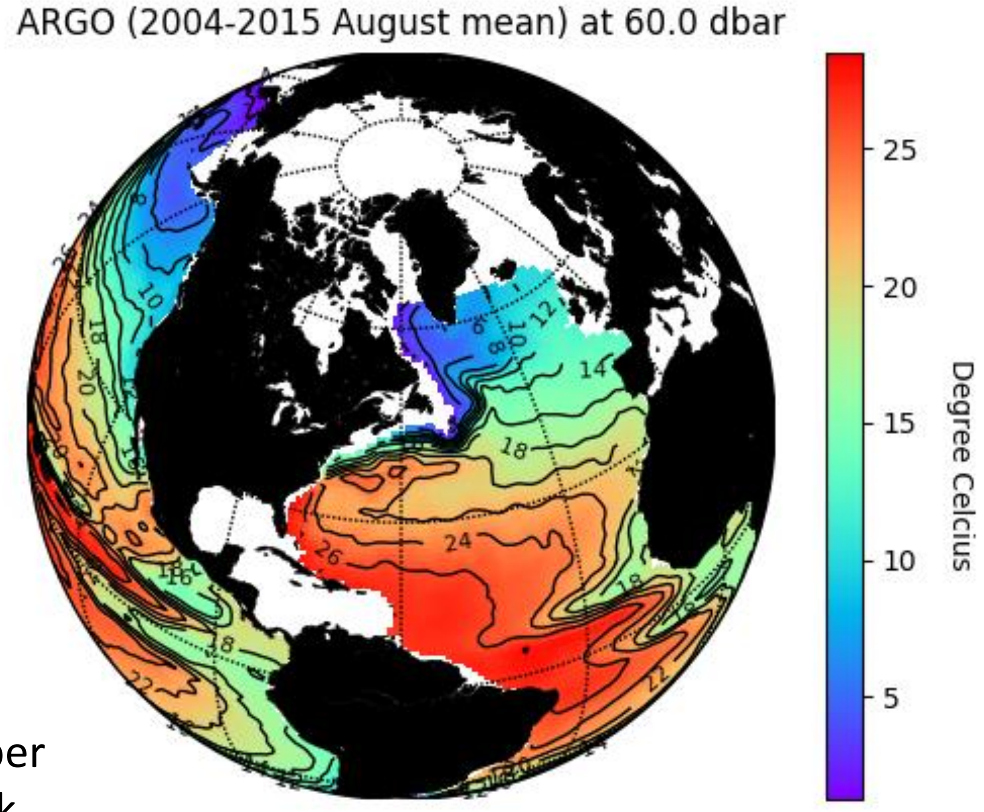
ω = -4 mb/hr = -11.2 X 0.01 Pa/s (deep heating at 15N)
ω = -0.8 mb/hr = -2.2 X 0.01 Pa/s (deep heating at 46N)
ω = +1.2 mb/hr = +3.4 X 0.01 Pa/s (shallow heating at 46N)

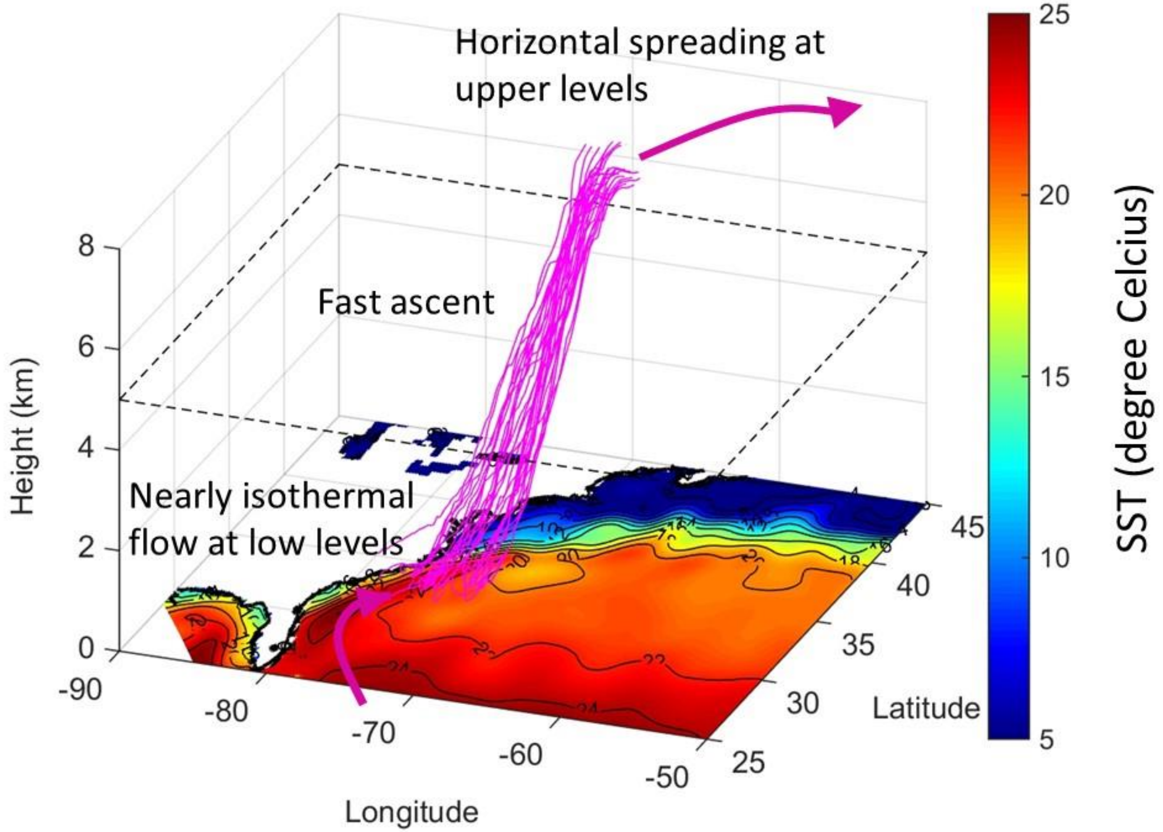
- Smirnov et al. produce, for a surface heat flux anomaly of **20Wm-2**, a midlevel vertical velocity of:

ω = - 0.01 Pa/s (40N SST anomaly in HR)

Thus the HR experiment generates midlevel upward motion of similar magnitude as subtropical forcing

Ascent in weather fronts is sensitive to the Gulf Stream (Sheldon et al., 2017)





Remember Bob's talk yesterday!

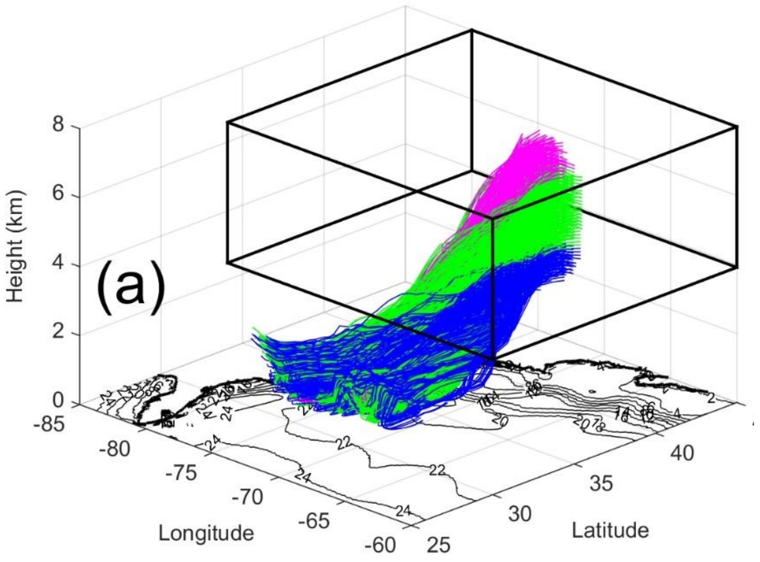
Ascent in weather fronts is sensitive to the Gulf Stream (Sheldon et al., 2017)

zi>7km

5km< zi < 7km

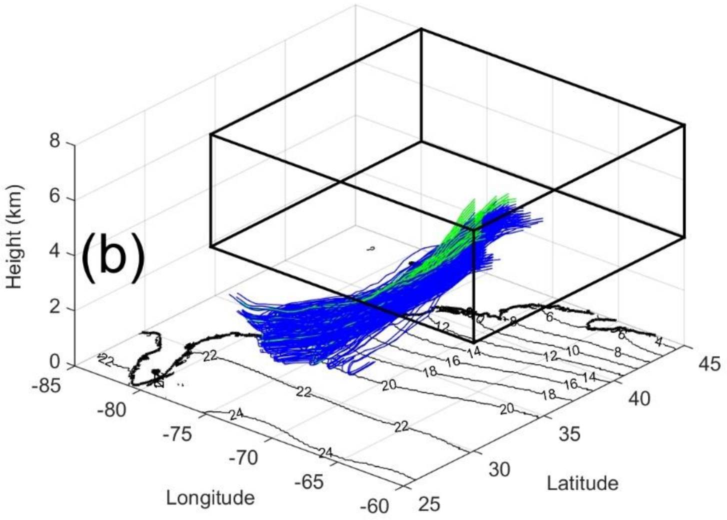
zi<5km

CNTL

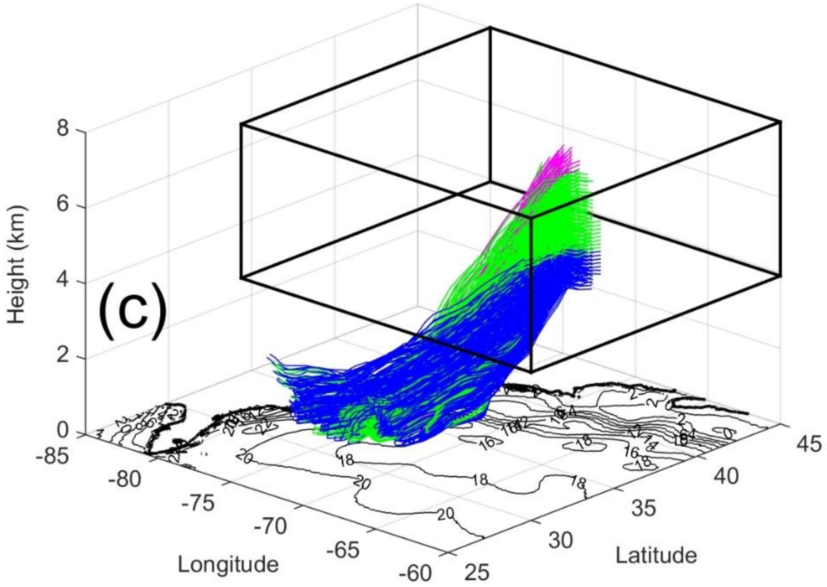


Initial release volume

SMTH



COOL



Back trajectories originating from low levels in *12km resolution runs*

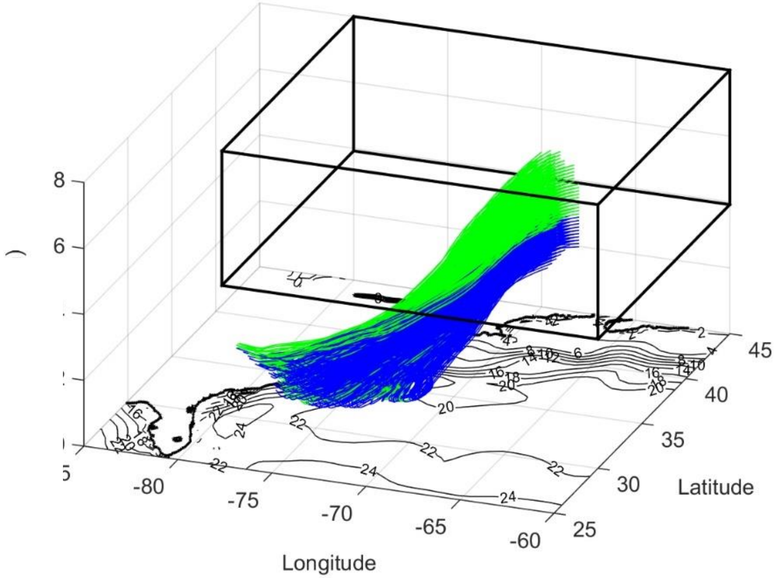
Ascent in weather fronts is sensitive to the Gulf Stream (Sheldon et al, 2017)

zi>7km

5km< zi < 7km

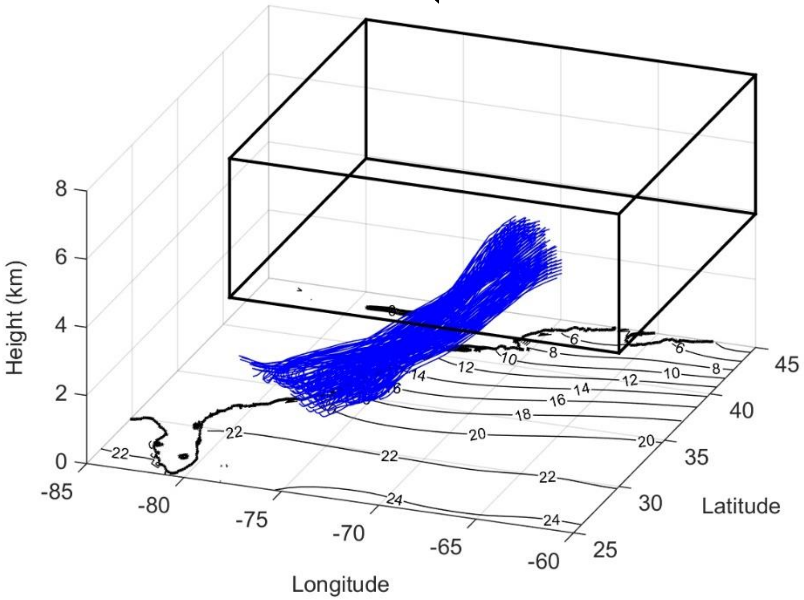
zi<5km

CNTL

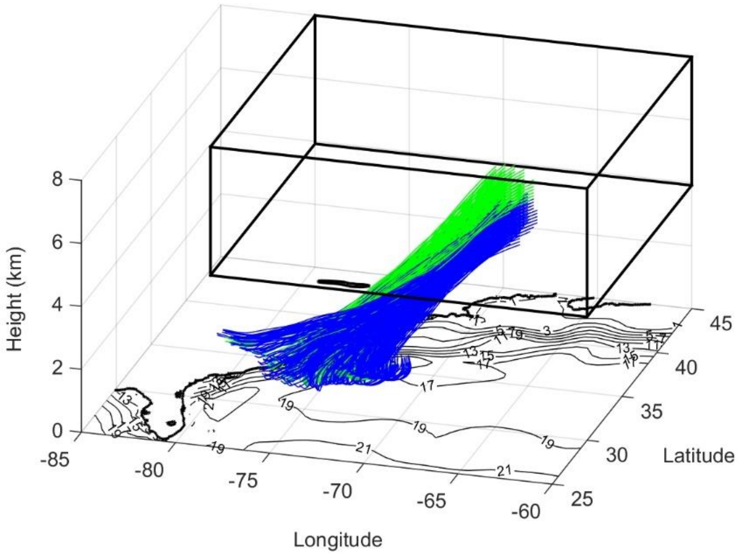


Initial release volume

SMTH



COOL



Back trajectories originating from low levels in *__40km resolution runs__*

Summary

- The observational record shows well defined “patterns” of co-variability between the ocean and atmosphere on monthly, interannual and possibly decadal timescales. There is a rich range of behaviour between basins in terms of temporal signatures and timescales.
- The standard theory of thermal forcing provides insight into how a warmer/colder upper ocean affects wind and temperature distribution but one must acknowledge the complexity of the link SST → diabatic heating
- There is suggestion that the midlatitude oceanic forcing increases with spatial resolution and might become comparable in magnitude to the low latitude oceanic forcing