Mixed Layer Model Assessment of Boundary Layer Convergence and the MJO

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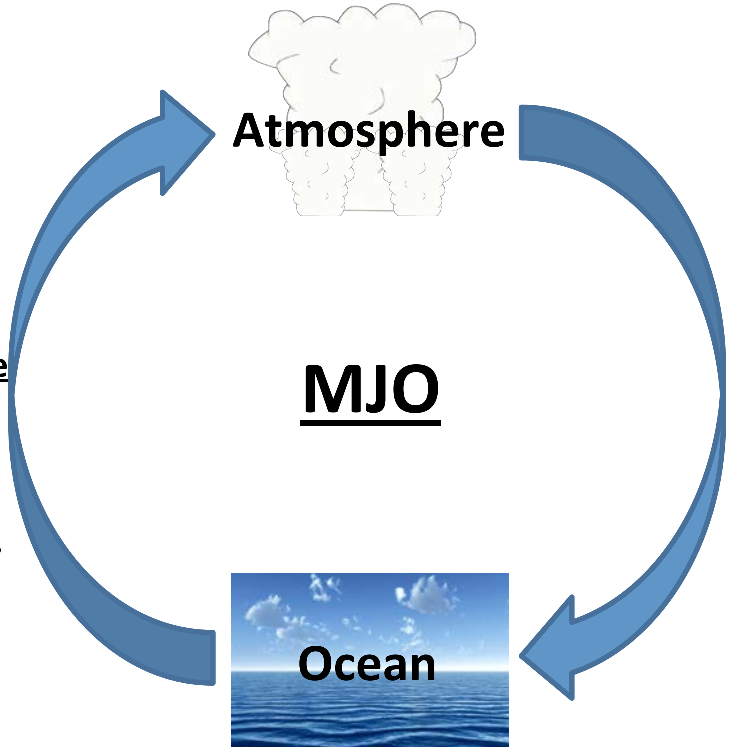
Photo: Walter Hannah

Ocean-Atmosphere Feedbacks and the MJO



Oceanic Forcing of Atmosphere

Latent and Sensible Heat Flux
Momentum
Chemical and Biological Fluxes

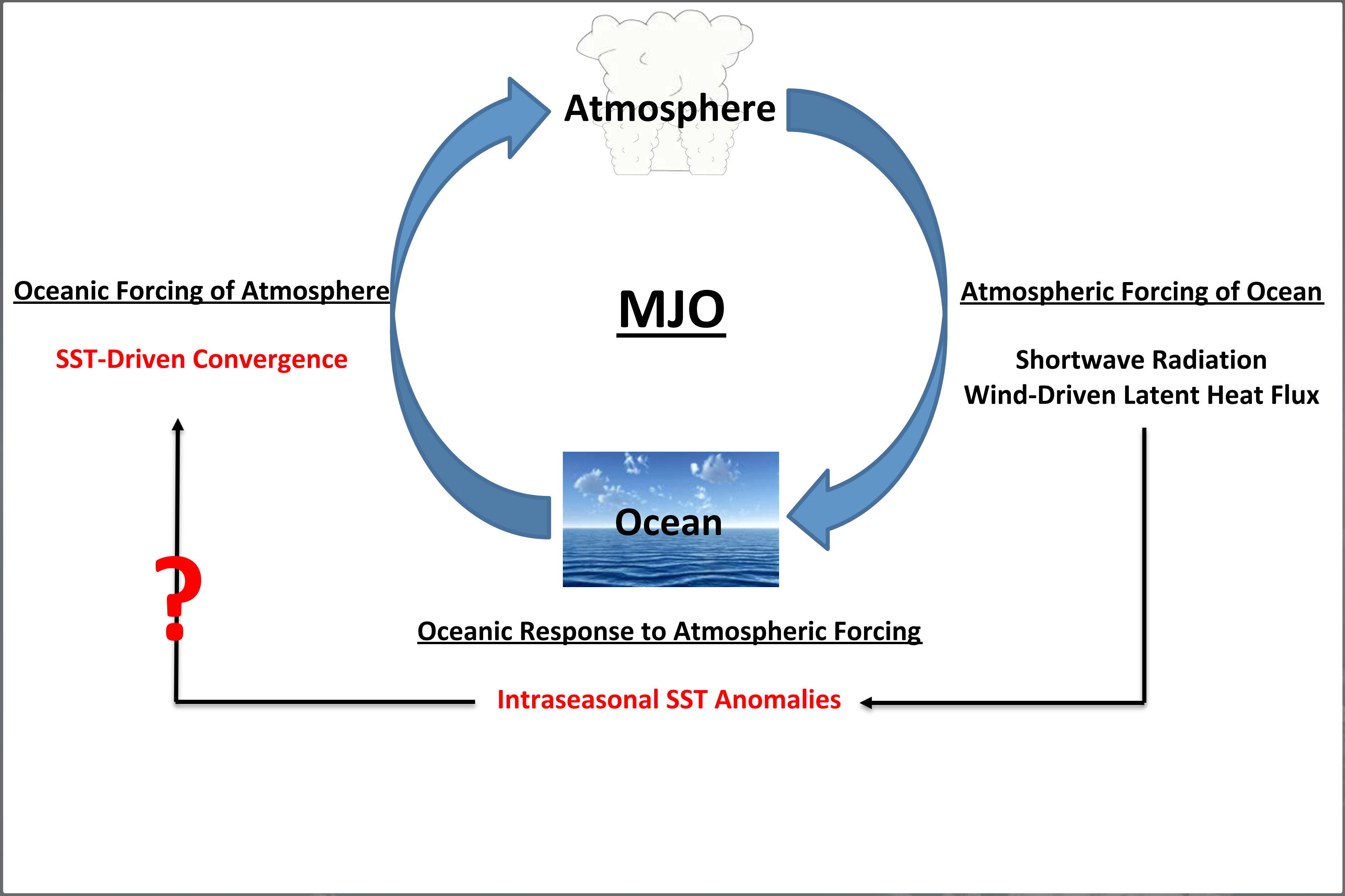
Atmospheric Forcing of Ocean

Radiation
Latent and Sensible Heat Flux
Precipitation
Momentum

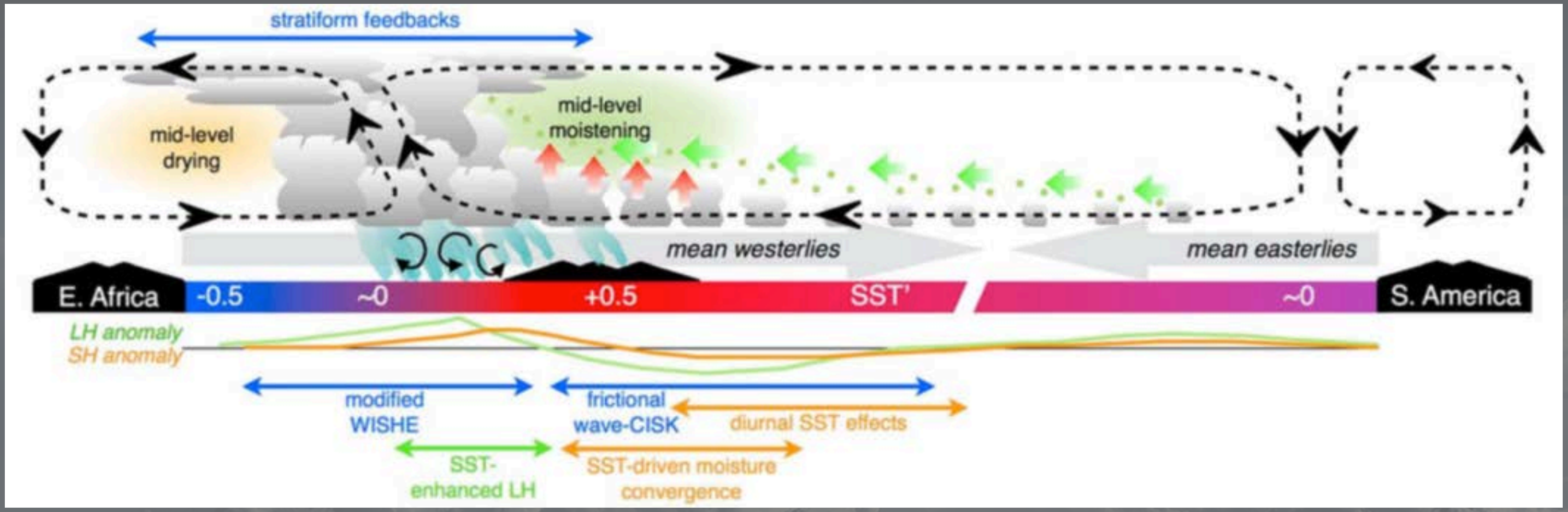
Oceanic Response to Atmospheric Forcing

Heat Content
Stratification
Dynamics
Chemistry and Biology

Ocean-Atmosphere Feedbacks and the MJO



Ocean-Atmosphere Feedbacks and the MJO



DeMott et al. 2015

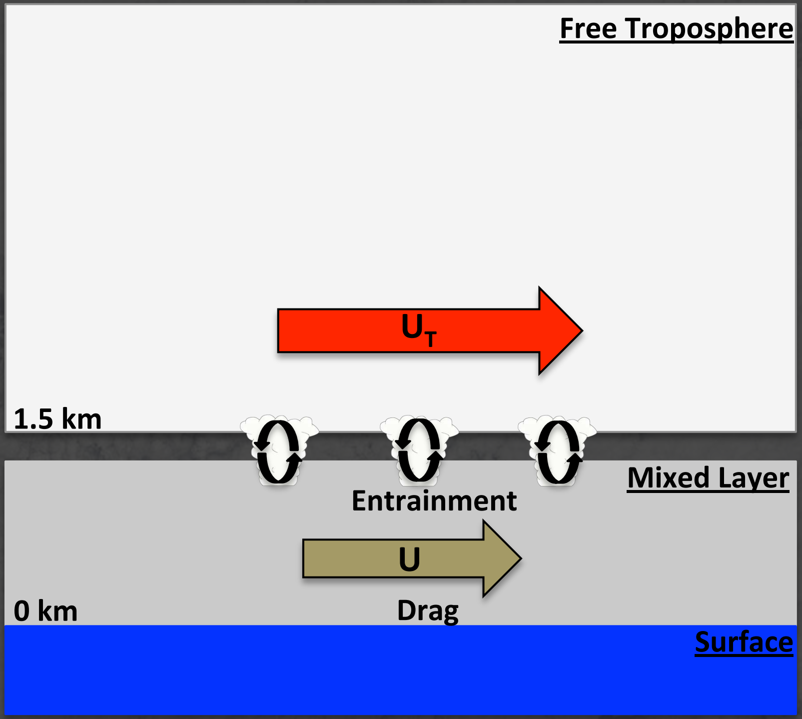
Do intraseasonal SST anomalies drive considerable boundary layer convergence?

Mixed Layer Model (MLM)

MLM of Stevens et al. (2002) used by Back and Bretherton (2009)

Assumes force balance in mixed layer of constant depth h

Mixed layer capped by trade inversion where shallow convection communicates between mixed layer and free troposphere

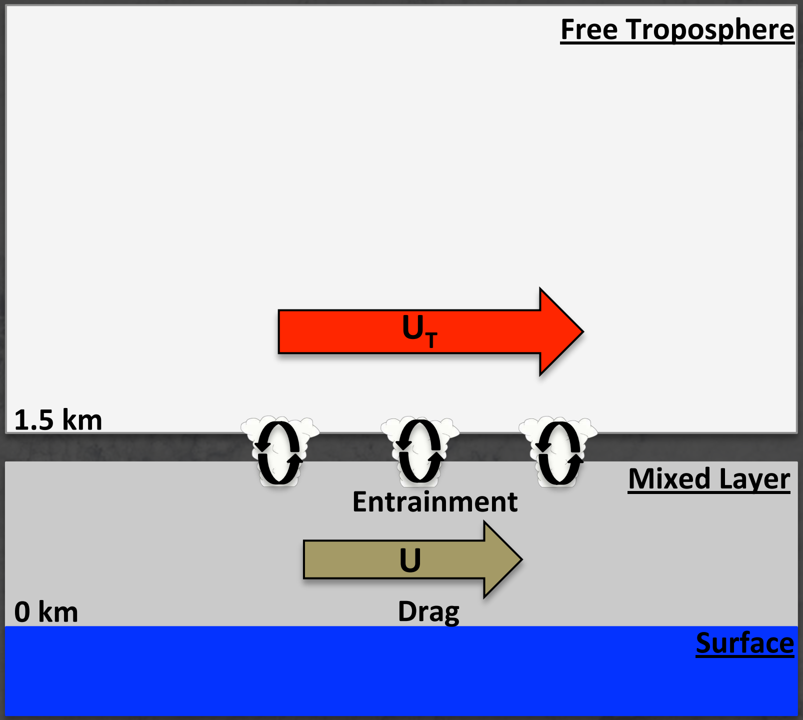


Mixed Layer Model (MLM)

Steady-State BL Force Balance

$$f\mathbf{k} \times \mathbf{U} + \rho_0^{-1}\nabla P_s = \frac{w_e(\mathbf{U}_T - \mathbf{U})}{h} - \frac{w_d(\mathbf{U})}{h}$$

Coriolis Acceleration | Pressure Gradient | Downward Momentum Mixing | Friction



Inputs to Standard MLM

U and V at 850 hPa

P_s

Mixed Layer Model (MLM)

Entrainment | Pressure Gradient

$$U = \frac{U_T \epsilon_i \epsilon_e + V_T f \epsilon_e - \rho_0^{-1}(f\partial P_s/\partial y + \epsilon_i \partial P_s/\partial x)}{\epsilon_i^2 + f^2}$$

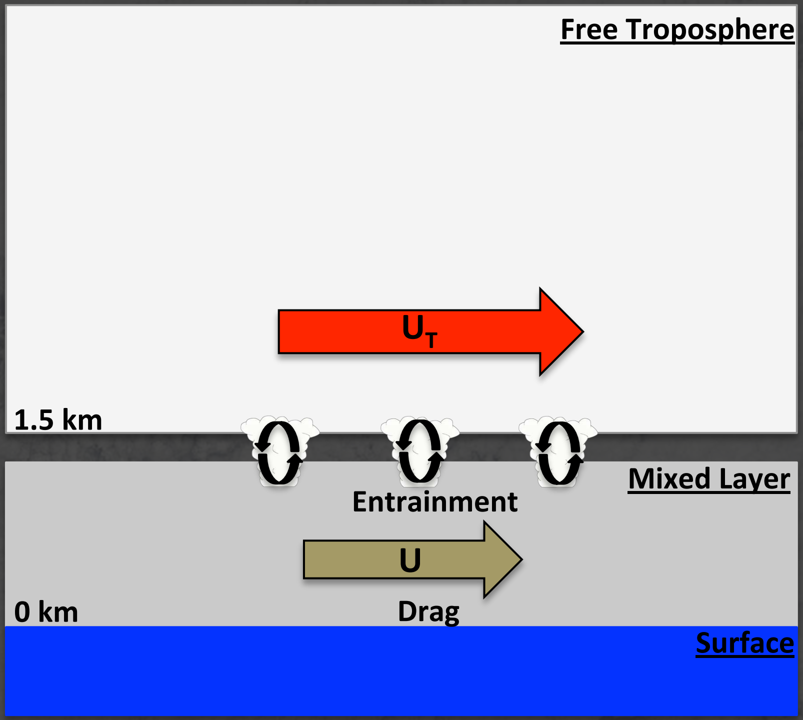
$$V = \frac{V_T \epsilon_i \epsilon_e - U_T f \epsilon_e + \rho_0^{-1}(f\partial P_s/\partial x - \epsilon_i \partial P_s/\partial y)}{\epsilon_i^2 + f^2}$$

Both entrainment and pressure gradient are modified by drag and Coriolis Acceleration.

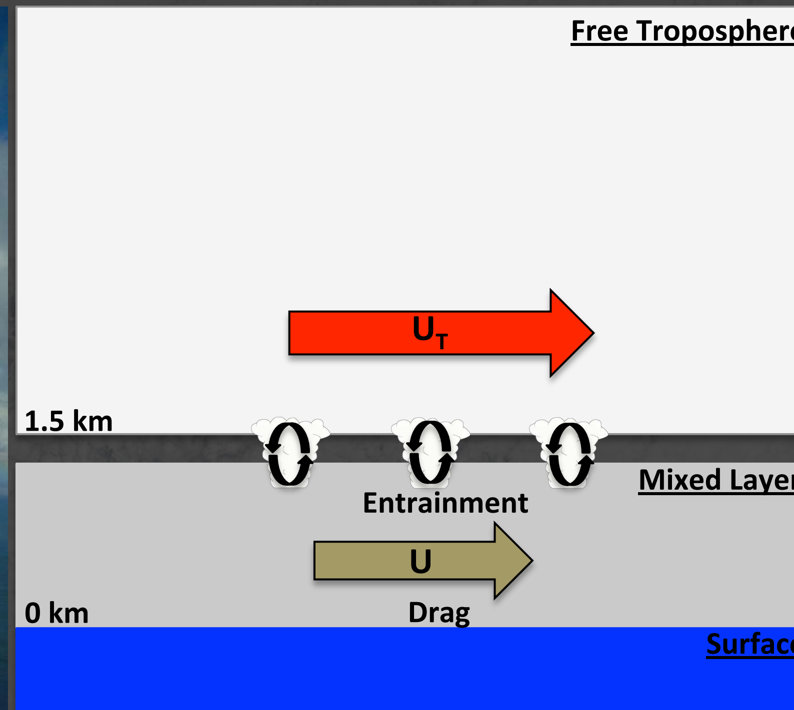
E_e = entrainment timescale (w_e / h from previous slide)

E_i = entrainment timescale plus drag timescale (w_e / h + w_d / h from previous slide)

E_e and E_i are our tunable parameters, left at standard values from Back and Bretherton 2009



MLM vs Reality



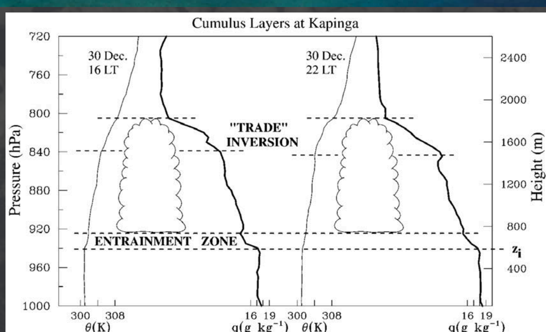


FIG. 3. Example of soundings (potential temperature θ in thin solid lines and specific humidity q in heavy solid lines) at Kapinga at 1600 and 2200 LST 30 Dec 1992 depicting a period when the mixed layer was topped with a shallow cumulus layer. Also indicated are the entrainment zone atop the mixed layer, and a "trade-type" inversion layer between 840 and 800 hPa.

Johnson et al. 2001

Mixed Layer (ML) = properties well mixed

Boundary Layer (BL) = layer up to trade inversion that is directly influenced by surface. Because drag is felt up to this level, WTG balance does not hold in BL. It is more difficult to establish horizontal temperature gradients above the BL.

Data and Methodology

Inputs to MLM

ERAi → Geo_{850} U_{850} V_{850} T_{850} $P_{surface}$

NOAA OISST (AVHRR + AMSR) → SST

Validation

Cross-Calibrated Multi-Platform (CCMP) V2 Wind Analysis → $U_{10\text{-}meter}$ $V_{10\text{-}meter}$

ERAi → $Omega_{850}$ / 150 hPa

Composites

Filtered MJO OLR (FMO) Index

TRMM 3B42 Precipitation

All data is:

2002-2011

0.75 x 0.75

Data and Methodology

Validation

$$FC = \frac{\iiint M(x,y,t)V(x,y,t)\partial x \partial y \partial t}{\iiint V(x,y,t)^2 \partial x \partial y \partial t}$$

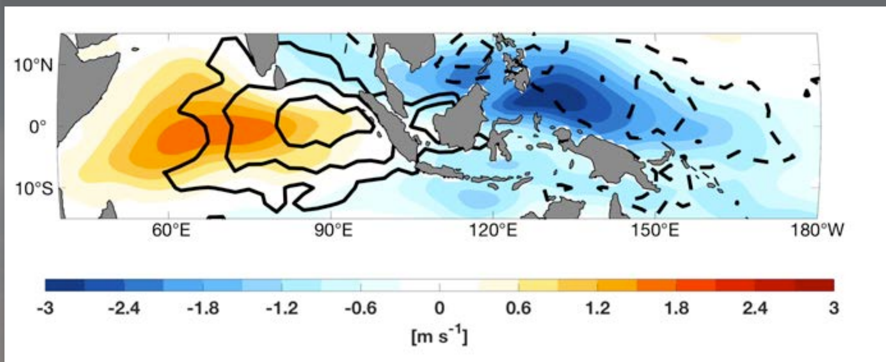
dx = 40E to 180
dy = 15N to 15S
dt = FMO MJO lifecycle

The fractional contribution (FC) of some model field (M) to the observed variable (V) is equal to the covariance of M and V integrated over the domain and over a given time period (the numerator) divided by the variance of the observed variable V integrated over the same domain and time period.

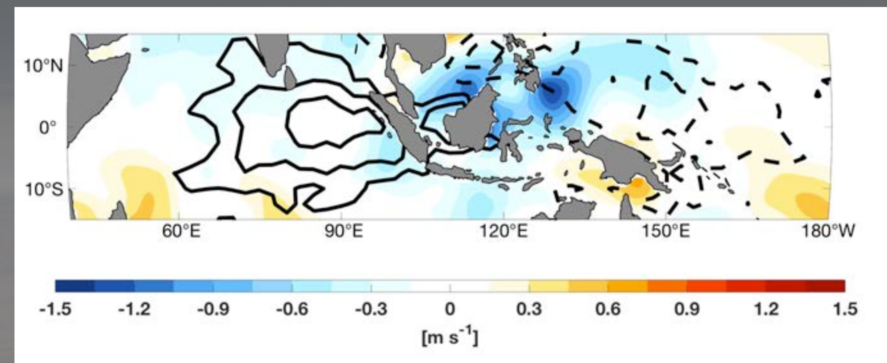
Essentially assessing spatial correlation, temporal correlation, and magnitude all in one analysis.

Full MLM Results

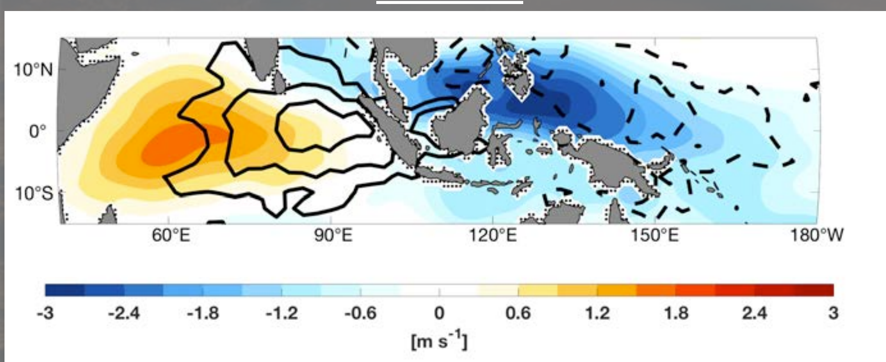
CCMP U 10m



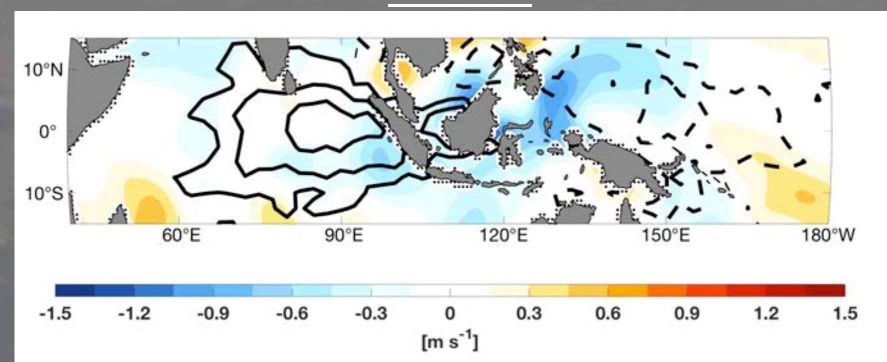
CCMP V 10m



U MLM



V MLM

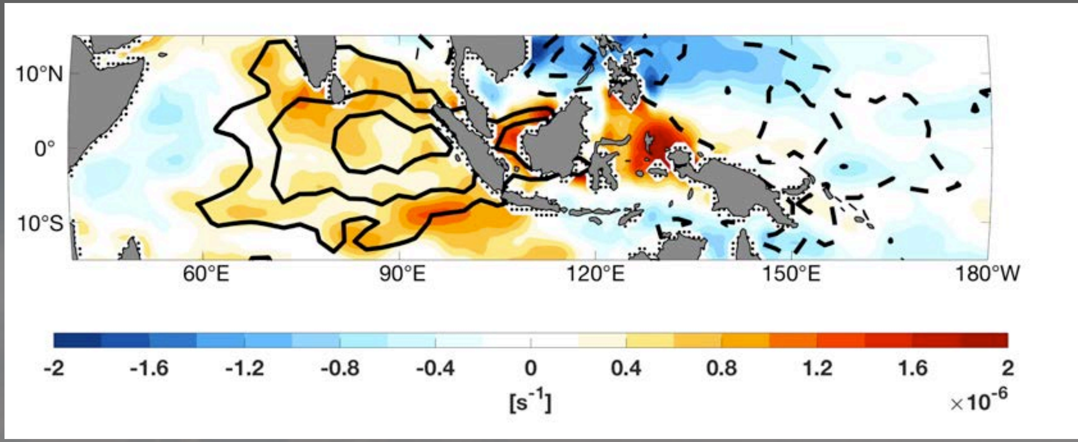


FC = 1.01

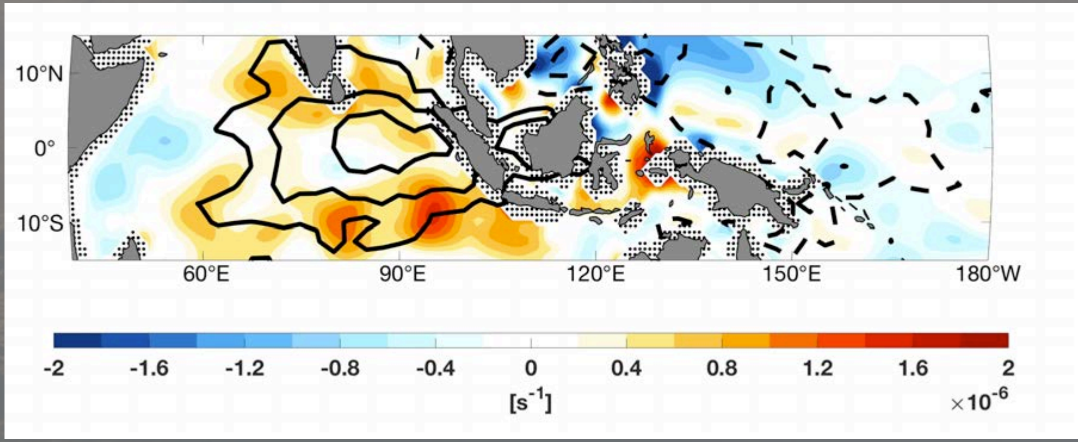
FC = 0.90

Full MLM Results

CCMP 10m Convergence



Full MLM Convergence

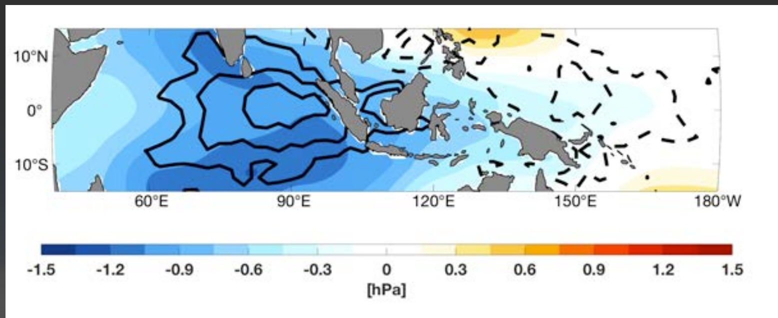


FC = 0.70

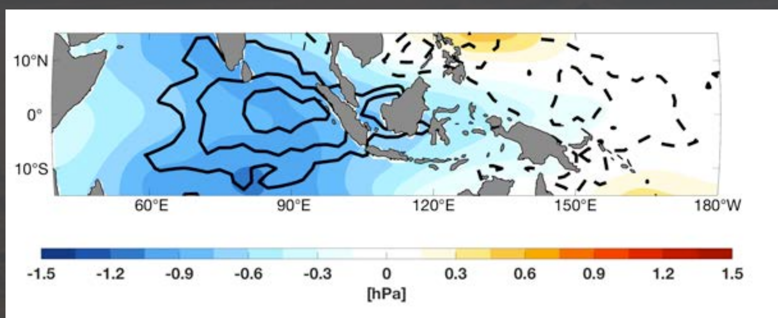
(FC = 0.75 for layer convergence)

Separating BL and Free Tropospheric Contributions

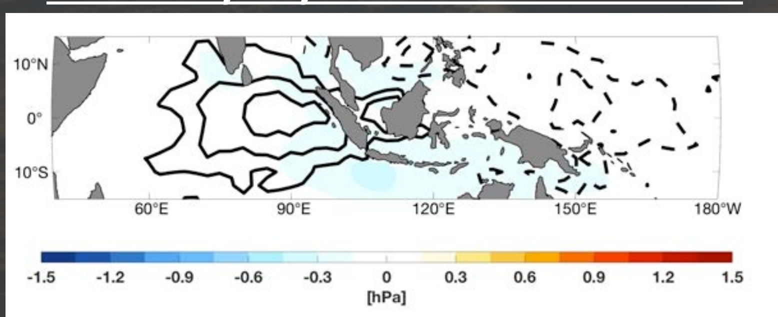
Surface Pressure

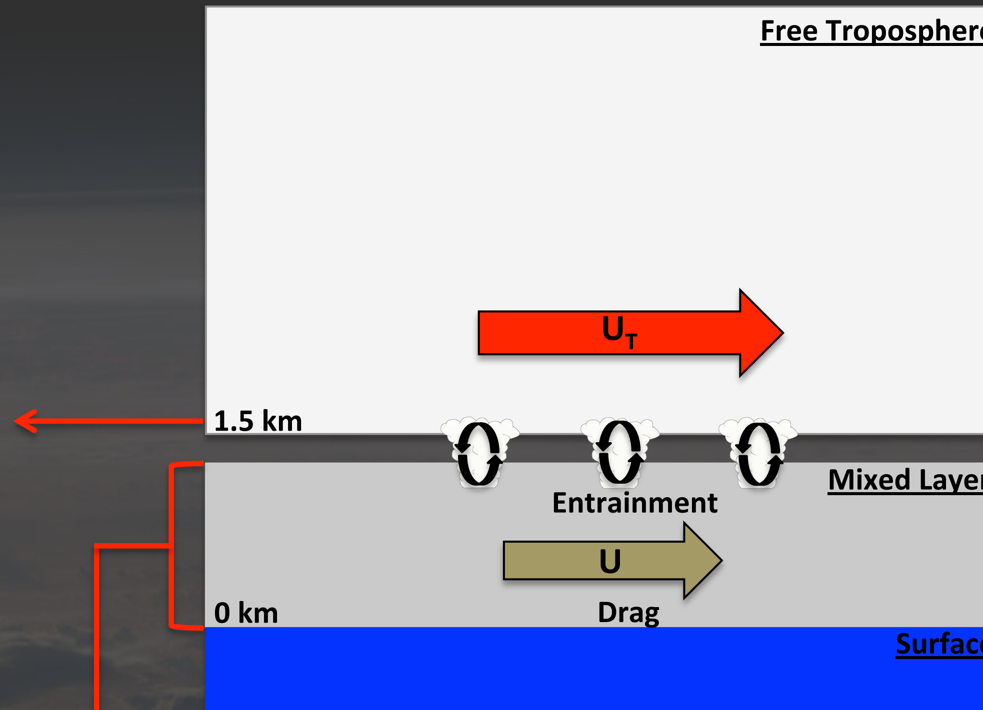


Pressure at 1.5 KM



Boundary Layer Pressure Thickness





$$P_s = P_i + \Delta P_{BL} \quad \text{and} \tag{3}$$

$$P_i = 850\,\text{hPa} + \rho_{850}(\Phi_{850} - \overline{\Phi}_{850}). \tag{4}$$

Here, P_i, which is approximately the pressure at the mean height of the 850-hPa surface, is calculated from zonally smoothed ERA-40 output. The boundary layer contribution ΔP_{BL}, calculated as a residual, is proportional to the mean temperature between the surface and 850 hPa.

Separating BL and Free Tropospheric Contributions

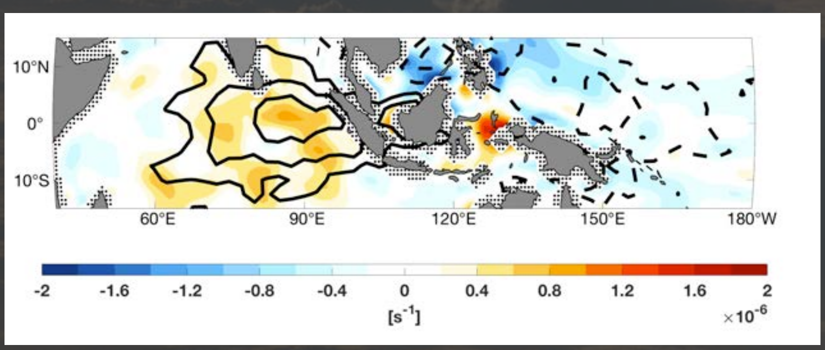
		Zonal Wind	Meridional Wind	Surface Convergence	Layer Average Convergence
		U10m	V10m	10m Convergence	-(850 hPa Omega)/150hPa
Free Tropo. + B.L. →	MLM	1.01	0.90	0.70	0.75
Free Troposphere →	MLMDEEP	0.96	0.69	0.43	0.52
Boundary Layer →	MLMBL	0.05	0.21	0.27	0.23

Zonal wind driven by free troposphere

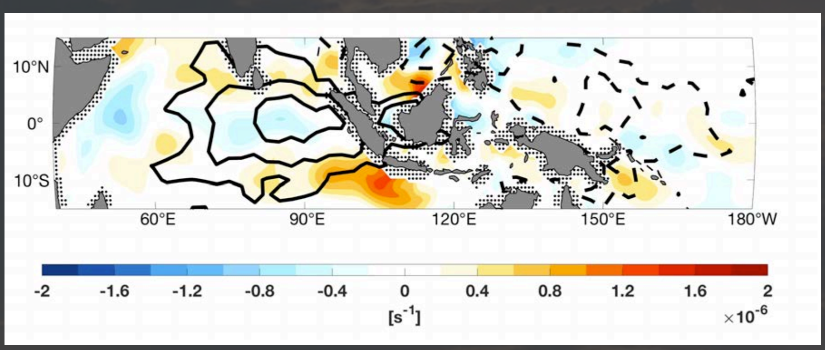
Meridional wind driven mostly by free troposphere, but BL contributes

Boundary layer convergence driven both by free troposphere (2/3) and boundary layer (1/3)

MLM Conv: Free Troposphere

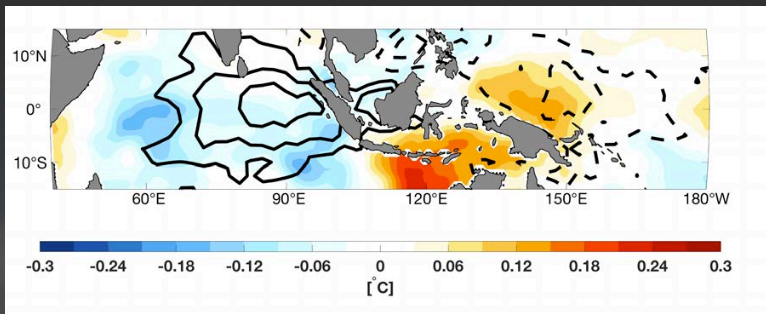


MLM Conv: Boundary Layer

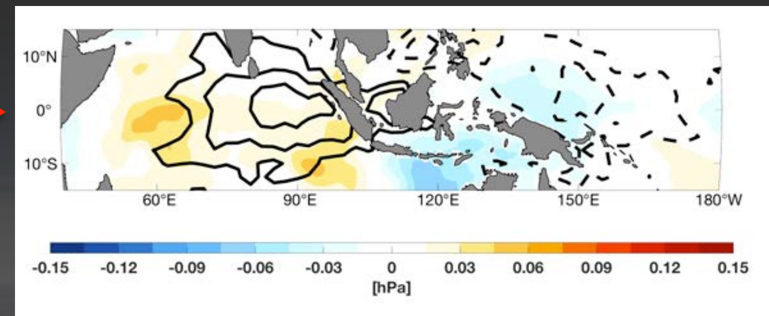


SST Contributions to Boundary Layer Pressure, Winds, and Convergence

NOAA OI SST



Estimated SST Driven Pressure Anomalies

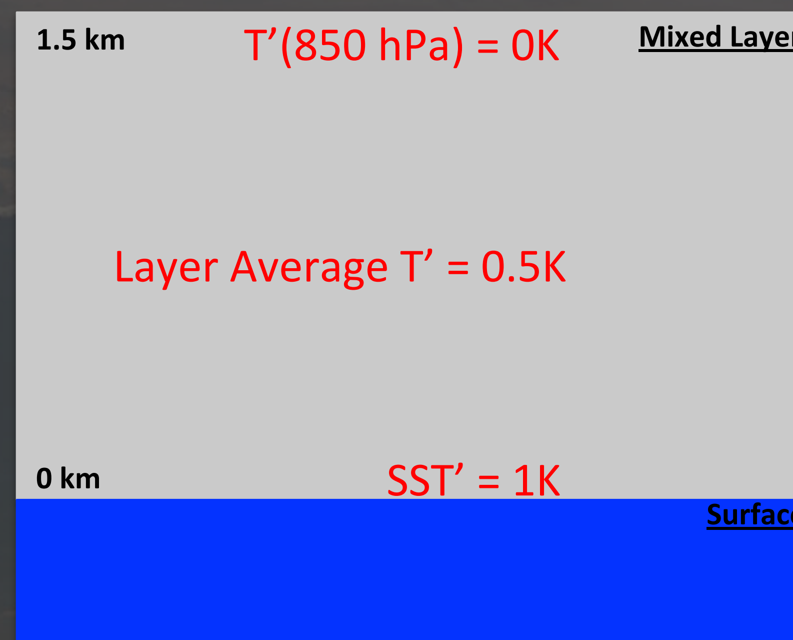


Assume air temperature varies linearly between SST’ and T’@850 hPa

Linearize density around mean temperature

Integrate through depth of boundary layer

These pressure anomalies are very small!



SST Contributions to Boundary Layer Pressure, Winds, and Convergence

		BL Pressure	Zonal Wind	Meridional Wind	Surface Convergence
		Delta P_{BL}	Residual U10m	Residual V10m	Residual 10m Convergence
SST + T850 →	MLM (SST + T850)	0.21	0.10	0.07	0.05
T850 only →	MLM (T850)	0.19	0.06	0.05	-0.01
SST only →	MLM (SST)	0.03	0.04	0.02	0.03

Linear combination of SST and T@850 hPa only captures 20% of pressure anomalies caused by BL temperature anomalies

Almost all of this 20% is contributed by downward mixing of T@850 hPa

SST anomalies contribute little to boundary layer temperature anomalies

Processes other than influence of SST and downward mixing of T@850 hPa dominate BL temperature budget

What Do These Results Tell Us?

Repeated Analysis

1. Conditioned on enhanced and suppressed phases separately
2. Full ERAi time period (1979 – 2016)
3. Composites of selected “Large MJO Event”

The following conclusions are robust across these analyses

What Do These Results Tell Us?

Conclusions

1. Basin scale MJO boundary layer winds and convergence are primarily driven by free tropospheric processes communicated downward
2. Boundary layer temperature anomalies play a lesser but still important role in driving MJO boundary layer convergence
3. No evidence of "large" scale SST anomalies (~0.5 degrees C) playing a major role in determining boundary layer pressure anomalies, winds or convergence in a consistent/coherent manner
4. SST anomalies must be of larger magnitude and/or smaller spatial scale to play a first order role in driving boundary layer convergence

What Don't These Results Tell Us?

These results do not suggest that:

1. SST driven boundary layer convergence can not be important for individual MJO events
2. Smaller scale SST anomalies (e.g. diurnal warm layers) do not play a consistent/coherent role in modifying boundary layer convergence in most MJO events

For SST anomalies to drive first order convergence they must be:

1. Stronger

and/or

2. Smaller spatial scale

and

3.) O(1) in boundary layer temperature budget

SST Gradients and Their Influence on Boundary Layer Pressure Gradients

From BnB

Connection to SST

To construct an LN-like model, we must relate gradients in the boundary layer pressure contribution ΔP_{BL} to gradients in SST. A simple approximation is to hydrostatically estimate ΔP_{BL} by assuming the air temperature varies linearly between a surface value equal to the SST and the ERA-40 850-hPa temperature, which is roughly at the mean inversion height; density can then be linearized about the mean temperature and integrated (as in LN). Figure 5a shows the corresponding

From Lindzen Nigam

$$\rho = \rho_0[1 - n(T - T_0)], \qquad (2a)$$

where

$$\rho_0 = \rho(T_0) = 1.225\ \mathrm{kg\,m^{-3}}, \quad T_0 = 288\ \mathrm{K},$$

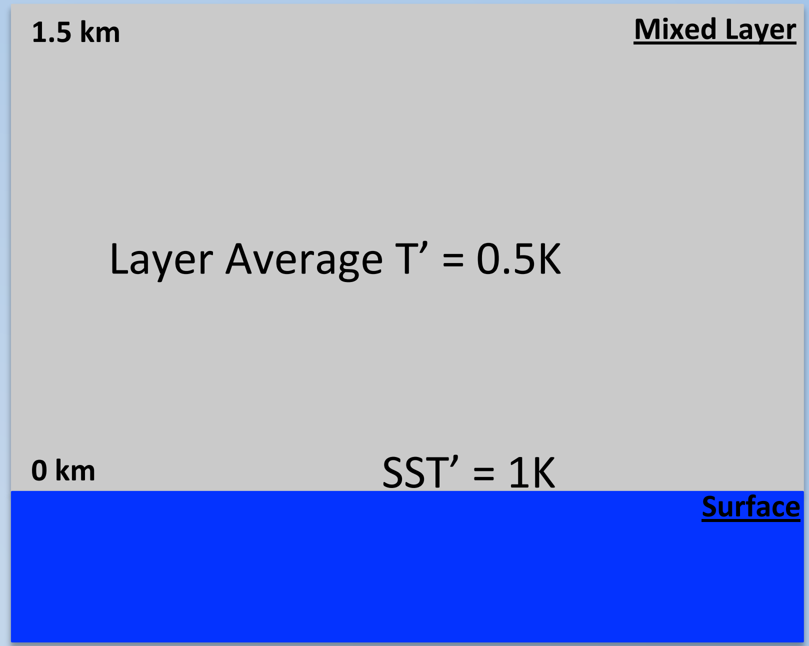
$$n = -\left[\frac{1}{\rho}\frac{\partial \rho}{\partial T}\right]_{T_0} = 1/T_0,$$

so that

$$\rho = \rho_0[2 - nT]. \qquad (2b)$$

From the knowledge of density and temperature fields, the three-dimensional pressure field can be constructed using the hydrostatic equation and a boundary condition or an integration constant; assuming the latter to be a specification of the geopotential height field

T'(850 hPa) = 0K



Intraseasonal Example

SST' = 1K

T'(850 hPa) = 0K

Assume linear variation of T →

Layer Average T' = 0.5K

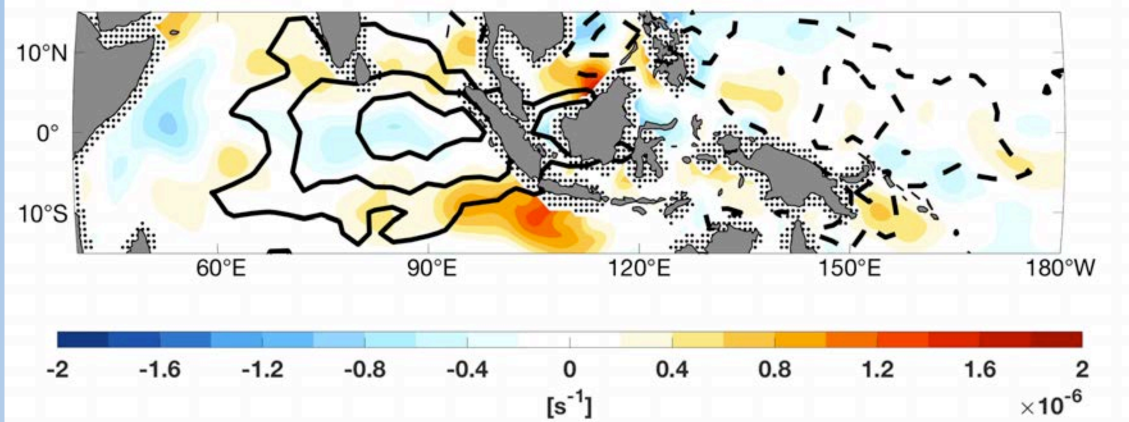
Linearize density about T, with reference T and density from LN above:

$\rho' = \rho_0[1-(T_0+T')/T_0] = 1.225\ \mathrm{kg\ m^{-3}}\ [1 - (288\mathrm{K}+0.5\mathrm{K})/288\mathrm{K}] = -0.0021\ \mathrm{kg\ m^{-3}}$

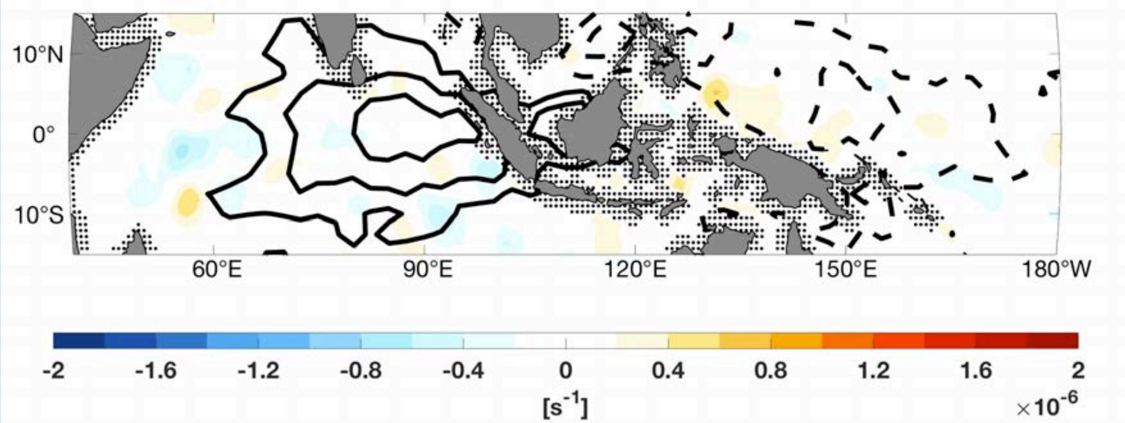
Use hydrostatic to integrate density over depth of mixed layer to get pressure:

$\Delta P' = \rho' * g * \Delta z$ → $\Delta P'_{BL} = \rho' * g * \Delta z_{BL} = -0.0021\ \mathrm{kg\ m^{-3}} * 9.8\ \mathrm{m\ s^{-2}} * 1500\ \mathrm{m} = -0.32\ \mathrm{hPa}$

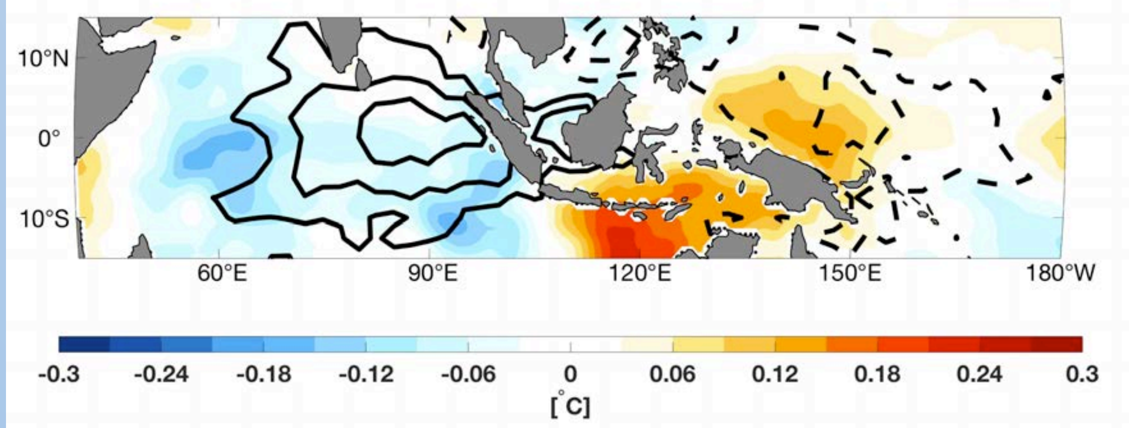
MLM BL



MLM – SST Only



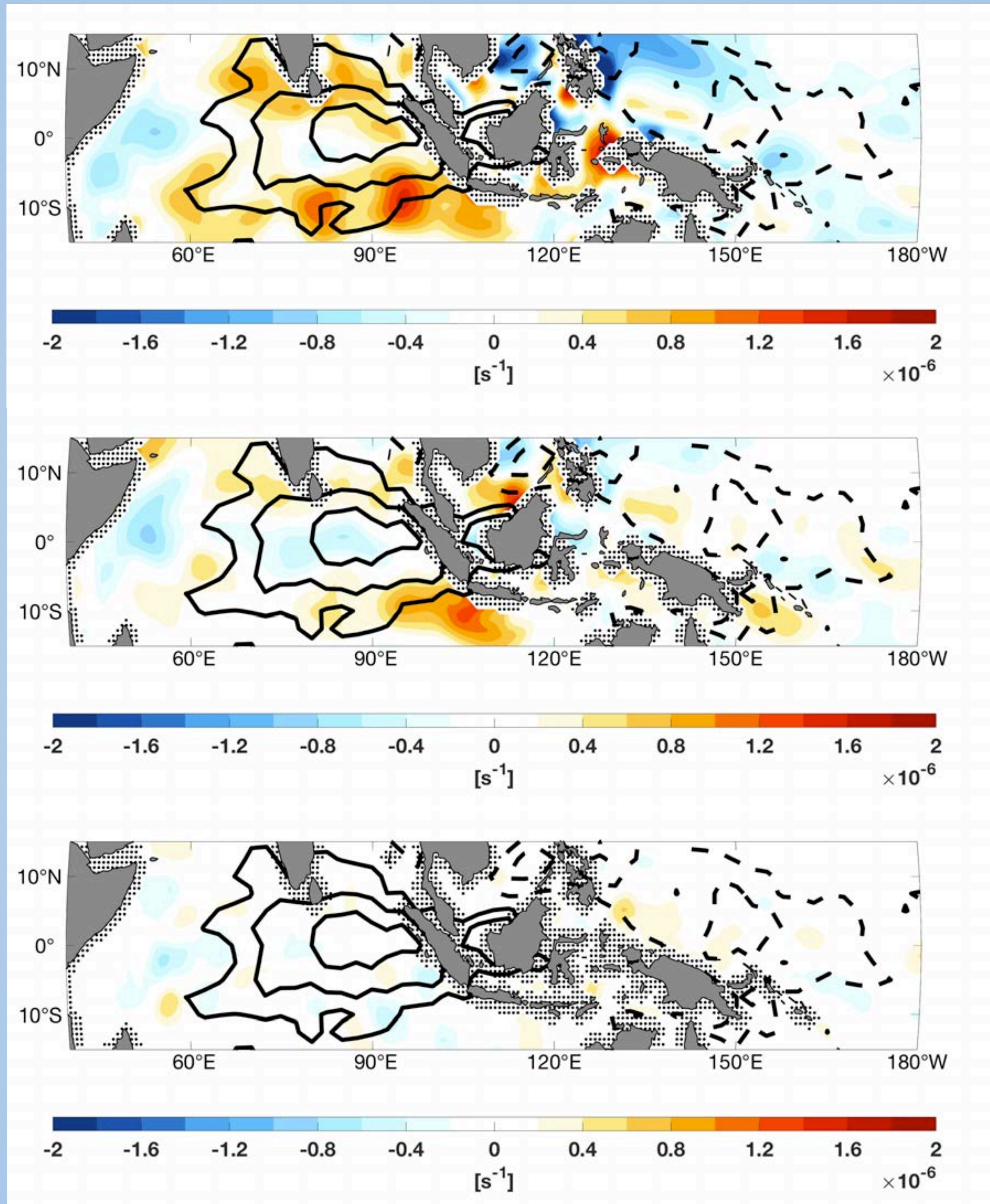
SST



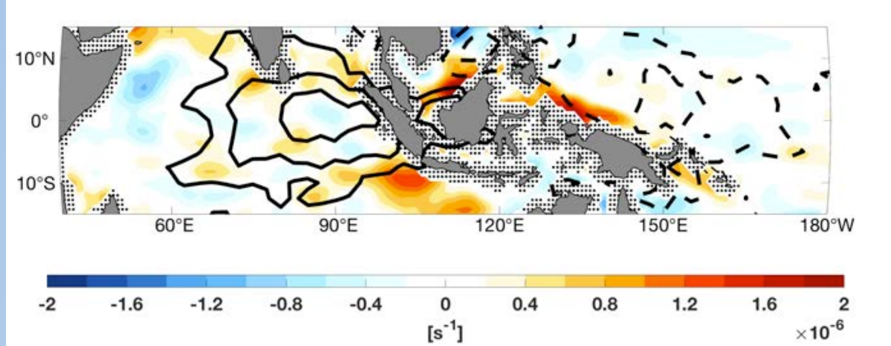
Full MLM

MLM BL

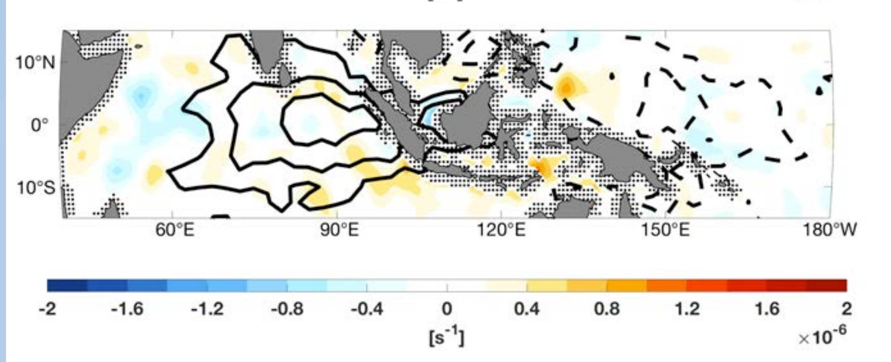
MLM – SST Only



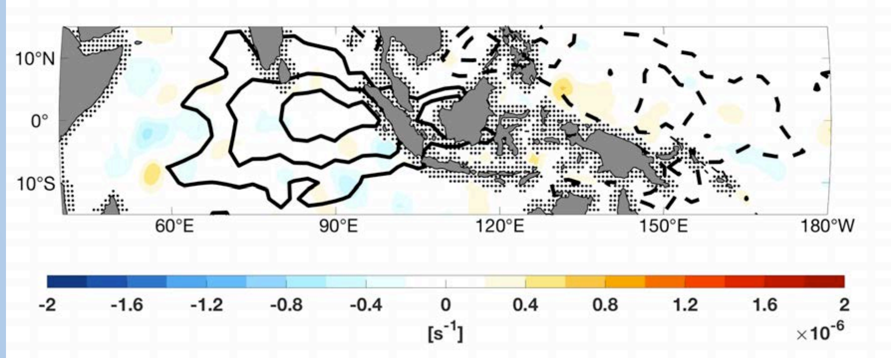
Residual Convergence



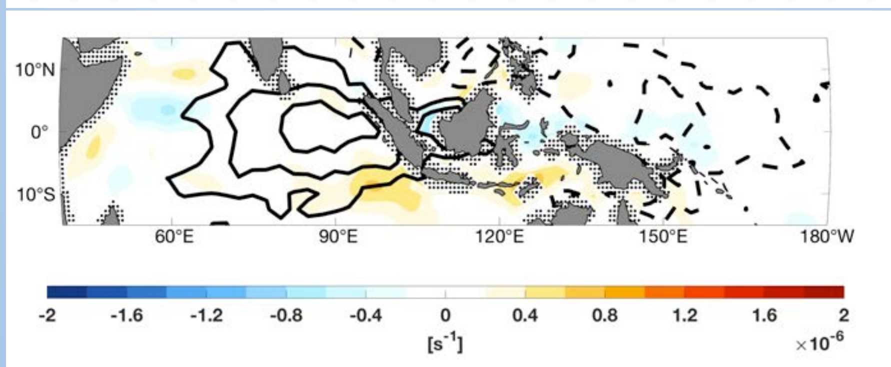
MLM SST + T850



MLM SST Only



MLM T850 Only



Sanity Check!

So far we have found that BL convergence is driven primarily (2/3) by processes occurring in the free troposphere, and secondarily (1/3) on smaller scales by pressure gradients originating in the BL. Back and Bretherton found the opposite for monthly to annual timescales.

Do our findings actually make sense?

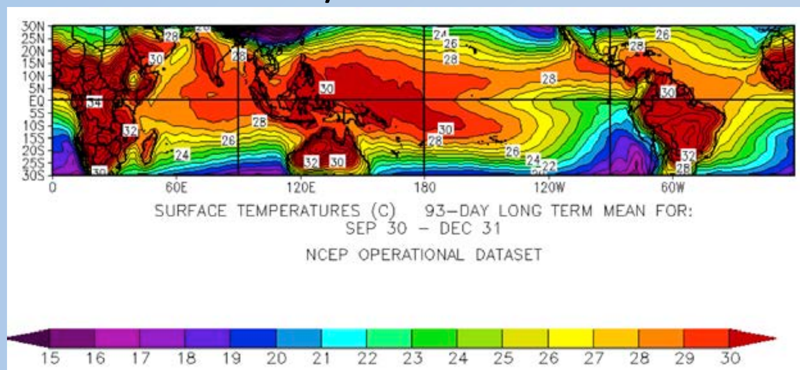
Sanity Check!

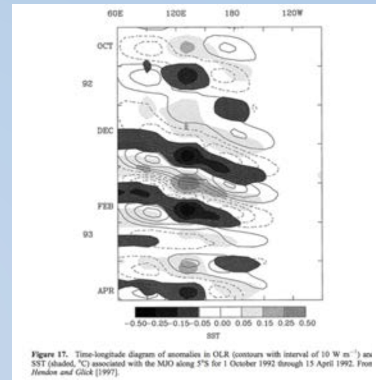
Back of envelope calculation, assume:

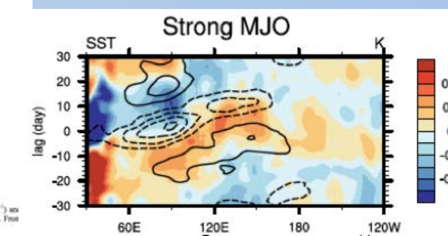
1.) Boundary Layer T = SST

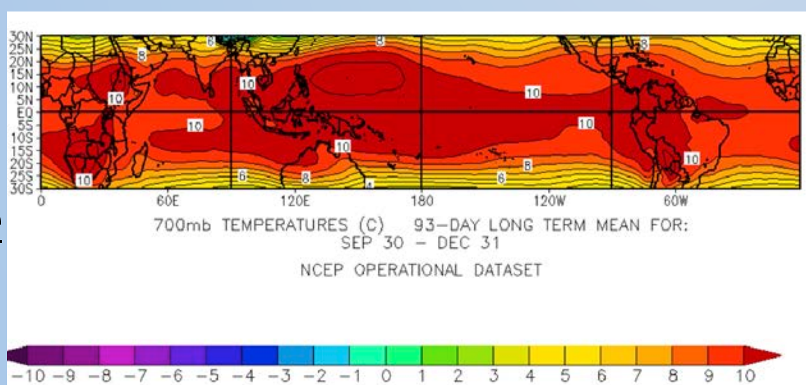
2.) dP/dx goes as dT/dx

	Climatology	**MJO**
Boundary Layer	dT/dx ~ 10K	dT'/dx ~ 1K
Free Troposphere	dT/dx ~ 1K	dT'/dx ~ 1K
Order Of Magnitude	BL / Free Troposphere ~ 10/1	BL / Free Troposphere ~ 1/1









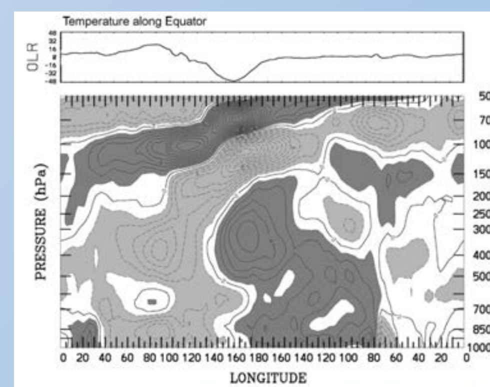


FIG. 7. As in Fig. 3, except for anomalous temperature along the equator. Contour interval is 0.1 K.

The MLM Worked for Back and Bretherton (BnB)

Annual Mean Convergence

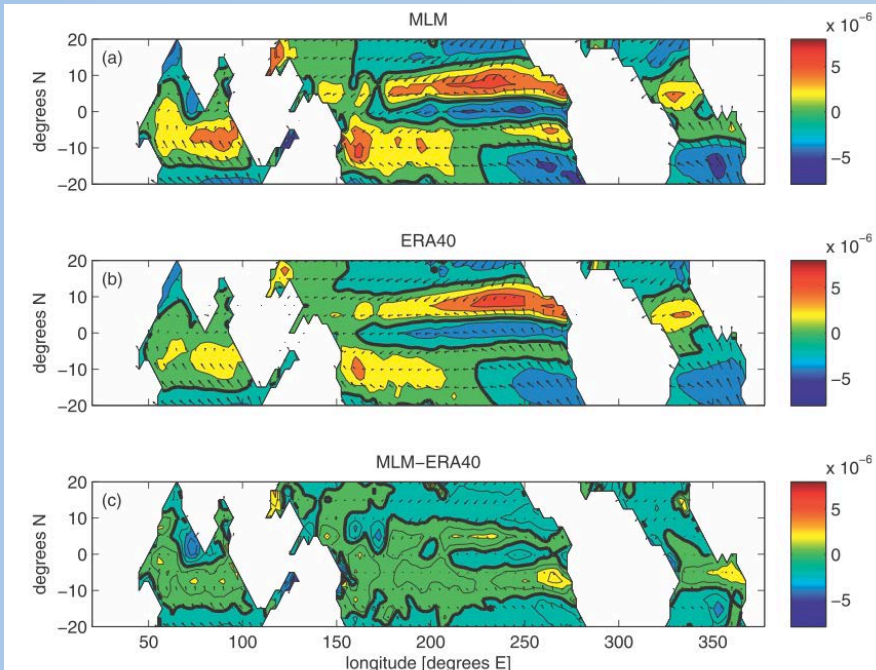


FIG. 2. Annual-mean surface winds and convergence from (a) MLM $[\mathbf{V}(\partial P_{BL}, U_T, \partial P_i)]$ forced by ERA-40 and (b) ERA-40, and (c) their difference. Here, 1° arrow length is 2 m s^{-1} wind and there are 2×10^{-6} s^{-1} convergence contours, except for in (c), where the contour interval is halved and the arrow length is doubled. There is a heavy zero contour for (a)–(c).

Comparison to QuickSCAT Obs.

	ERA-40	MLM
Correlations:		
Mean winds	0.99	0.98
Convergence	0.95	0.84
Seasonal winds	0.97	0.94
Convergence	0.91	0.82
Monthly winds	0.89	0.85
Convergence	0.76	0.65
RMS:		
Mean winds	0.45	0.74
Convergence $\times 10^6$	0.74	1.42
Seasonal winds	0.40	0.58
Convergence $\times 10^{-6}$	0.71	0.98
Monthly winds	0.89	1.11
Convergence $\times 10^{-6}$	1.42	1.80

MLM does a good job with annual mean winds and convergence. Winds are always explained better than convergence.

MLM performance decreases as timescale decreases… but still not bad.

Will the MLM Work on Intraseasonal Timescales ?

Separating the BL and Free Tropospheric Contributions

Separating the BL and Free Tropospheric Contributions

Why try to make this separation at 1500m?

From Johnson et. al. 2001

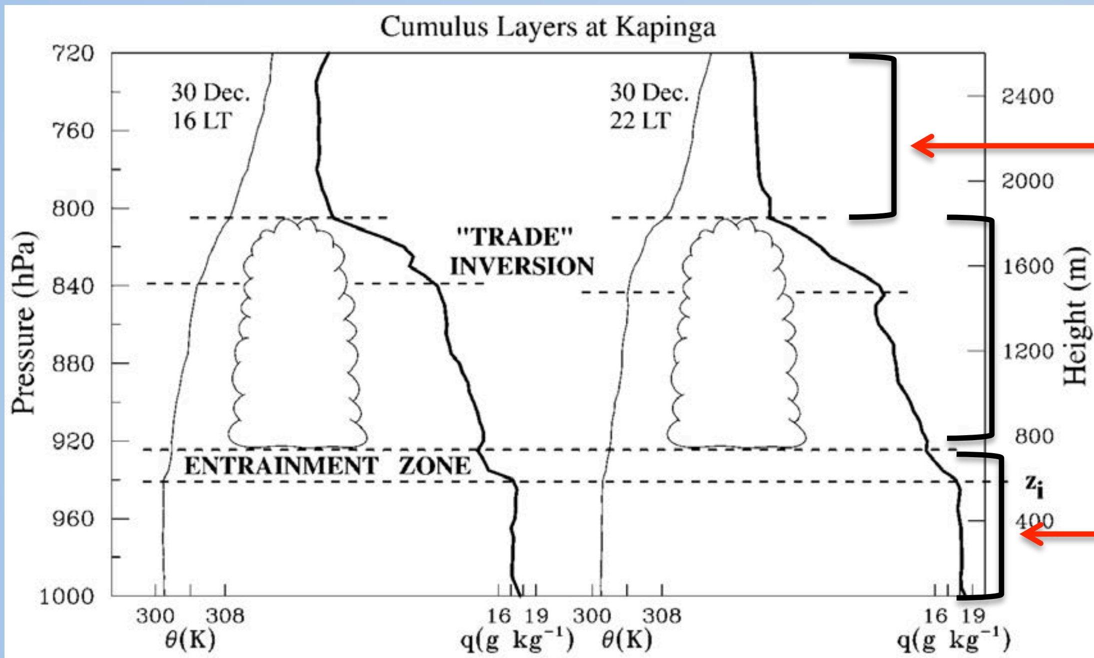


FIG. 3. Example of soundings (potential temperature θ in thin solid lines and specific humidity q in heavy solid lines) at Kapinga at 1600 and 2200 LST 30 Dec 1992 depicting a period when the mixed layer was topped with a shallow cumulus layer. Also indicated are the entrainment zone atop the mixed layer, and a "trade-type" inversion layer between 840 and 800 hPa.

Difficult to establish T gradients

Still strongly influenced by ML T gradients

Easier to establish T gradients

T gradients corresponds to pressure gradients. Trade inversion represents the upper bounds of where the surface may have a strong direct influence. 1500m is a decent approximation of the trade inversion, though it is likely to be higher in deep tropics and lower in subtropics. Keep in mind that different phases of the MJO may substantially modify the schematic above.

From Johnson et. al. 2001

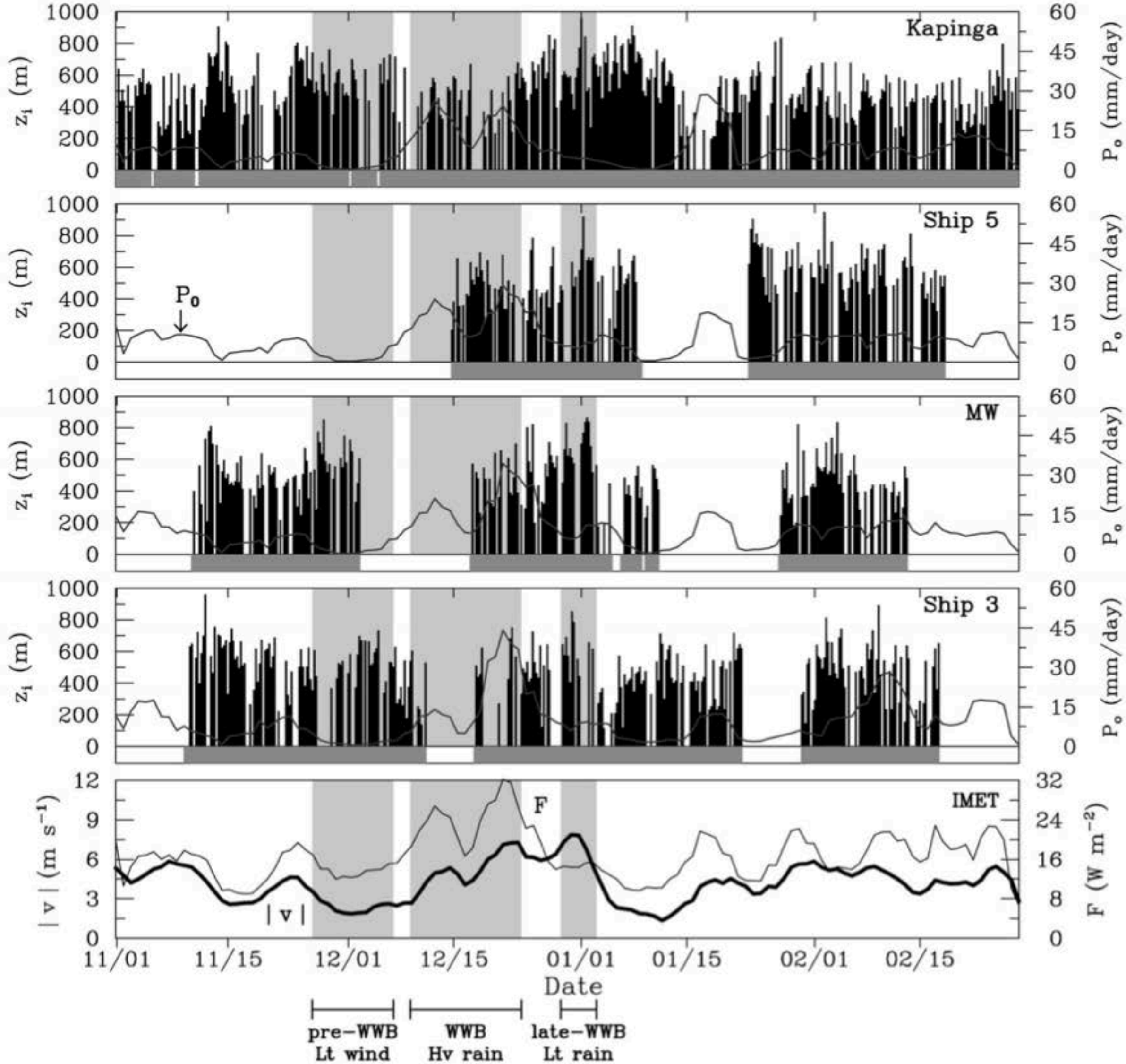


FIG. 7. (top four panels) Time series of mixed layer tops (top of vertical solid bars with scale to left) and 5-day running mean rainfall rate (thin gray line with scale to right) at four sounding sites. Light-shaded regions indicate three periods associated with the passage of an MJO and its associated westerly wind burst (WWB), characterized by different weather conditions noted at the bottom of the figure. Dark-shaded regions at bottom of panels indicate times at which soundings were taken. (bottom panel) Five-day running mean time series of surface wind speed (heavy line with scale to left) and bouyancy flux (thin line with scale to right) at the IMET buoy.

Separating the BL and Free Tropospheric Contributions

For this partitioning, we define the 850-hPa level as the BL top because this is near the climatological trade inversion, which caps active vertical mixing due to turbulence and shallow cumulus convection over much of the tropics. In addition, the 850-hPa surface is convenient because our ERA-40 dataset has been interpolated to pressure surfaces, including 850 hPa, which is close to where bottom-heavy vertical motion profiles peak. Introducing the geopotential Φ and denoting a mean over the belt 20°S–20°N by an overline, we write

$$P_s = P_i + \Delta P_{BL} \quad \text{and} \tag{3}$$

$$P_i = 850\,\text{hPa} + \rho_{850}(\Phi_{850} - \overline{\Phi}_{850}). \tag{4}$$

Here, P_i, which is approximately the pressure at the mean height of the 850-hPa surface, is calculated from zonally smoothed ERA-40 output. The boundary layer contribution ΔP_{BL}, calculated as a residual, is proportional to the mean temperature between the surface and 850 hPa.

From BnB

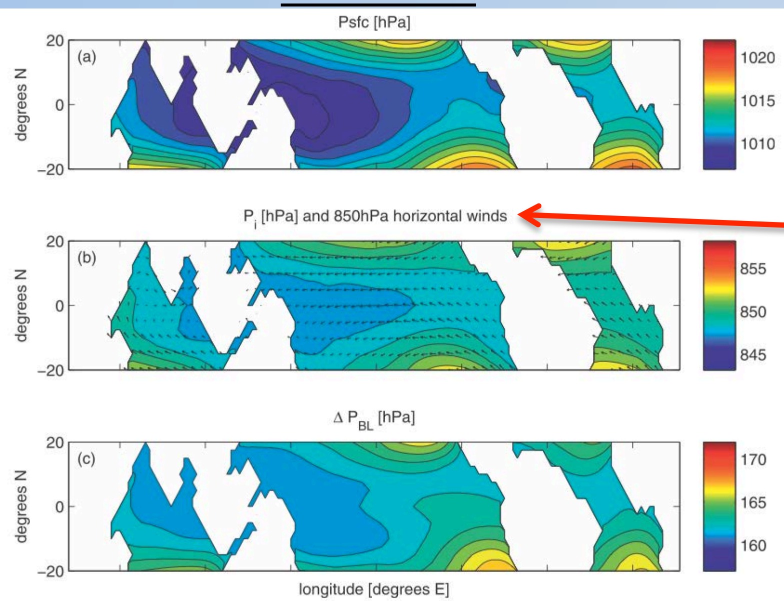
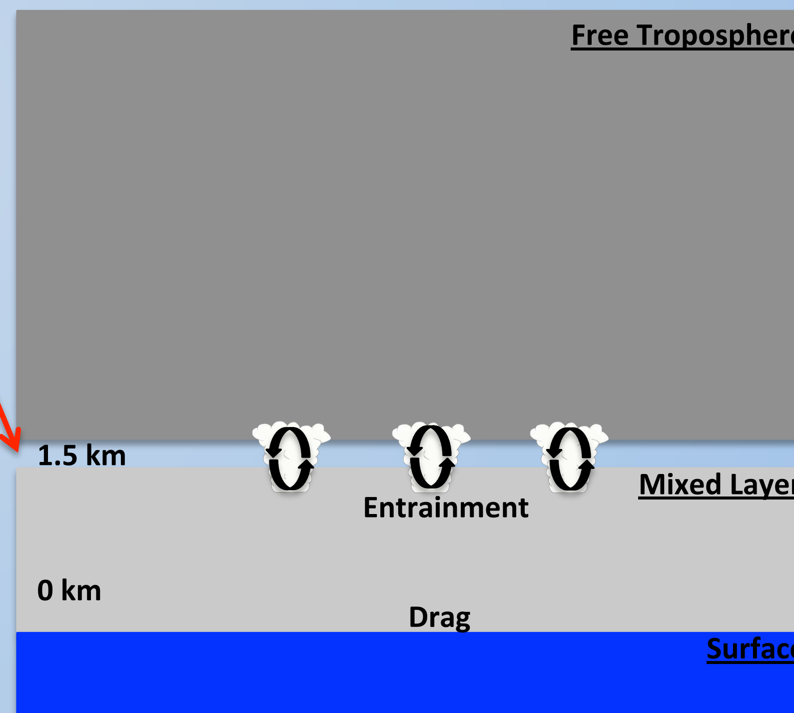


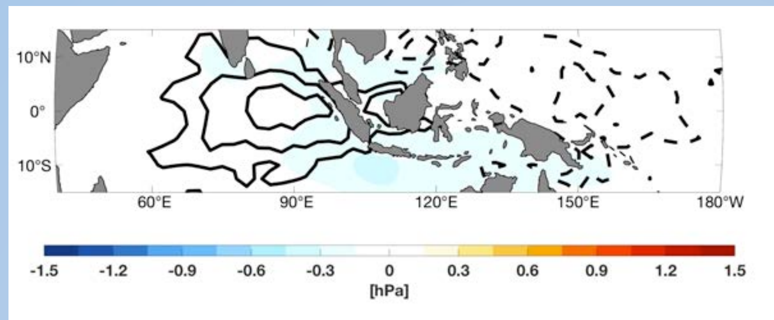
FIG. 3. ERA-40 1998–2001 mean (a) surface pressure, (b) free-tropospheric contribution to surface pressure [P_i in (4)] and 850-hPa vector winds, and (c) boundary layer contribution to surface pressure from below 850 hPa [ΔP_{BL} in (4)]. Here, 1° arrow length is 2 m s^{-1} wind.

Pressure at the z – height (which is fixed everywhere) of the top of the mixed layer



Separating the BL and Free Tropospheric Contributions

$\Delta P_{bl}'$
BL induced surface pressure anomaly



This very large-scale structure of boundary layer pressure anomalies (i.e. BL temperature anomalies) appears to be consistent with the findings of Kiladis et al. 2005 and Tian et al. 2010.

From Kiladis et al. 2005

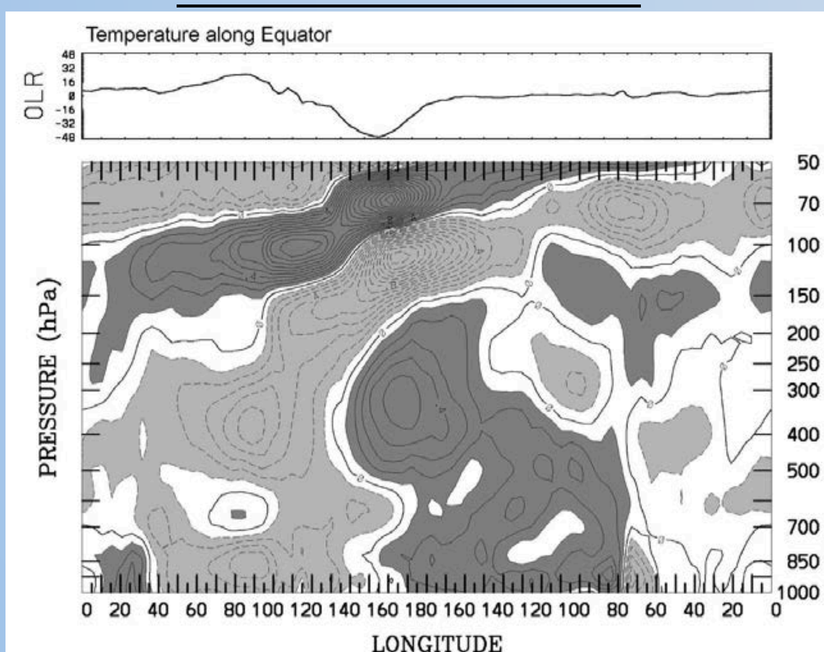


FIG. 7. As in Fig. 3, except for anomalous temperature along the equator. Contour interval is 0.1 K.

From Tian et al. 2010

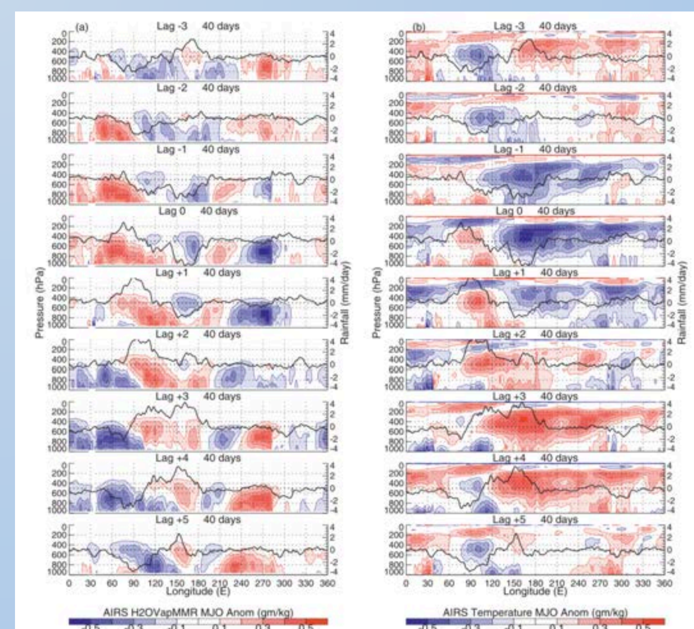


FIG. 2. Composite MJO cycle of equatorial mean (8°S–8°N) pressure–longitude cross sections of (a) specific humidity and (b) temperature anomalies (color shading) based on the 2.5-yr pentad V5 AIRS data and the MJO analysis method 1. The overlaid solid black lines denote TRMM rainfall anomalies (scales at right) for the same period for the AIRS data.

From BnB **Separating the BL and Free Tropospheric Contributions**

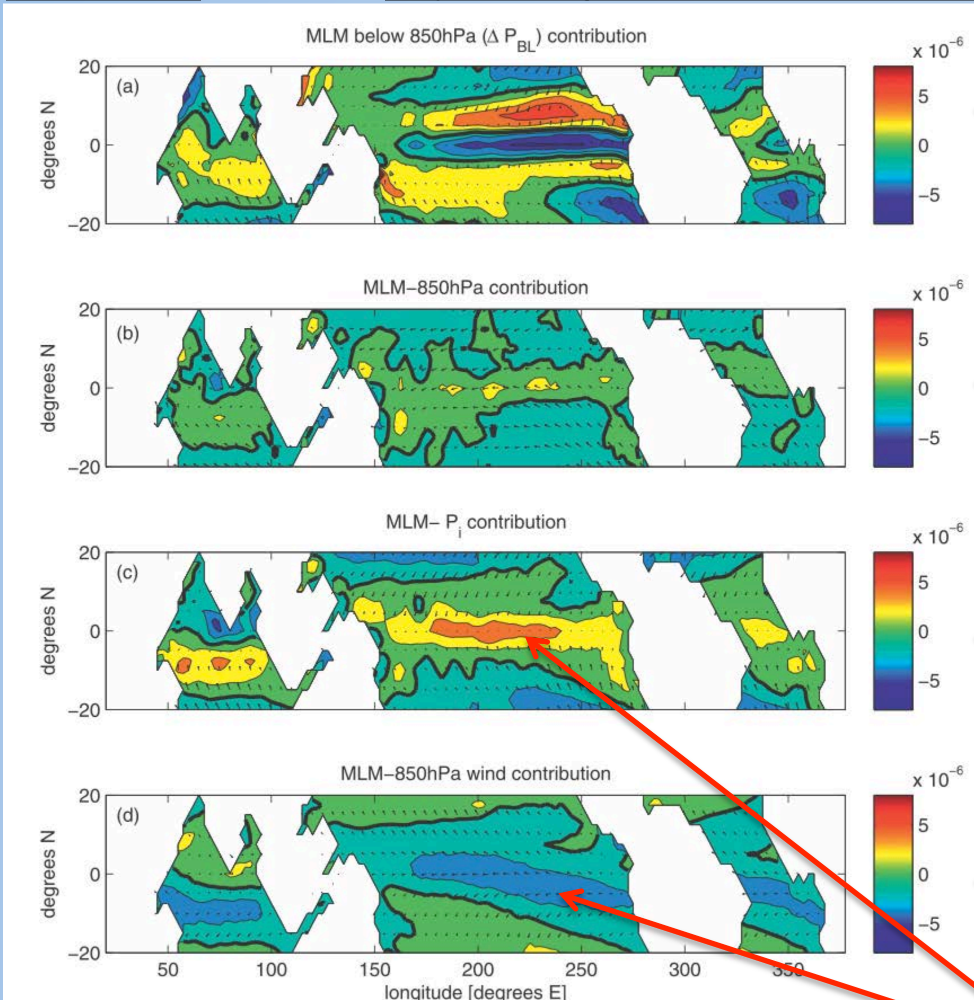


FIG. 4. Components of MLM solution resulting from (a) ΔP_{BL}, the boundary layer contribution to surface pressure gradients $[\mathbf{V}(\partial P_{BL}, U_T = 0, \partial P_i = 0)]$; (b) above-850-hPa (free tropospheric) processes $[\mathbf{V}(\partial P_{BL} = 0, U_T, \partial P_i)]$; (c) P_i, the above-850-hPa contribution to surface pressure gradients $[\mathbf{V}(\partial P_{BL} = 0, U_T = 0, \partial P_i)]$; and (d) downward momentum mixing $[\mathbf{V}(\partial P_{BL} = 0, U_T, \partial P_i = 0)]$. The plotting conventions are the same as Fig. 2. (b) is the sum of (c) and (d).

BL induced pressure gradient contribution

Free tropospheric contribution (wind mixing and pressure gradient)

=

Free tropospheric induced pressure gradient contribution

+

Free tropospheric wind mixing contribution

Notice the large degree of cancellation!

Convergence results primarily from BL induced pressure gradients (i.e. top panel).

Separating the BL and Free Tropospheric Contributions

From BnB

Comparison to QuickSCAT Obs.

	ERA-40	MLM	MLM BL	MLM deep	MLM-SST	MLM-SST BL	MLM-SST SST only
Correlations:							
Mean winds	0.99	0.98	0.83	0.90	0.97	0.76	0.56
Convergence	0.95	0.84	0.85	0.13	0.65	0.62	0.62
Seasonal winds	0.97	0.94	0.77	0.81	0.91	0.65	0.49
Convergence	0.91	0.82	0.78	0.14	0.51	0.48	0.48
Monthly winds	0.89	0.85	0.68	0.72	0.82	0.54	0.40
Convergence	0.76	0.65	0.65	0.12	0.40	0.38	0.38
RMS:							
Mean winds	0.45	0.74	2.43	1.70	0.89	2.66	3.21
Convergence $\times 10^{6}$	0.74	1.42	1.31	2.51	2.49	2.53	2.53
Seasonal winds	0.40	0.58	0.98	0.88	0.67	1.13	1.30
Convergence $\times 10^{-6}$	0.71	0.98	1.08	1.75	1.72	1.72	1.72
Monthly winds	0.89	1.11	1.35	1.27	1.17	1.54	1.67
Convergence $\times 10^{-6}$	1.42	1.80	1.72	2.28	2.70	2.59	2.59

Convergence results primarily from BL induced pressure gradients

The Role of SST Anomalies

SST Gradients and Their Influence on Boundary Layer Pressure Gradients

From BnB

Connection to SST

To construct an LN-like model, we must relate gradients in the boundary layer pressure contribution ΔP_{BL} to gradients in SST. A simple approximation is to hydrostatically estimate ΔP_{BL} by assuming the air temperature varies linearly between a surface value equal to the SST and the ERA-40 850-hPa temperature, which is roughly at the mean inversion height; density can then be linearized about the mean temperature and integrated (as in LN). Figure 5a shows the corresponding

From Lindzen Nigam

$$\rho = \rho_0[1 - n(T - T_0)], \tag{2a}$$

where

$$\rho_0 = \rho(T_0) = 1.225\ \mathrm{kg\ m^{-3}}, \quad T_0 = 288\ \mathrm{K},$$

$$n = -\left[\frac{1}{\rho}\frac{\partial \rho}{\partial T}\right]_{T_0} = 1/T_0,$$

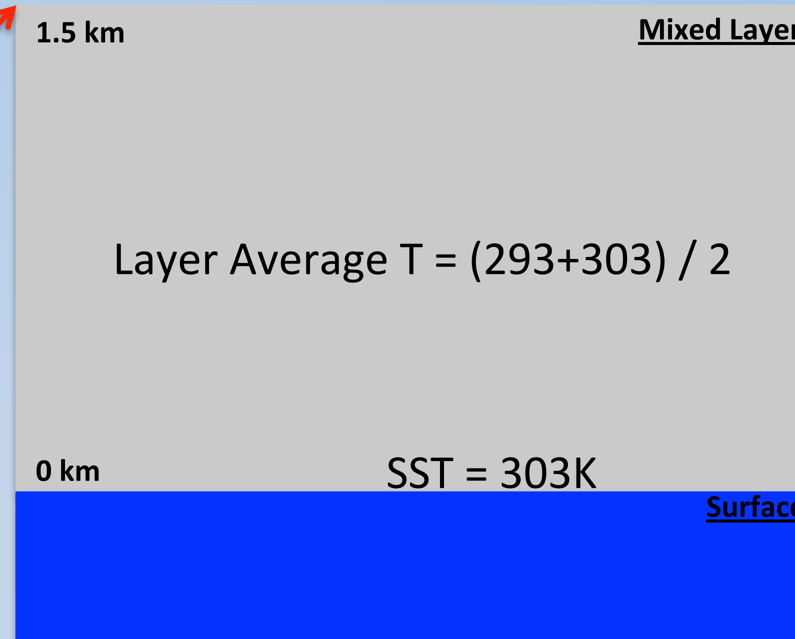
so that

$$\rho = \rho_0[2 - nT]. \tag{2b}$$

From the knowledge of density and temperature fields, the three-dimensional pressure field can be constructed using the hydrostatic equation and a boundary condition or an integration constant; assuming the latter to be a specification of the geopotential height field

T(850 hPa) = 293K

We don't actually know T at top of mixed layer, so we estimate using 850 hPa T



Climatological Example

SST = 303K

T(850 hPa) = 293K

Assume linear variation of T →

Layer Average T = 298K

Linearize density about T, with reference T and density from LN above:

$\rho = \rho_0[2\text{-}nT] = 1.225\ \mathrm{kg\ m^{-3}}\ [2 - 298\mathrm{K}/288\mathrm{K}] = 1.18\ \mathrm{kg\ m^{-3}}$ ← Layer average T is higher than reference T, so makes sense that density is less than reference density

Use hydrostatic to integrate density over depth of mixed layer to get pressure:

$\Delta P = \rho * g * \Delta z$ → $\Delta P_{BL} = \rho * g * \Delta z_{BL} = 1.18\ \mathrm{kg\ m^{-3}} * 9.8\ \mathrm{m\ s^{-2}} * 1500\ \mathrm{m} = 173.5\ \mathrm{hPa}$

SST Gradients and Their Influence on Boundary Layer Pressure Gradients

From BnB

Connection to SST

To construct an LN-like model, we must relate gradients in the boundary layer pressure contribution ΔP_{BL} to gradients in SST. A simple approximation is to hydrostatically estimate ΔP_{BL} by assuming the air temperature varies linearly between a surface value equal to the SST and the ERA-40 850-hPa temperature, which is roughly at the mean inversion height; density can then be linearized about the mean temperature and integrated (as in LN). Figure 5a shows the corresponding

From Lindzen Nigam

$$\rho = \rho_0[1 - n(T - T_0)], \tag{2a}$$

where

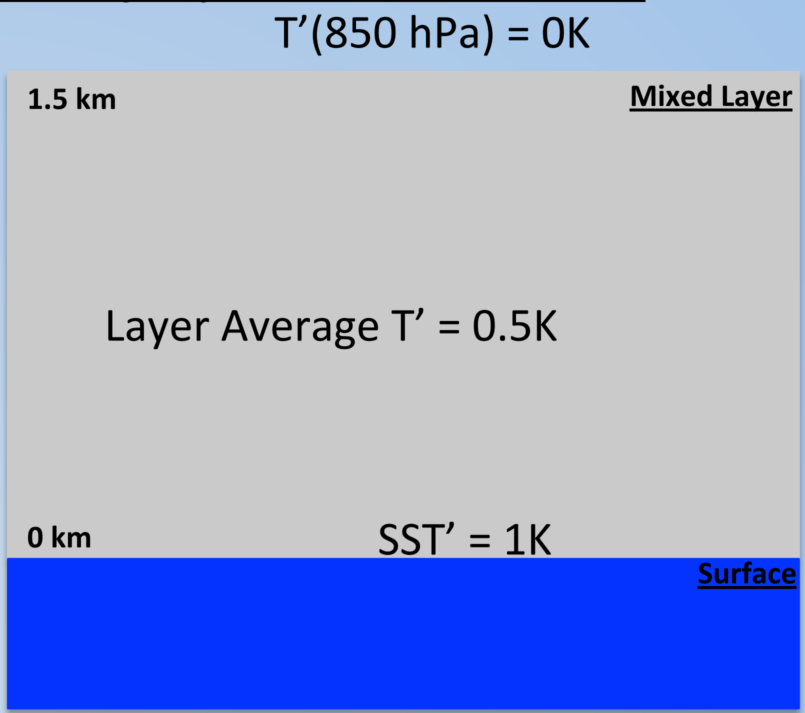
$$\rho_0 = \rho(T_0) = 1.225\,\text{kg m}^{-3}, \quad T_0 = 288\,\text{K},$$

$$n = -\left[\frac{1}{\rho}\frac{\partial\rho}{\partial T}\right]_{T_0} = 1/T_0,$$

so that

$$\rho = \rho_0[2 - nT]. \tag{2b}$$

From the knowledge of density and temperature fields, the three-dimensional pressure field can be constructed using the hydrostatic equation and a boundary condition or an integration constant; assuming the latter to be a specification of the geopotential height field



Intraseasonal Example

SST' = 1K

T'(850 hPa) = 0K

Assume linear variation of T →

Layer Average T' = 0.5K

Linearize density about T, with reference T and density from LN above:

$\rho' = \rho_0[1-(T_0+T')/T_0] = 1.225\text{ kg m}^{-3}\,[1 – (288\text{K}+0.5\text{K})/288\text{K}] = -0.0021\text{ kg m}^{-3}$

Use hydrostatic to integrate density over depth of mixed layer to get pressure:

$\Delta P' = \rho' * g * \Delta z$ → $\Delta P'_{BL} = \rho' * g * \Delta z_{BL} = -0.0021\text{ kg m}^{-3} * 9.8\text{ m s}^{-2} * 1500\text{ m} = -0.32\text{ hPa}$

SST Gradients and Their Influence on Boundary Layer Pressure Gradients

We can also use these values to sanity check our results:

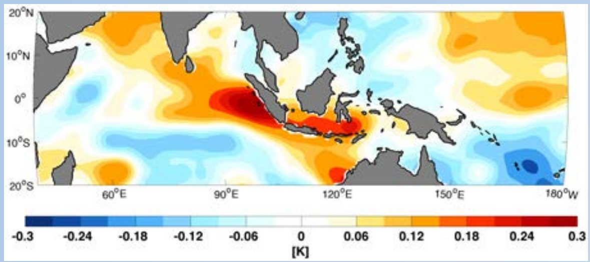
Layer Average T’ = 0.5K → $\Delta P'_{BL}$= -0.32 hPa

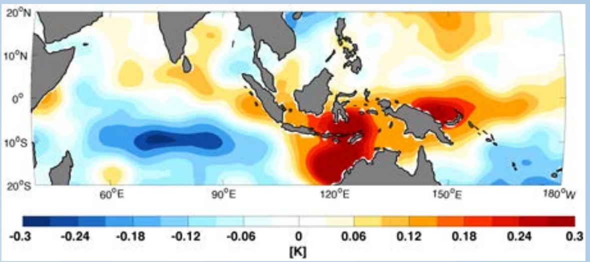
Layer Average T’ = 1K → $\Delta P'_{BL}$= -0.64 hPa

Strong Event, Day -5

Strong Event, Day 0

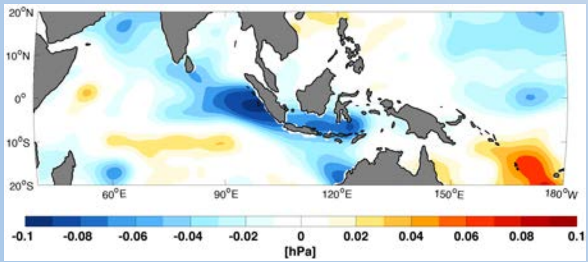
SST’

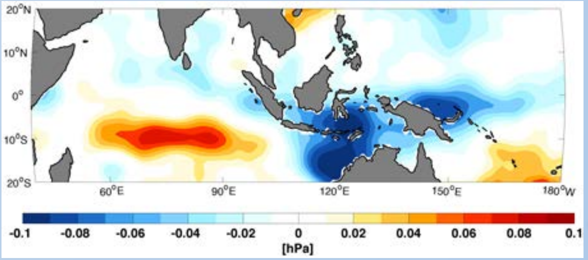




If we assume layer average T’ is ½ of SST’

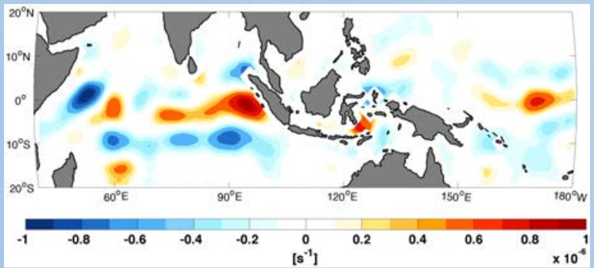
Corresponding Pressure Anomaly

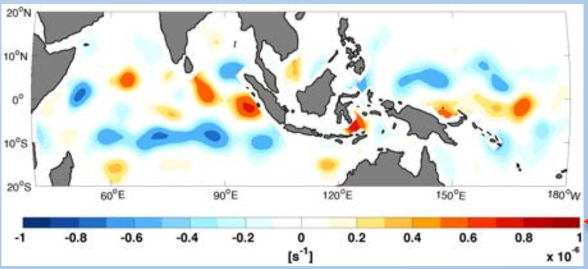




These P anomalies are very small…

Corresponding Convergence Anomaly





… but this convergence is actually pretty strong, especially at day -5