The El Niño Southern Oscillation (ENSO)

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Contents

1 Introduction

This is intended to be a brief overview of the El Niño Southern Oscillation (ENSO) and some important characteristics. It won't dig too far into the dynamics but should cover enough to give an idea of the mechanisms at play.

2 Mean climate

A key component of ENSO dynamics is the mean state of the Tropical Pacific. While this seems obvious, we tend to look at ENSO through the lens of anomalies in the wind and sea surface temperature (SST) field, and so it can be easy to lose sight of the mean field.

2.1 Tropical Pacific ocean-atmosphere climatology

We'll start by covering the mean winds in the tropical Pacific, and how those winds drive an ocean circulation. As a quick review, remember that in the tropics there are the easterly Trade winds blowing along the equator at the surface (see Fig. 1). There is also a Hadley cell, consisting of rising motion at the equator, and poleward movement of air aloft. The rising motion of the Hadley cell coincides with convection and the Intertropical Convergence Zone (ITCZ).

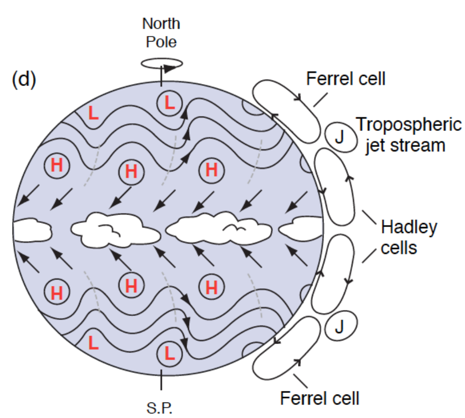


Figure 1: Schematic of the annual mean zonally symmetric general circulation. H represents high pressure, L represents low pressure systems.
Credit: Wallace and Hobbs (2006)

If we consider a zonal atmospheric circulation, we have the Walker circulation (see Fig. 2). The Walker circulation consists of convection over the maritime continent (coincident with consistently warm SSTs) and convection over the Amazon and Africa.

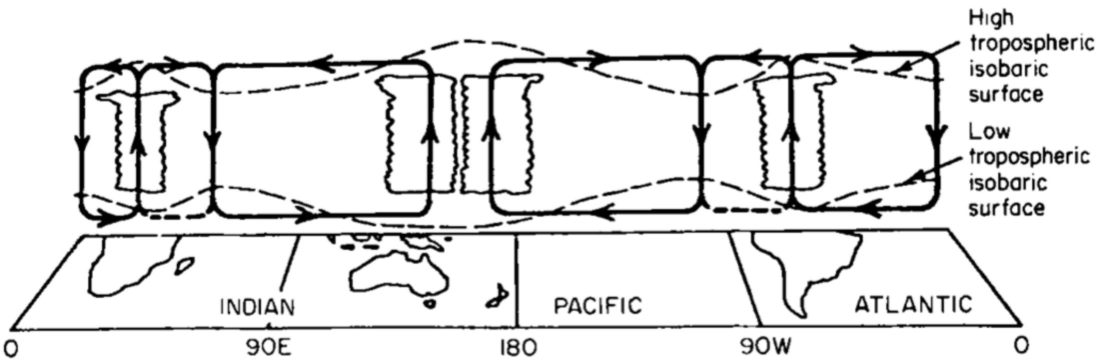


Fig. 6.22 Schematic view of the east-west Walker circulation along the equator indicating low-level convergence in regions of convection where mean upward motion occurs. [From Webster (1983).]

Figure 2: Schematic of the Walker CirculationCredit: Vallis (2006)

We can next consider how this mean atmospheric circulation drives an oceanic circulation. The Trade winds in the Tropics dive a North and South Equatorial Current (Fig. 4), which run from East to West a few degrees off the equator. This is the dominant flow at the equator and will be of the greatest importance in our conceptual model of ENSO

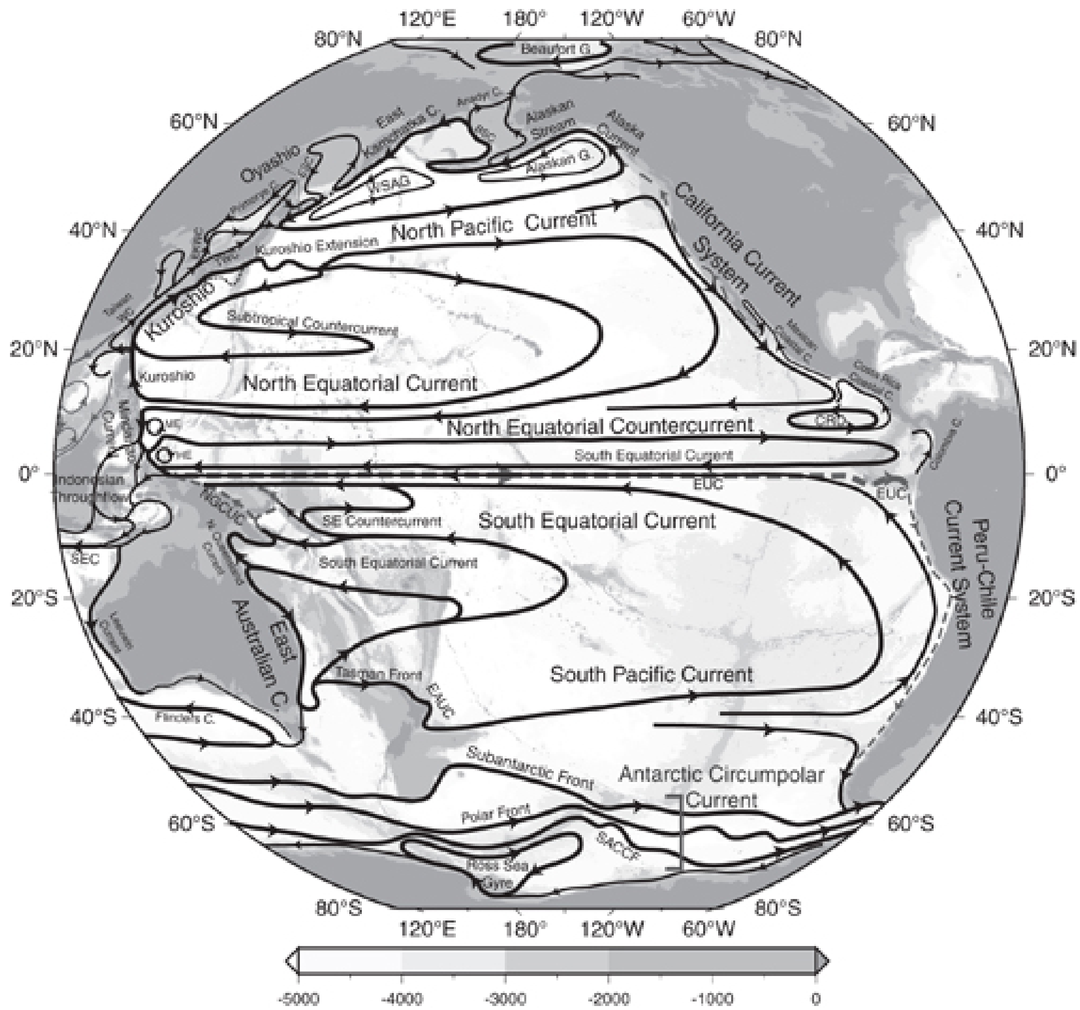


Figure 3: Credit: Talley 2010. Surface circulation scheme. Major near-surface undercurrents at the equator and along the eastern boundary are also shown (dashed). The South China Sea circulation represents the winter monsoon.

Note, however, that there is a narrow band near the equator in which water flows from west to east, transporting warm water from the west Pacific warm pool through to the Americas (this is referred to as the North Equatorial Countercurrent (NECC)). The NECC is somewhat counterintuitive because it runs against the Trade winds. However, this is a succinct demonstration that zonal currents are in geostrophic balance with pressure gradient force due to the slope of the sea surface height (i.e. the Coriolis force must act opposite to the pressure gradient force, and because the Coriolis force acts to the right of the flow, you can deduce what the geostrophic flow must be). See Figure **??** for a zonally averaged depiction of sea surface height (and therefore meridional pressure gradient), which is balanced by the Coriolis force (therefore dictating the zonal direction of the flow).

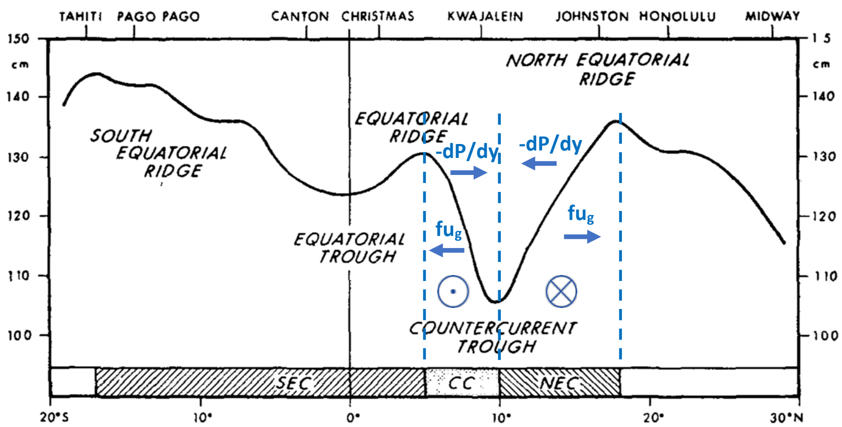


FIG. 1. Meridional profile of dynamic height relative to 500 db averaged zonally between 140W and 170E together with positions of sea level stations, topographic features, and currents.

Figure 4: Annotated figure from Wyrtki (1974) describing geostrophic balance and the north equatorial counter current (NEC)

Despite this thin latitude band of west-to-east flow, keep in mind that the dominant flow is in the opposite direction from east to west. The consistent east-to-west flow near the equator means that water must upwell from depth to maintain conservation of mass. Bringing cold water up from below the thermocline has two important consequences: (1) it creates a surface temperature gradient that is cold in the East and relatively warmer in the west and (2) it means that the thermocline on the equator tends to be tilted from deep in the West to shallow in the East (see Fig. 5 and Fig. 6). Both of these aspects of the mean ocean state are critical to ENSO dynamics.

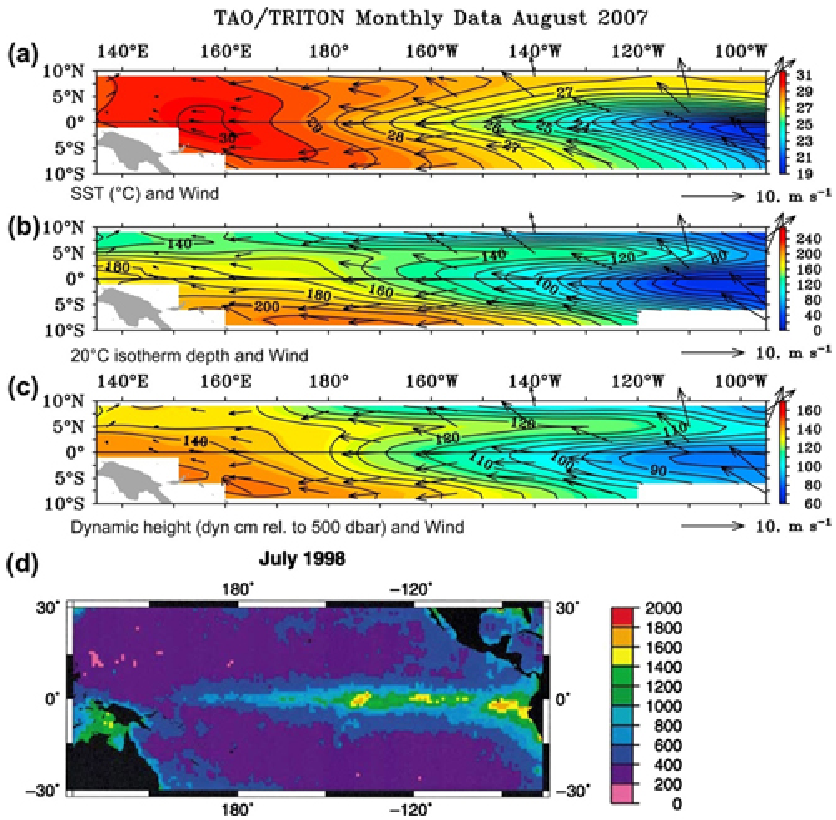


Figure 5: Credit: Talley 2010. (a) SST; (b) depth of the 20 degree C isotherm, which is an indicator of thermocline depth; and (c) dynamic height (dyn cm), with superimposed wind velocity vectors, during a period of a well-developed cold tongue (La Nina; August 2007). Source: From TAO Project Office (2009a). (d) Primary productivity (mg C m2 day1) based on ocean color, during a La Nina (July 1998). Source: From McClain et al. (2002).

Finally, we can explore the zonal mean circulation with depth in the Pacific basin (see Fig 6). The prominent features here are the slope of the sea surface height from low in the east to high in the west as a result of the mean easterly Trade winds (note the west-ward wind-driven surface current). This sea surface height variation sets up a mean pressure gradient across the basin that drives the Equatorial Undercurrent, which runs from west to east above the thermocline but below the wind-driven circulation. This figure also makes aparent the slope of the thermocline, a result of the easterly surface winds that drive Ekman divergence at the surface, which in turn necessitates upwelling of cold waters from below. The tilt of the thermocline plays an integral role in triggering and sustaining ENSO events.

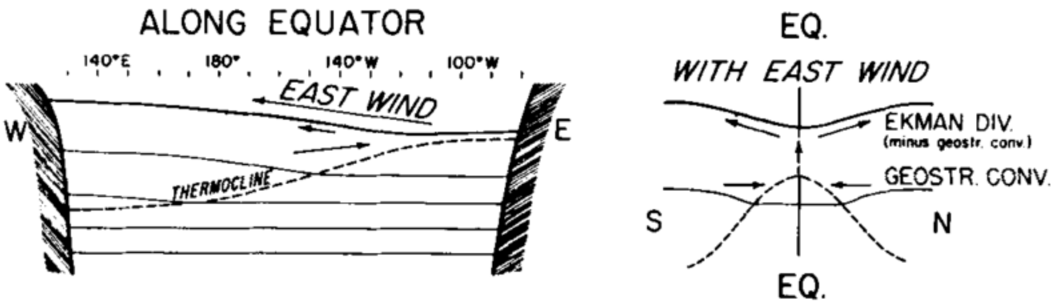


Figure 6: Credit: Figure 5 of (Bjerknes, 1966). The mean circulation at depth in the Pacific (left panel) from west to east across the basin and (right panel) zonal mean circulation in the east Pacific

2.2 Extratropics

Although the Tropics are the main region of interest for ENSO, if we want to think about ENSO dynamics on longer (decadal) timescales, we need to also consider the mean state of the extratropics.

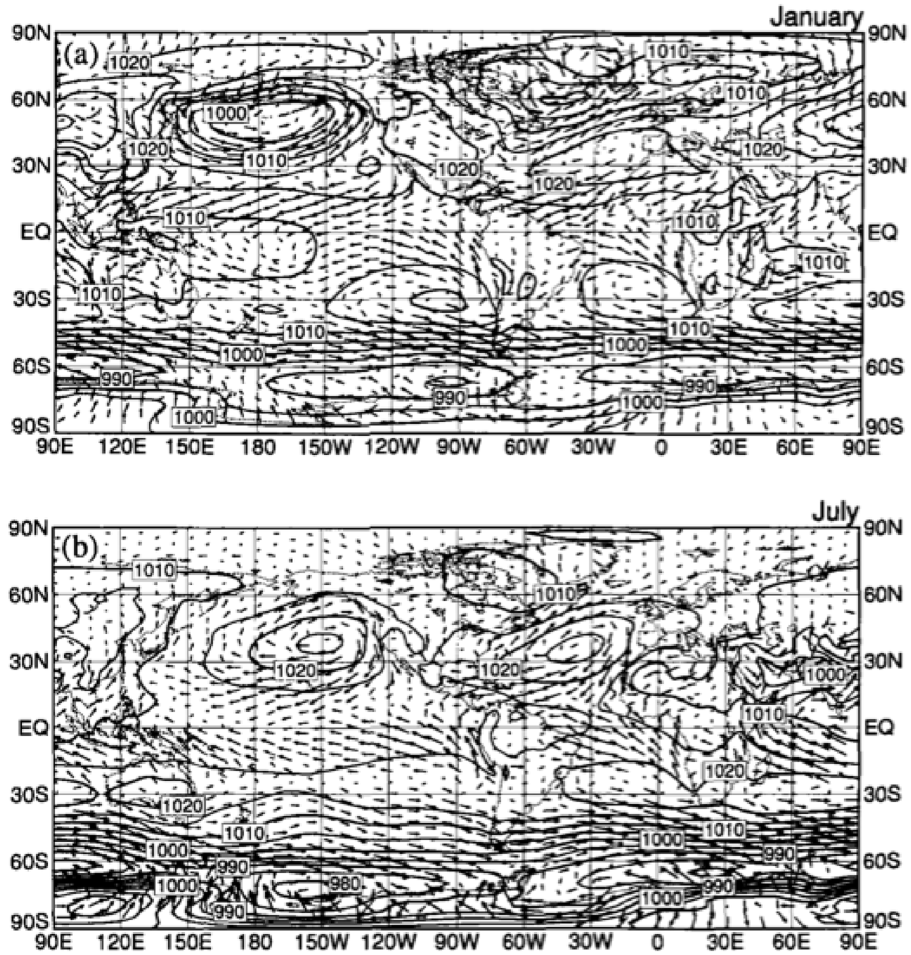


Fig. 6.18 Maps of mean sea-level pressure for (a) January and (b) July. Wind vectors for the 1000-mb level are superimposed. Data are 1980–87 analyses from a forecast model. Contour interval is 5 mb and largest vector represents a wind speed of 12 m s^{-1}.

Figure 7: Credit: (Hartmann, 2015)

3 The El Niño Southern Oscillation

Now that we have set the stage, we can start talking about ENSO. You may already know that ENSO refers to a semi-regular oscillation with a preferred frequency of 3-7 years. And that El Niño refers to an anomalously warm tropical Pacific while La Niña refers to a cold tropical Pacific. In the following sections we'll talk about how the system oscillates from neutral to warm to cold, and we'll describe the spatial and temporal characteristics of the system.

3.1 Physical mechanisms

Bjerknes was the first to propose a feedback mechanism between the equatorial Trade winds and the mean ocean state that could lead to the observed warmings of the tropical Pacific. Figure 8 describes how relaxing of the mean trade winds means that you also halt Ekman divergence and weaken (or remove) the wind-driven East-to-West surface current. The mean gradient in sea surface height (higher in the west, lower in the east) then drives a current eastward. The lack of Ekman divergence also deepens the thermocline in the east. Both the weakened upwelling and the advection of warm waters from west-to-east makes the eastern Pacific anomalously warm. This is the El Niño state of the system.

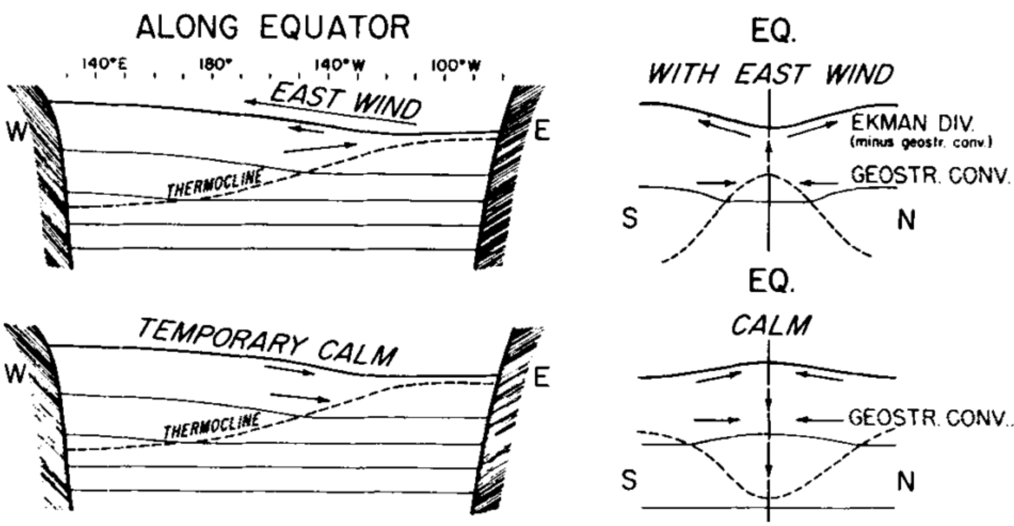


Fig. 5. Schematic equatorial flow patterns, with and without westward wind stress.

Figure 8: Credit: Bjerknes, 1966.

Now that the eastern Pacific is anomalously warm, the center of convection in the pacific migrates further east (see Fig. 9), following the warm SSTs (note that it is the total SST not the anomaly that sets the location of convection). The shifted convection leads to lower atmospheric pressure further east than in normal conditions, which induces westerly winds on the equator (as opposed to the climatological easterlies), thus completing the feedback loop.

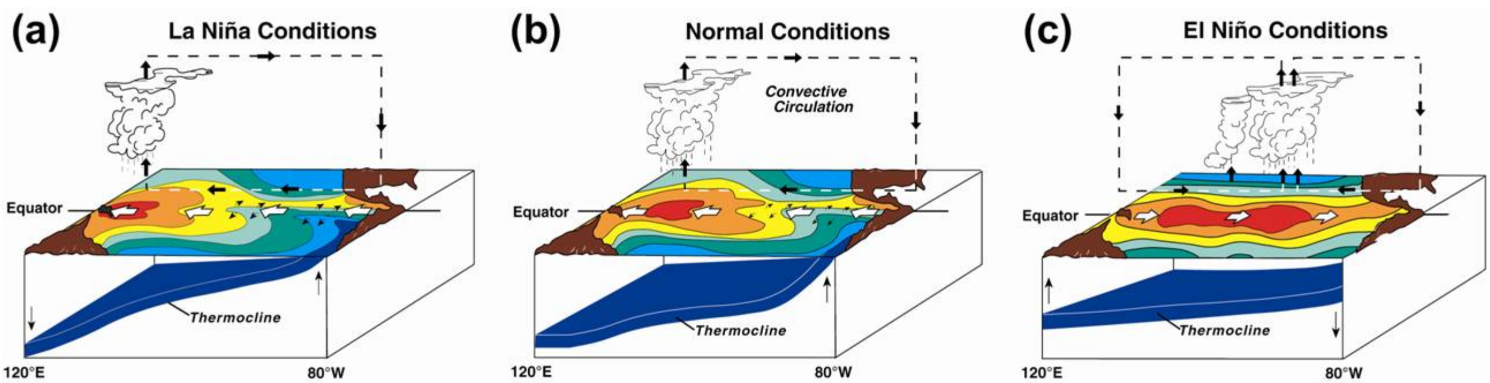


Figure 9: Credit: Talley 2010. (a) La Nina, (b) normal, and (c) El Nino conditions. This figure can also be found in the color insert. Source: From NOAA PMEL (2009b).

The La Niña phase of ENSO is the mirror of the El Niño phase. Anomalously strong trades on the equator lead to increased upwelling, which brings up more cold water than normal and makes the eastern Pacific anomalously cold. This means that convection is shifted slightly further West, and the altered SST gradient drives stronger yet trades, completing the feedback loop.

3.2 Temporal characteristics

The feedback described above occurs irregularly on an interannual timescale (happening every 3-7 years on average) but has a much more reliable pattern of development within a year. ENSO events reliably develop in the late boreal summer, peak in the early winter and decay in the spring. So what makes ENSO locked in its phase to the seasonal cycle but erratic on interannual timescales?

3.2.1 Seasonal phase-locking

In previous sections we described the annually averaged mean state of the Pacific, but its' important to consider also the mean climatology. As described in Zebiak and Cane (1987), the growth of anomalies into an ENSO event depends (in part) on the coupling strength between the ocean and the atmosphere at that time of year. This coupling is strongest in the summer and fall, weakest in the spring and intermediate in the winter. This is why ENSO events develop and decay in summer and fall (as the positive or negative feedback grows) and events tend to peak and change sign in the winter (when the coupling can't maintain the feedback). It is also why events don't develop in the spring, and why forecasting what an ENSO event will do going into the spring is difficult. So what causes the coupling to strengthen or weaken?

Here we turn back to the mean field. In the spring the trade winds are weakest, meaning that the associated upwelling is also weakest. With a weaker mean SST gradient and a thermocline that is deeper in the east than during other times of year, anomalies have less of an impact on the mean state and are therefore less likely to grow into the feedback necessary to initiate an ENSO event. The weaker mean state means you need greater temperature anomalies to alter a deeper mixed layer in the east, and you are less likely to advect anomalies from west to east with weaker ocean circulation. During the fall and summer the trade winds are strongest, which leads also to a strong SST gradient and upwelling. This mean state tends to communicate anomalies effectively (i.e. shallow mixed layer is strongly affected by SST anomalies).

The spring, while not a time of strong coupling between the atmosphere and the ocean, is an important season for the atmosphere. This is when the ITCZ extends far enough south to reach the equator, and when the eastern Pacific is at its warmest. This means that during this season atmospheric anomalies may organize (i.e. warm east Pacific leads to convection, latent heat warming air to make it rise, leading to low pressure and more convection). There is evidence that the atmospheric anomalies begin to organize in the spring but are only subsequently coupled to the ocean to begin the feedback critical for ENSO.

3.2.2 ENSO onset

Model simulations have shown that an important aspect for initiating an ENSO event is total ocean heat content in the mixed layer (equivalent to the depth of the thermocline). A deepening of the thermocline in the east pacific appears to be a prerequisite for an ENSO event, perhaps even pre-dating the anomalous westerly wind bursts. Without this change in thermocline depth, which has the important aspect not only of affecting the surface temperatures but also of giving the system 'thermal inertia' (i.e. a deeper, warmer mixed layer can maintain warm SST anomalies even against a noisy atmosphere, while a shallow mixed layer will reflect atmospheric forcing in SST anomalies immediately). This is at least partly responsible for the chaotic nature of ENSO on interannual timescales. Without the possibility of this thermal inertia, the tropical Pacific tends to lock into regular annual cycles that are phase locked to the seasonal cycle.

Equatorial Kelvin / Rossby waves stochastically forced and optimal growth modes that are ENSO-like

3.2.3 Decadal ENSO-like variability

A subject of constant interest is the low-frequency ENSO-like variability that is observed outside of the tropics, and whether it is an independent mode of variability, or whether it is a forced response to ENSO. Authors tend to try and separate the two timescales using a low-pass filter or by linearly regressing ENSO out of SSTs and doing an EOF on the remaining smoothed data. The result of these analyses is that interannual variability directly attributable to ENSO is strongest in the tropics, whereas the interdecadal variability is equally strong in the extratropics and midlatitudes as it is in the tropics (see SST anomalies in Fig. 10).

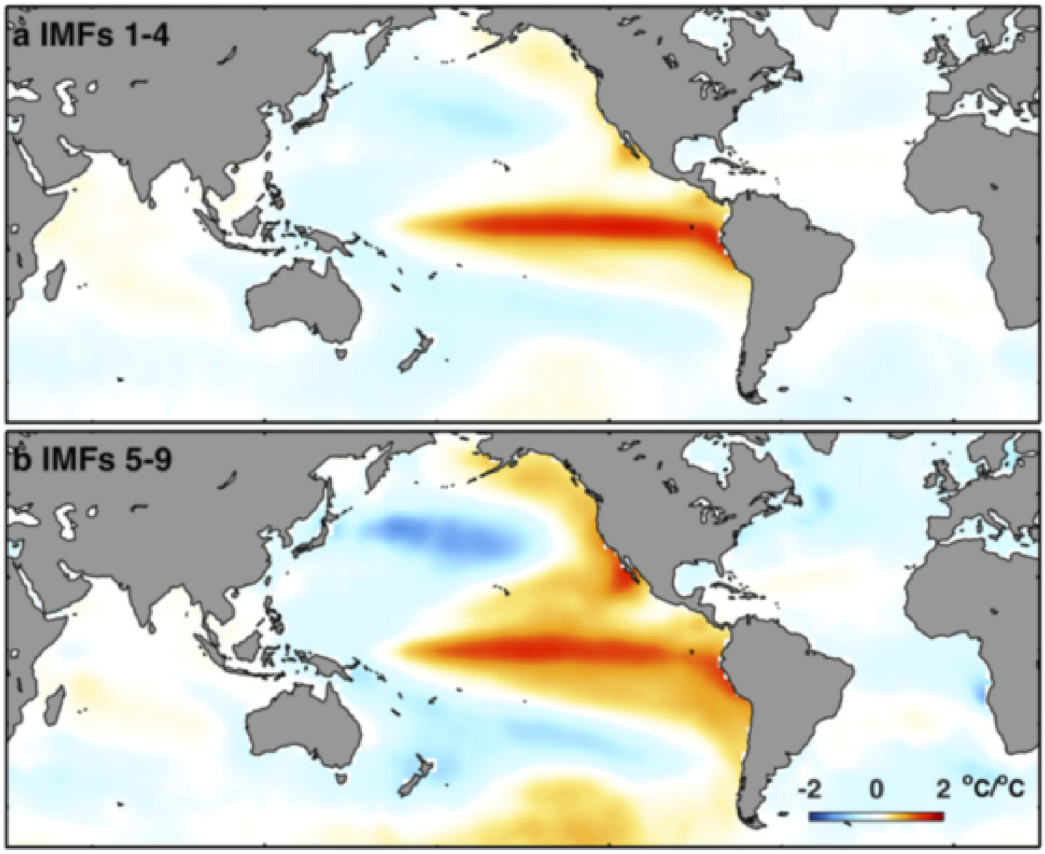


Figure 10: Credit: Figure 15 of Chen and Wallace (2015). The (a) dominant mode of interannual variability in the SST field and (b) dominant mode of decadal variability in the SST field.

The interdecadal variability generally has a center of action over the gulf of Alaska, and closely resembles the pattern associated with the Pacific North American (PNA) pattern (which is a preferred mode of internal variability in the atmosphere). It is indisputable that the PNA-forced teleconnections (SST, Sea level pressure, heat flux etc) strongly resemble the interdecadal ENSO-like variability. However, the causality is unclear. It is possible that the PNA pattern forces a red-noise response in the ocean that is communicated to the tropics, but it is also possible that ENSO forces the extratropical red-noise that resembles the PNA pattern. Both have been shown to be possibilities through various model experiments

4 ENSO teleconnections

Below we'll briefly discuss some mechanisms of ENSO teleconnections and describe their structure, although our listing is far from complete. We mainly focus on teleconnections over North America and the north Pacific, although ENSO teleconnections are a global phenomena.

4.1 Matsuno-Gill Model

To understand how the atmosphere will respond to an equatorial heating anomaly, we can consider the simplified linear model of Gill (1980), which uses the shallow water equations. If a heat source is added in the atmosphere on the equator, then the atmospheric response will be in the form of equatorial waves. Kelvin waves will travel east, inducing tropical easterly wind anomalies on the equator, supplying inflow to the heating anomaly from the east. Westerlies will also develop on the equator to the west of the anomaly, generated by Rossby waves. Because the fastest Rossby wave travels at $\sim 1/3$ the speed of a Kelvin wave, we expect the westtward extent of the response to penetrate only one-third as far as the eastward response. Also somewhat inexpectedly the meridional flow isn't towards the heating anomaly (as would be the case in the absence of earth's rotation) but is instead away from the heating. This is because here advection of planetary vorticity is balancing the anomalous ascent ($\beta v = \frac{\partial \omega}{\partial z}$). Figure 11 illustrates the vertical ascent, horizontal winds, and pressure anomalies generated by an equatorial heat anomaly at lower levels. Because the response is baroclinic, at upper levels the structure is the same but the sign of everything is reversed (i.e. off-equatorial high pressure).

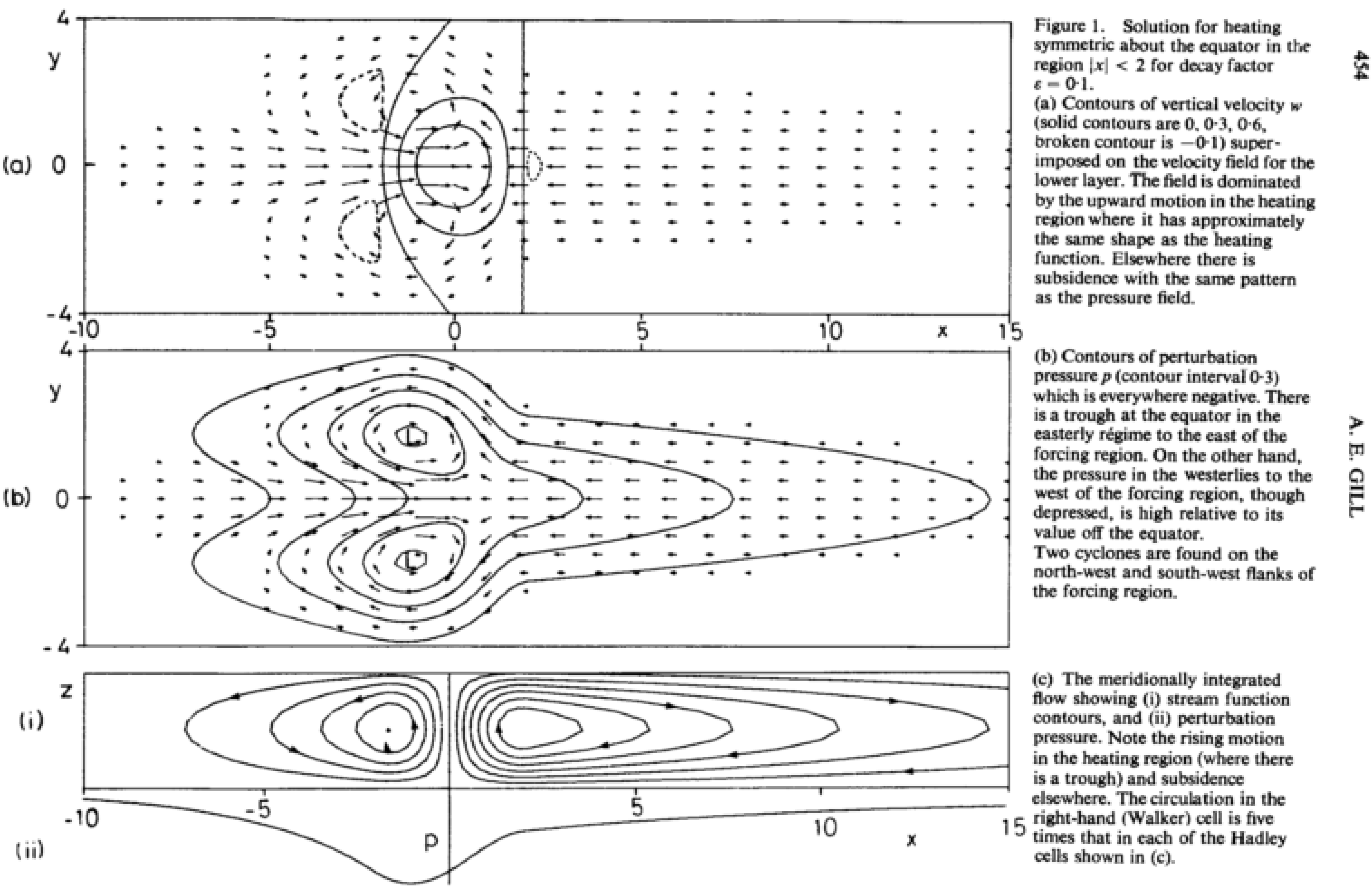


Figure 11: Credit: Gill (1980)

The zonal anomalies are somewhat intuitive: a heat source creates a low pressure anomaly, which forces inflow from both east and west. The meridional response is surprising, and must be understood from the perspective of vorticity. The vertical ascent (vortex stretching; $f\frac{\partial w}{\partial z}$) is balanced by advection of planetary vorticity (βv) such that poleward motion is associated with ascent. This is the equivalent of the Sverdrup balance in oceanography.

4.2 North Pacific teleconnections

While we previously described the Matsuno-Gill-type response, which generates Rossby and Kelvin waves in response to tropical heating, here we'll describe teleconnections from the perspective of how (1) ENSO shifts the heat source for the Hadley cell east and (2) provides an increased supply of moisture and energy into the Hadley cell.

During the warm phase of an ENSO event (El Niño), the jet streams tend to be shifted equatorward in agreement with what would be expected from increasing the temperature at the equator and therefore increasing the supply of moisture (and therefore energy) to the Hadley cell. In the eastern equatorial Pacific the thermal maximum is displaced from its usual 5-10 N to be centered on the equator. With these two changes in mind we will turn to teleconnections in the North Pacific

Typically teleconnections are strongest when an ENSO event is strongest, in the boreal wintertime. During boreal winter in the Northern Hemisphere the dominant atmospheric feature is a low pressure over the Aleutian Islands. Isobars usually run from southwest to northeast in the North Pacific, as you would expect from a thermal maximum that is at a higher latitude in the east pacific as compared to the west pacific.

During the warm phase of ENSO this low is intensified and shifts to the east as the isobars flatten out to be nearly directly east-west (see Fig. 12. Again, this is the result of shifting the thermal maximum to be directly at the equator for all longitudes. The intensification is in agreement with more intense westerlies as a result of an intensified Hadley circulation supplying more angular momentum to the midlatitudes. In his original paper, Bjerknes (1966) also argues that the anomalous Aleutian Low is the beginning of a wave-train that weakens the Icelandic low and displaces it from the southwest of the island to the east of the island.

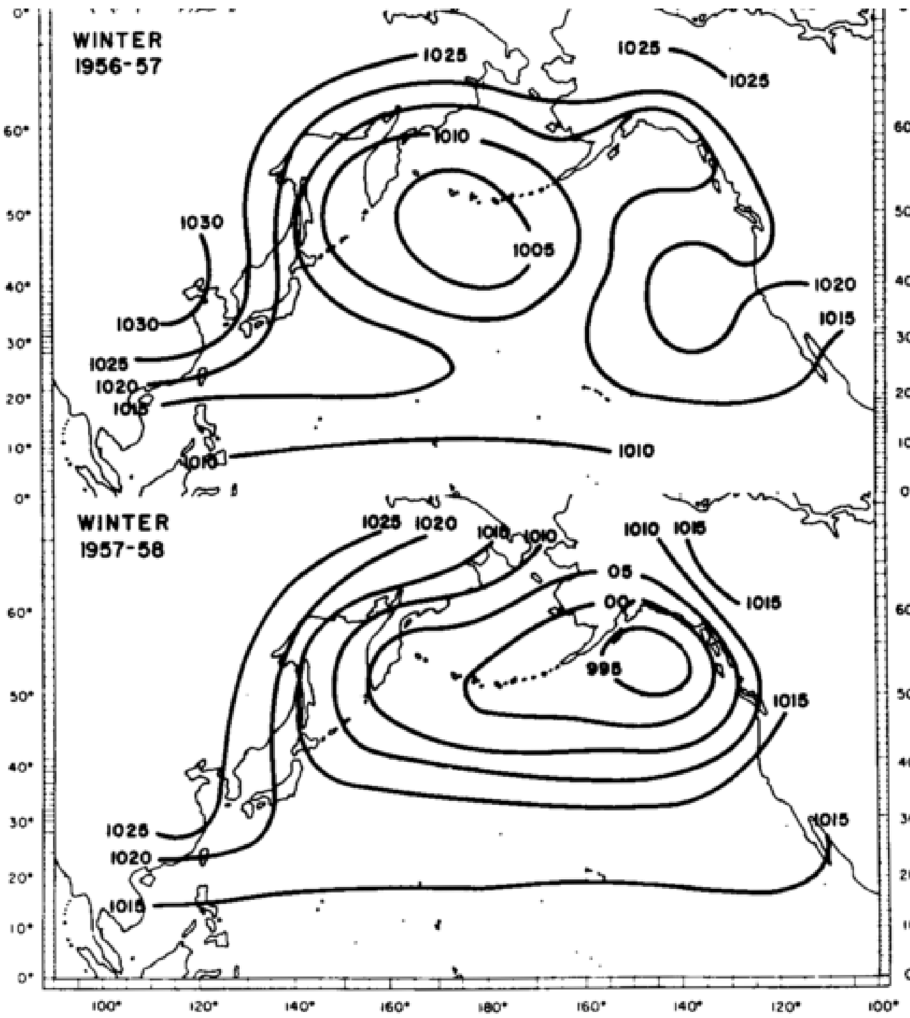


Figure 12: Credit: Figure 6 of Bjerknes (1966). Winter (DJF) sea level pressure for (top panel) the year prior to an El Niño and (bottom panel) during the peak of El Niño.

Another (perhaps more common) way of looking at these variations is as anomalies to the mean field. For a complete conceptual understanding using these fields, however, you will need to mentally add the anomalies (see panel (a) of Fig 13) to the mean field (see panel (a) of 7). Using this conceptual framework, the warm state of ENSO creates a wave train of an anomalous low in the Gulf of Alaska, an anomalous high over western Canada and an anomalous low over the southeast US. Chen and Wallace (2015) also identify anomalies associated with the decadal ENSO-like state in panel (b), which resemble intensified ENSO teleconnections.

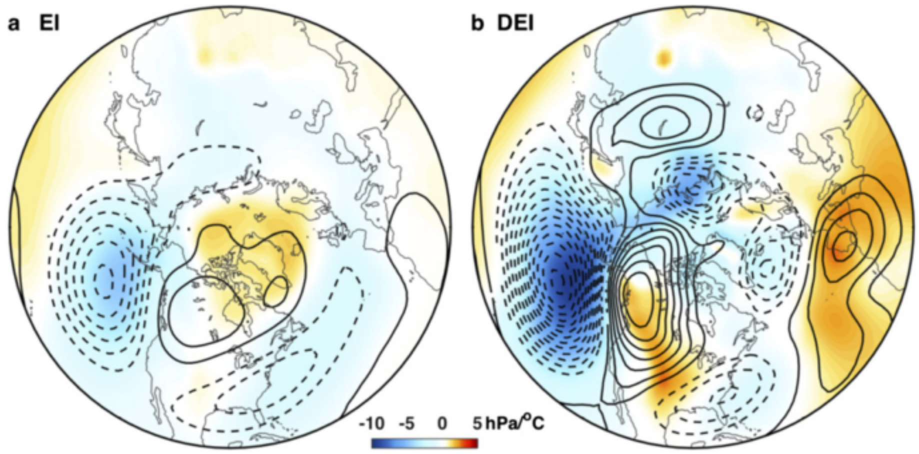


FIG. 15. Wintertime (November–March) SLP (colored shading) and 500-hPa height (contours) regressed upon the (a) EI and (b) DEI, based on data from the 1919/20 winter through the 2013/14 winter. The contour interval for 500-hPa height is 10 m (°C)$^{-1}$ of the CTI.

Figure 13: Credit: Figure 15 of Chen and Wallace (2015). ENSO teleconnections, found by regressing (a) an Index of ENSO and (b) a decadal ENSO index onto winter (Nov-Mar) sea level pressure (color shading) and 500 hPa height (contours).

5 Transition from El Niño to La Niña

ENSO events tend to develop and decay in a characteristic fashion, such that all La Niñas develop following El Niños, but not all El Niños decay into La Niñas. So the question arises, what drives the transition from warm to cold states in the system? There are four distinct mechanisms that have been proposed as being important: 1) Wave reflection at the western boundary, 2) wind-forced Kelvin waves in the western Pacific, 3) discharge of warm water due to Sverdrup transport and 4) Anomalous zonal advection. These mechanisms may each have varying degrees of influence in different scenarios, on different timescales or depending on who you ask. The bottom line is that we don't fully understand the system, although we do have some working hypotheses. The four mechanisms are outlined below

5.1 Delayed Oscillator (wave reflection)

This theory invokes wave reflection to explain the transition from El Niño to La Niña. Equatorial Rossby and Kelvin waves are generated where there is a coupling between the ocean and the atmosphere (see Figures 14 and 15). The Rossby waves propagate west, reflect off the western boundary and return east as equatorial Kelvin waves, which reverse the anomalies in the east Pacific and complete the transition from warm to cold events. Note that directly-forced (downwelling) Kelvin waves and reflected (upwelling) Kelvin waves have opposite signed influences on the thermocline, which is why the reflected-Kelvin waves act as a negative feedback to the system. Figure **??** a Hovmoller diagram of a eastward propagating, directly forced, downwelling Kelvin wave (shown in red) and a subsequent, weaker, upwelling Kelvin wave from 2014. Kelvin waves generally take 2-3 months to cross the Pacific ocean.

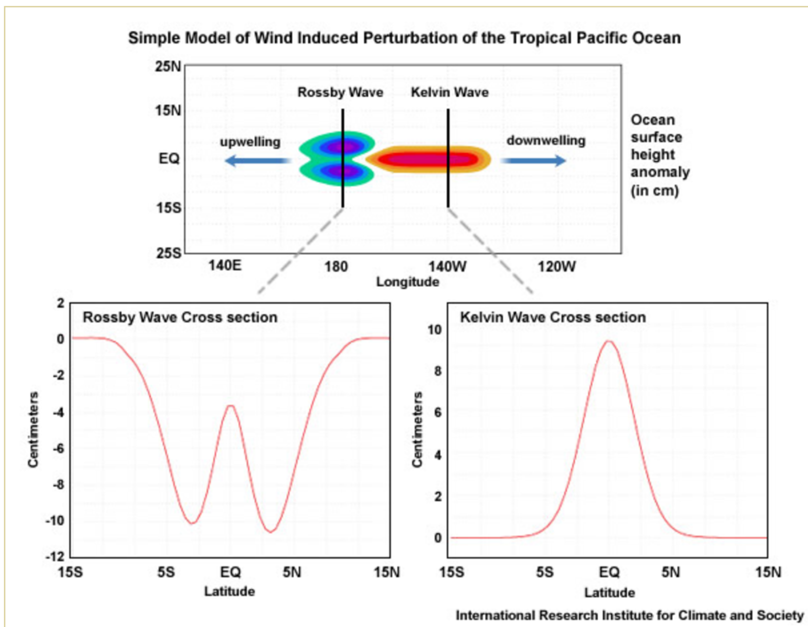


Fig. 4.29. Idealized model of a single equatorial eastward-moving Kelvin wave generated by wind stress anomaly (red and orange) and corresponding Rossby waves propagating westward.

Figure 14: Credit:

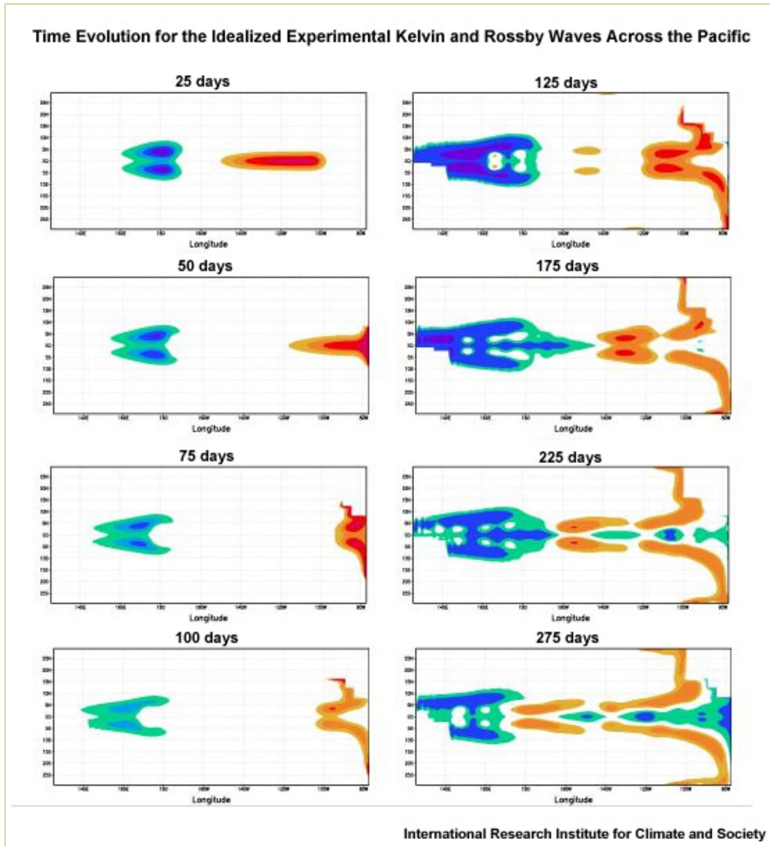
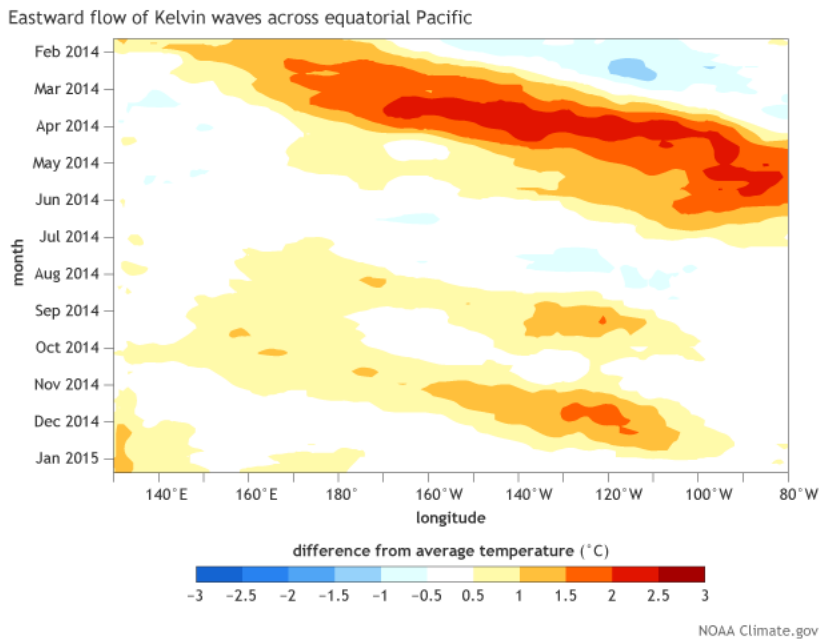


Fig. 4.30. Time evolution for the idealized experimental waves generated in Fig. 4.29 (IRI).

Figure 15: Credit:



Equatorial sub-surface temperature anomalies (averaged from 0-300 meters depth) shown for each longitude across the Pacific Ocean (see x-axis). Time is shown on the y-axis from mid-January 2014 (top edge) to mid January 2015 (bottom edge). Red shading shows where temperatures were above-average, and blue shading shows where they were below average. Data is from the NCEP Global Ocean Data Assimilation System (GODAS) with anomalies defined with respect to the average over 1981-2010. Figure is from NOAA Climate Prediction Center.

Figure 16: Credit: Figure from NOAA: https://www.climate.gov/news-features/blogs/enso/oceanic-kelvin-waves-next-polar-vortex.

While all versions of this theory acknowledge that wave reflection from the western boundary is important for terminating El Niños, there is debate about whether it is only the gravest Rossby mode that is important or whether off-equatorial (i.e. poleward of about 8°) Rossby wave reflection is also important. While this theory as a whole seems simple, in practice determining the reflection efficiency of the western boundary is complicated due to the paucity of observations, the complex topography and highly coupled nature of the west Pacific.

5.2 Wind-forced Kelvin waves in the western Pacific

A theory that doesn't explicitly rely on wave reflection is the western Pacific oscillator. This theory instead invokes a Gill-type circulation anomaly to explain the eastward-propagating Kelvin waves that help to dissipated the event. Below is a diagram of the Gill atmospheric circulation response to a heating in the tropical Pacific.

The Western Pacific Oscillator

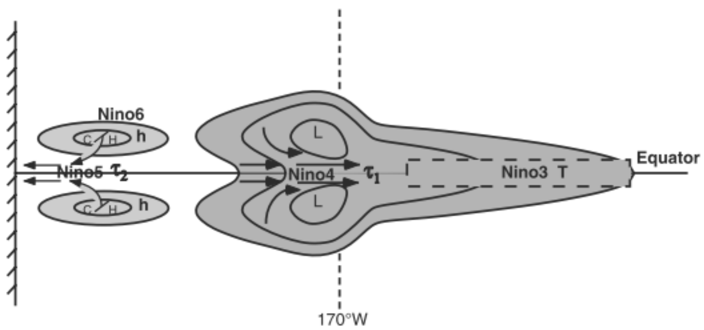


Figure 4. Schematic diagram of the western Pacific oscillator for ENSO. Condensational heating in the central Pacific induces a pair of off-equatorial cyclones with westerly wind anomalies in the Niño4 region. The Niño4 westerly wind anomalies act to deepen the thermocline and increase SST in the Niño3 region. On the other hand, the off-equatorial cyclones raise the thermocline there via Ekman pumping. Thus, a shallow off-equatorial thermocline anomaly expands over the western Pacific leading to a decrease in SST and an increase in SLP in the Niño6 region. During the mature phase of El Niño, the Niño6 anomalous anticyclone initiates equatorial easterly wind anomalies in the Niño5 region. The Niño5 easterly wind anomalies cause upwelling and cooling that proceed eastward as a forced Kelvin wave response providing a negative feedback for the coupled system to oscillate.

Figure 17: Credit: Figure 4 of Wang and Picaut (2004).

Rather than wave reflection or Sverdrup balance, this theory primarily invokes Ekman drift and Ekman pumping associated with the cyclonic or anticyclonic Gill anomalies. On the equator there are westerly wind anomalies that have equatorward Ekman drift, which acts to deepen the thermocline in the east Pacific through convergence. But in the center of the off-equatorial cyclones there is Ekman pumping due to divergence, which shoals the thermocline just off the equator. These cold anomalies expand over the course of an El Niño, lead to anomalous easterlies that reinforce the climatological trade winds and therefore act as a negative feedback for the system, which could move it from a warm state into a cold state.

5.3 Discharge of warm water due to Sverdrup transport

One prominent idea that dates back to (Wyrtki, 1975) is that a buildup of heat in the tropical Pacific is a precursor to El Niño events, and that the subsequent discharge of this heat during an El Niño triggers the transition to La Niña. Observational analyses have demonstrated that the total upper ocean heat content between 5N and 5S leads Niño3 SST anomalies by about two seasons. The physical processes proposed are as follows: During an El Niño westerly winds at the equator cause warm waters to be advected east. Sverdrup balance is the balance between the wind stress curl and meridional flow:

$$\beta v = \frac{1}{\rho_0 h} \nabla \times \tau$$

where β is the poleward variation of the Coriolis parameter (sometimes called the Rossby parameter), h is the sea surface height and τ is the wind stress. By applying this equation to the zonal-averaged wind stress, we can see the direction of Sverdrup transport as a function of latitude. The figure below illustrates the zonal mean wind, the associated Sverdrup flow (small arrows; a scalar function of wind stress curl, or the first derivative of wind velocities), the western boundary currents (large arrows) and the sign of the vorticity imparted by the wind stress curl in the clockwise or counter-clockwise arrows.

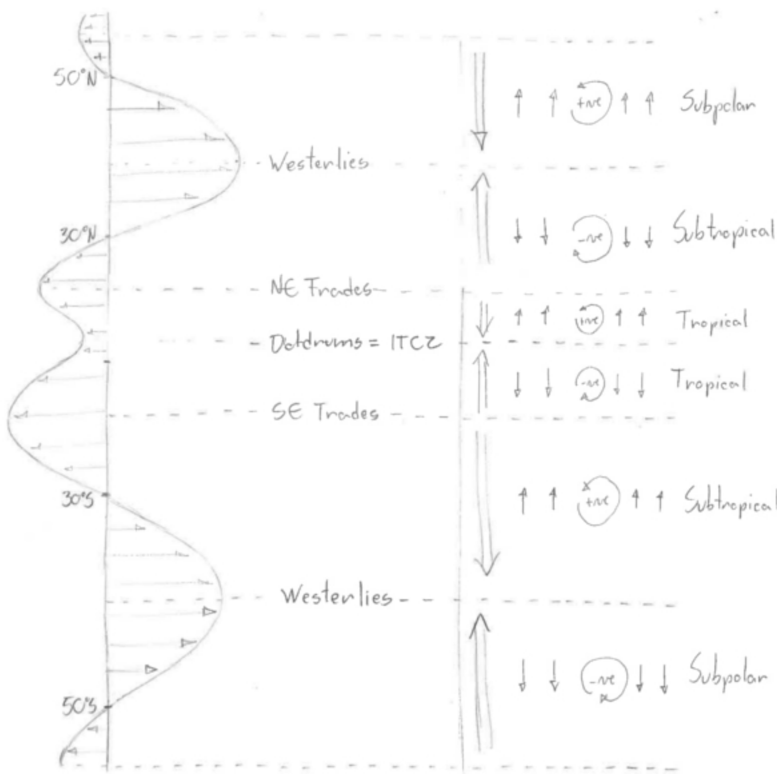


Figure 18: Courtesy of the notes of Dr. Lisa Beal, Lecture 15. http://www.rsmas.miami.edu/users/lbeal/MPO603/

This figure demonstrates that in the deep tropics Sverdrup flow is equatorward, but would be reversed by westerly wind bursts right at the equator. How this Sverdrup transport may act to bring warm water into or out of the deep tropics as a function of modifying the wind stress curl is demonstrated in the figure below.

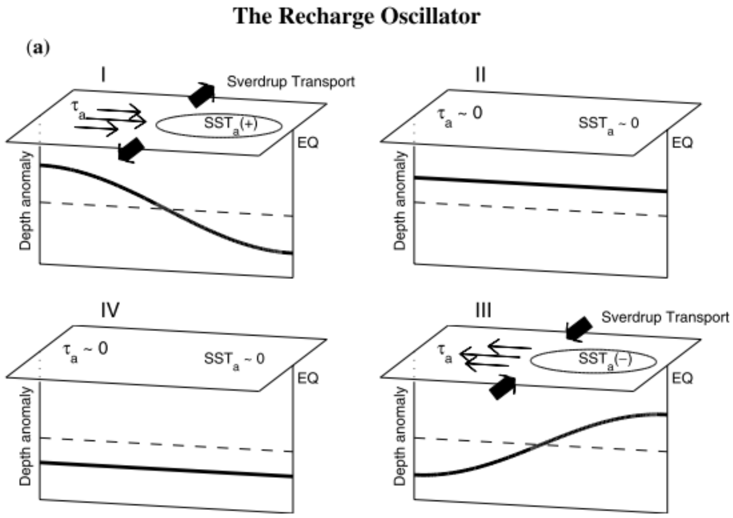


Figure 19: Credit: Figure 3 of Wang and Picaut (2004). The panels represent (I) El Nño conditions (II) transition period between warm to cold events, (III) La Niña conditions (IV) post-La Niña conditions

5.4 Anomalous zonal advection

This conceptual model focuses on the zonal currents. During a developing El Niño the western Pacific warm pool is advected eastward through anomalous zonal currents and positive feedbacks. There are three negative feedback processes that act to reverse this process: 1) an anomalous zonal current resulting from Kelvin waves reflecting on the eastern boundary of the Pacific, 2) an anomalous zonal current from Rossby waves reflecting off the western boundary of the Pacific, and 3) the mean zonal current converging at the eastern edge of the warm pool. The first two mechanisms can be understood as follows. Wind-forced equatorial Kelvin waves propagate eastward and reflect off the eastern boundary to become a downwelling Rossby wave. The same westerly-wind anomalies force Rossby waves, which reflect off the western boundary to become upwelling Kelvin waves. Both of these reflected waves are associated with a westward zonal current, which 'pushes' the warm pool back towards the west Pacific. Overlaid on the mean zonal current, these negative feedbacks are sufficient to move the system from a warm state to a cold state.

5.5 A unified view

Here we will reiterate that these mechanisms aren't mutually exclusive, and nothing constrains the system from behaving differently in different circumstances. As such, we can complete our conceptual model by combining all four of the above mechanisms into a single model, as illustrated below. Here we have a simultaneous illustration of 1) waver reflection at the western boundary, 2) wind forced equatorial Kelvin waves, 3) discharge / recharge of oceanic equatorial heat content due to Sverdrup transport and 4) anomalous zonal advection.

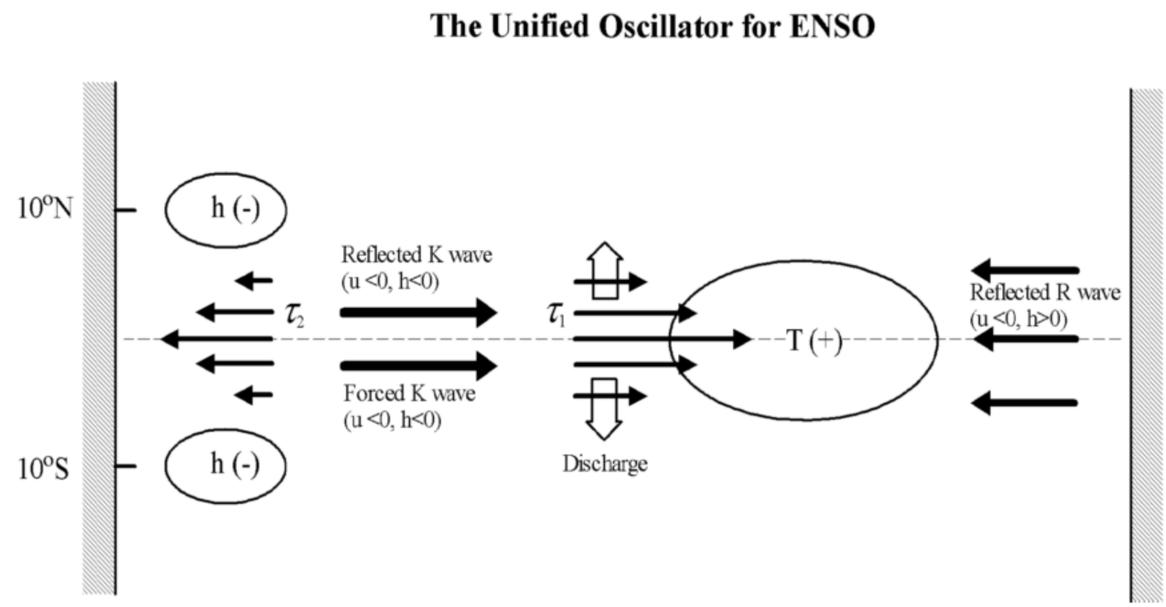


Figure 1. Schematic diagram of the unified oscillator for ENSO. Bjerknes positive ocean-atmosphere feedback leads the equatorial central/eastern Pacific to a warm state (El Niño). Four negative feedbacks, required to turn the warm state around, are (1) reflected Kelvin wave at the ocean western boundary, (2) discharge process due to Sverdrup transport, (3) western Pacific wind-forced Kelvin wave, and (4) reflected Rossby wave at the ocean eastern boundary. These negative feedbacks correspond to the delayed oscillator, the recharge oscillator, the western Pacific oscillator, and the advective-reflective oscillator. The unified oscillator suggests that all of the four negative feedbacks may work together in terminating El Niño warming. The four ENSO oscillators are special cases of the unified oscillator.

Figure 20: Credit: Figure 1 of Wang and Picaut (2004).

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