OCEAN-ATMOSPHERE INTERACTION AND TROPICAL CLIMATE

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Contents

Abstract. The ocean and atmosphere are in constant exchange of heat, water, and momentum. The interaction of the ocean and atmosphere adds shades and rhythms in the structure of tropical climate. Ocean-atmosphere interaction research has experienced rapid growth in studying El Nino/Southern Oscillation (ENSO), and yielded tremendous benefits by enabling and improving the prediction of ENSO and important modes of climate variability. This chapter reviews major ocean-atmospheric feedbacks that give rise to ENSO and other variations in tropical climate.

1. Introduction

Solar radiation is the ultimate source of energy for motions in the atmosphere and ocean. Most absorption of solar radiation takes place on the Earth surface, the majority of which is occupied by oceans. Thus oceanic conditions, sea surface temperature (SST) in particular, are important for atmospheric temperature conditions and circulation. Fueled by water vapor evaporated from the surface, deep convection in cumulonimbus clouds and resultant condensation and freezing are the dominant mechanism for heating the atmosphere. Atmospheric convection is strongly regulated by SST on the one hand and affects the ocean on the other by modulating surface momentum and heat fluxes. Latent heating in atmospheric convection drives surface winds and modulates cloud cover. Surface winds drive ocean circulation and affect ocean surface heat flux while clouds modulate surface radiative flux.

Thus, the ocean and atmosphere are a coupled system and their interaction helps shape tropical climate and its variability. Examples are abundant. Figure 1a shows the climatology of precipitation and SST. While the annual-mean solar radiation at the top of the atmosphere (TOA) is zonally uniform and symmetric about the equator, rainfall and SST are highly asymmetric both in the east-west and north-south directions. On the equator, SST in the eastern Pacific (say, near the Galapagos Islands, 90^{o}W, equator) features a pronounced annual cycle with the maximum in March (Fig. 2) despite a TOA solar radiation dominated by a semi-annual cycle. Such departures in space and time from the solar radiation distribution are the result of ocean-atmosphere interaction.

This Topic concerns major ocean-atmospheric feedbacks important for spatial and temporal variations of tropical climate. They involve changes in cloud cover, surface evaporation, and ocean dynamical adjustments. The scope of this review is limited to ocean-atmosphere interaction operating on large (>100 km) spatial and long (> 1 week) temporal scales. While variability on weather and shorter timescales and beyond the instrumental record is not covered, the feedback mechanisms discussed here are expected to operate for paleoclimate variability (Chiang et al. 2003; Timmermann et al. 2007) and future climate change (Vecchi and Soden 2007). For a more comprehensive discussion of ocean-atmosphere interaction and its applications, readers are referred to a recent monograph on this topic (Wang et al. 2004).

2. Bjerknes feedback

Strong east-west asymmetry is found over the equatorial Pacific, characterized by the Walker circulation (with the easterly winds at the surface) in the atmosphere and the cold tongue in SST (Fig. 1a). To a lesser extent, similar asymmetry is observed over the equatorial Atlantic.

Under easterly winds, surface ocean currents flow poleward a few degrees away from the equator following the Ekman dynamics. The divergence of these poleward Ekman flows requires water to upwell into the surface layer on the equator. If the thermocline is close to the surface, this equatorial upwelling brings the cold thermocline water into the mixed layer, causing an SST cooling. On the equator where the Coriolis force vanishes, surface ocean currents flow in the wind direction, resulting in the westward South Equatorial Current (SEC) in the Pacific and Atlantic. The equatorial upwelling and the westward (SEC) shoals the thermocline in the east and deepens it in the west (Fig. 1b). In steady state on the equator, the easterly wind stress is nearly balanced by the pressure gradient force associated with the eastward shoaling of the thermocline.

From an oceanographic point of the view, the easterly winds shoal the thermocline eastward and induce the equatorial upwelling, keeping the eastern basin cool. From a meteorological point of the view, on the other hand, the eastward SST cooling limits deep convection to the west and maintains a sea pressure gradient that drives the easterly winds along the equator. This circular argument indicates that ocean-atmosphere interaction is at the heart of the cold tongue formation.

The feedback may be described as follows (Fig. 3). Let us begin with modest easterly winds over the equatorial Pacific, which induce upwelling and tilt the thermocline, both acting to cool the eastern ocean and suppressing atmospheric convection there. This reduction in deep convection in the east raises sea level pressure, intensifying the initial easterly winds at the surface. The amplification of the initial perturbation indicates a positive feedback resulting from ocean-atmosphere interaction.

Bjerknes (1969) first proposed this feedback hypothesis for El Nino/the Southern Oscillation (ENSO), to which Wyrtki (1975) added the thermocline adjustment. Therefore it is also called the Bjerknes-Wyrtki feedback. The thermocline depth in the eastern Pacific controls how much equatorial upwelling cools SST, and is determined by the transport of warm upper-ocean water above the thermocline between the east and west, and in and out of the equatorial belt. Such warm water transport, governed by ocean wave dynamics, is essential for ENSO (see Topic xx in this Theme). In coupled ocean-atmosphere models, the most unstable or least damped mode often displays an interannual cycle of SST warming and cooling in the eastern basin, with the deepening and shoaling of the thermocline, respectively. Besides ENSO, such a Bjerknes mode of coupled ocean-atmospheric variability is observed in the equatorial Atlantic and Indian Oceans on interannual timescales (Chang et al. 2006).

The equatorial cold tongue displays annual expansion and retreat in July-September and February-April (Fig. 2), respectively, over the Pacific and Atlantic. This annual cycle in SST is forced by that in meridional wind superimposed on the annual-mean southerly cross-equatorial winds, the latter being part of meridional climatic asymmetry. During July-September, the southerlies intensify in response to the seasonal warming (cooling) in the Northern (Southern) Hemisphere, enhancing the upwelling cooling in the equatorial belt (Mitchell and Wallace 1992). Conversely, the relaxed southerlies cause equatorial SST to warm up during February-April. Unlike ENSO, thermocline adjustments are of a secondary importance for the annual cycle but the interaction of SST and surface winds causes a westward phase propagation of the annual cycle.

3. Wind-evaporation-SST (WES) feedback

Evaporation is the major means for tropical oceans to balance the net downward radiative flux at the surface. Surface evaporation is a function of SST and atmospheric variables such as wind speed. The interaction of surface wind and SST via surface evaporation yields a positive feedback as follows. Consider a meridional dipole of SST perturbations, positive north and negative south of the equator (Fig. 4). The SST-induced pressure anomalies drive a southerly cross-equatorial wind, which gains an easterly south and westerly component north of the equator because of the Coriolis force. Superimposed on the prevailing easterly trades, these wind anomalies increase (decrease) the wind speed south (north) of the equator, intensifying (reducing) evaporative cooling there. This dipole of latent heat flux anomalies acts to amplify the initial SST dipole. Appendix presents a simple mathematical model for the WES interaction. Xie and Philander (1994) first proposed this wind-evaporation-SST feedback (WES) feedback to explain the pronounced climatic asymmetry of the northward-displaced rain bands over the eastern Pacific and Atlantic.

Over the eastern tropical Pacific and Atlantic, the northeast and southeast trades converge onto not the equator but to the north (Fig. 1a). This northward displaced intertropical convergence zone (ITCZ) is collocated with a basin-scale rain band, both anchored by a zonal band of high SST. Consistent with the WES feedback, the easterlies and scalar wind speed are much weaker in the northern high-SST band than on the other side of the equator. While helping break the equatorial asymmetry, the WES feedback does not favor either hemisphere. Asymmetry in continental geometry, such as the coastline and mountains, has been demonstrated to initiate the development of climatic asymmetry. For example, in response to the prevailing easterlies, a northwestward tilted coastline (such as the west coast of the American continents) induces upwelling south and downwelling north of the equator. Though initially trapped near the coast, this asymmetry in upwelling triggers the WES feedback, displacing the ITCZ north of the equator across the basin.

The WES feedback is found to be important in forming the so-called meridional mode in tropical oceans, a pattern of interannual SST anomalies with subtropical centers of action and opposite signs across the equator (Chang et al. 2006). Over the tropical Atlantic in particular, the meridional mode is associated with the anomalous displacement of the ITCZ into the warmer hemisphere, causing floods or droughts in the semi-arid region of northeast Brazil.

The WES feedback can cause meridional propagation. Consider a positive SST anomalies centered off the equator. The SST heating of the atmosphere generates a low pressure, with westerly (easterly) geostrophic winds to the south (north) of the SST warming. Under the mean easterlies, these zonal wind anomalies act to weaken (strengthen) wind speed and warm (cool) SST south (north) of the SST maximum via latent heat flux, causing the initial SST anomalies to propagate toward the equator. This equatorward propagation due to WES feedback helps subtropical wind variability to influence the deep tropics including the equator, in the so-called fingerprinting mechanism (Vimont et al. 2003).

Under the westerly mean flow, the above argument indicates a poleward phase propagation due to the WES feedback. Such conditions are met in the Asian summer monsoon from the North Indian Ocean to Northwest Pacific, where atmospheric intraseasonal oscillation indeed displays a pronounced northward propagation, along with significant SST co-variability (Fu et al. 2003). Such northward-propagating intraseasonal oscillations are important for the subseasonal, active-break cycle of the Indian summer monsoon.

4. Cloud feedback

Much of tropical rainfall is confined to regions of SST greater than 27^{o}C (Fig. 1a), indicating that SST regulates strongly deep convection of the atmosphere. Over the tropics, SST-precipitation is positively correlated but the relationship is highly nonlinear, with the correlation strong in the SST range of 26-28^{o}C but weak for lower and higher SSTs. In climatological rain bands, cumulonimbus clouds with extensive anvils (collectively convective clouds hereafter) prevail. Thus, convective cloud-SST feedback is negative over warm ocean surface: an increase in SST intensifies convection and the increased cloud cover feeds back to dampen the initial SST warming (Fig. 5a).

Over subtropical oceans of SST below the 26-27^{o}C convective threshold and under the slow subsidence of the Hadley circulation, low clouds capped by a temperature inversion prevail. On a flight from Los Angeles to Hawaii, the following cloud regime transition is typically observed: solid stratus cloud decks prevail immediately off the coast of California; then the cloud decks begin to break up into stratocumulus with significant clear regions embedded; by the time the plane approaches Hawaii, the ocean is dotted with fair-weather cumulus clouds of small cloudiness. This cloud regime transition and the attendant decrease in cloudiness are due to the increase in SST along the trajectory of an air parcel riding on the northeast trades. The SST increase enhances surface turbulent heat flux, promoting shallow convection below the inversion and creating subsidence that dissipates cloud droplets.

Of high liquid water content, low-level clouds are highly reflective of solar radiation. Over extensive subtropical oceans, interannual anomalies of low cloud cover is often negatively correlated with the underlying SST (Fig. 5b; Klein and Hartmann 1993), indicative of a positive feedback: an SST cooling increases low-cloud cover, reducing solar radiation at the sea surface and amplifying the initial cooling. Subtropical low clouds are difficult to represent in state-of-art climate models, resulting in major errors in simulations, such as the failure to keep the Pacific and Atlantic ITCZ north of the equator. Low cloud changes in response to the increase in greenhouse gases are a major uncertainty in model projections of future climate.

Cloud-SST feedback may be positive or negative depending on the cloud type. Over warm oceans with deep convection, the convective cloud-SST feedback is negative while over cool subtropical oceans, the low cloud-SST feedback is positive.

5. Water vapor feedback

With an increase in SST, the ocean surface emits more upward longwave radiation, following the Stephan-Boltzman law. Meanwhile the SST increase raises water vapor content in the atmosphere, causing an increase in downward atmospheric infrared radiation. At the sea surface, the water vapor-induced increase in the downward longwave radiation is usually much larger than the Stephan-Boltzman increase in the upward radiation, helping amplify the initial SST warming. The water-vapor feedback is the major mechanism for amplifying the global warming induced by increased greenhouse gas concentrations in the atmosphere (Hall and Manabe 1999)

6. Ocean front-atmosphere interaction

Over much of the open ocean, surface air temperature (SAT) is close to the underlying SST. Near ocean fronts, however, surface air-sea temperature difference displays large cross-frontal variations (Fig. 6). This happens because of the spatial scale of ocean fronts is much smaller than that of the atmosphere. SAT adjusts to the sharp SST front by cross-frontal advection and/or the geostrophic adjustment. As a result, SAT is often warmer (colder) than SST on the colder (warmer) flank of the SST front. On the warmer flank, the reduced static stability strengthens the vertical mixing, bringing faster winds from aloft and accelerating surface winds. For the same reason, the surface winds decelerate on the colder flank of the ocean front. Thus the ocean front-atmosphere interaction is characterized by a positive correlation between SST and wind speed. Recent studies suggest that atmospheric pressure variations caused by SST effects on atmospheric boundary layer temperature are also important in driving coherent wind patterns near ocean fronts.

Recent high-resolution satellite observations reveal ubiquitous co-variability of SST, surface wind velocity, and clouds along major fronts of the ocean, with positive correlation between SST and wind speed (Fig. 7; Small et al. 2008). In the tropics, such ocean-atmospheric co-variability is observed along the sharp SST front on the northern flank of the equatorial cold tongue in both the Pacific and Atlantic. Lateral shears of major equatorial currents sustain tropical instability waves that cause the equatorial front to meander. The interaction of the ocean front and atmosphere constitutes a negative feedback on the ocean thermodynamically. On the warmer (colder) flank of the front, the increase (decrease) in both wind speed and sea-air temperature intensifies turbulent heat flux from the ocean, acting to dampen the original SST perturbation.

7. Summary

Ocean-atmospheric feedbacks described above may be divided into two types: the Bjerknes feedback involves ocean dynamical adjustments while the WES and cloud/water vapor-SST feedbacks involves surface heat flux and are often called thermodynamical feedbacks. The Bjerknes feedback operates in the east-west direction within the equatorial belt where the Coriolis force vanishes. (Studies indicate that the Bjerknes feedback becomes negative away from the equator where geostrophy dominates ocean and atmospheric flow.) While it may operate in the zonal direction, the WES feedback breaks equatorial symmetry, creating its signature dipole pattern of coupled ocean-atmospheric anomalies in the meridional direction. The ocean's interaction with clouds and water vapor, by contrast, acts locally and does not generate characteristics spatial patterns. (It may, in combination with the Bjerknes and/or WES feedback.)

Rich structures of tropical climate owe much to ocean-atmosphere interactions, without which the Earth climate would be much more uniform in space and constant in time. Owing to the Bjerknes feedback, Indonesia on the western side of the Pacific is covered by rainforests while the Galapagos in the east is arid with major precipitation limited to a brief warm season centered in March. Because of the WES feedback aided by low cloud-SST interaction, Panama north of the equator sees towering cumulonimbus with convective rain year round while the Pacific coast of Peru to the south is among the driest places on Earth. The barren Galapagos and Peruvian coast are blessed with heavy rainfall once every few years during El Nino events, a result of the Bjerknes feedback.

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Appendix. Coupled ocean-atmosphere model

We present a simple model for the WES feedback to illustrate the concept of ocean-atmosphere coupling. Surface latent heat flux (E) is often calculated using an empirical bulk formula: $E = L\rho_a C_E |U| [q_s(T) - RHq_s(T_a)]$, where L is the latent heat of evaporation, ρ_a the air density, C_E the transfer coefficient, |U| is the wind speed (for simplicity, we consider only the zonal wind), q_s is the saturated specific humidity following the Clausius-Clayperon equation, RH the relative humidity, T and T_a are SST and SAT, respectively. Consider a simple case where the net radiation and ocean advection are set constant. Since SAT anomalies generally follow SST because of the large heat content of the ocean, we set $T'_a = T'$ so that sensible heat flux vanishes. The governing equation for perturbation SST may be cast as $\frac{\partial T'}{\partial t} = \frac{E'}{\rho c_p H}$, where ρ is the sea water density, c_p the specific heat at constant pressure, and H the mixed layer depth.

Under the prevailing background wind $\overline{U}$, scalar wind speed is cast as $\left|\overline{U}+U'\right| = \mathrm{sgn}(\overline{U})(\overline{U}+U')$, where the overbar and prime denote the mean and perturbation, respectively. Linearizing the latent heat flux formula yields

$$\frac{\partial T}{\partial t} = aU - bT , \qquad (1)$$

where a=sgn(- $\overline{U}$) $\overline{E}$ /($\rho c_p H$), and b= $\overline{E} L/(R_v \overline{T}^2)$ with R_v the gas constant for water vapor. Note that we have omitted the prime for clarity. Under the prevailing easterly trades ($\overline{U}$ <0), the coupling coefficient in the ocean model, *a*, is positive: a westerly wind perturbation reduces surface wind speed and evaporation, causing SST to rise.

Consider two strips north and south of the equator, whose SST is T_N and T_S, respectively. The perturbation cross-equatorial wind may be modeled as

$$V = \alpha(T_N - T_S) , \qquad (2)$$

where α is a positive coefficient. Off the equator, the meridional wind induces a zonal wind component due to the Coriolis force,

$$\delta U = fV / \varepsilon = (f\alpha/\varepsilon)\delta T, \qquad (3)$$

where δ denotes the north-south difference, f is the Coriolis parameter in the northern strip and ε the friction coefficient.

Eqs. (1) and (3) constitute a simple *coupled model*, from which we can obtain the equation for the north-south SST difference

$$\frac{\partial}{\partial t}\delta T = (\sigma - b)\delta T ,$$

where σ = f (aα)/ε is the WES *coupling coefficient*. Without the coupling, the initial SST dipole would decay at the rate of b. With the WES feedback, the decay is slowed down at the rate of (b-σ). If the WES is strong enough (σ>b), the dipole can overcome the Newtonian cooling (b) and grows in time.

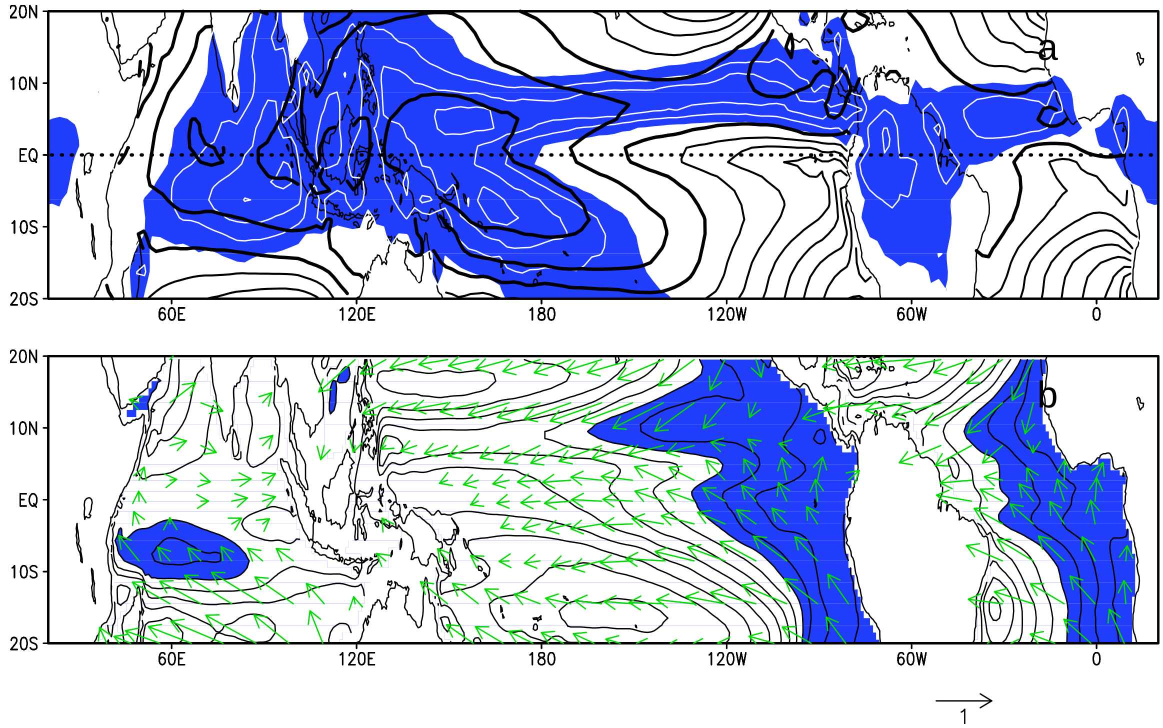

Figure 1. Annual-mean climatology: (a) SST (black contours at 1^{o}C intervals; contours of SST greater than 27^{o}C thickened) and precipitation (white contours at 2 mm/day; shade > 4 mm/day); (b) surface wind stress vectors (Nm^{-2}) and the 20^{o}C isotherm depth (contours at 20 m intervals; shade < 100 m).

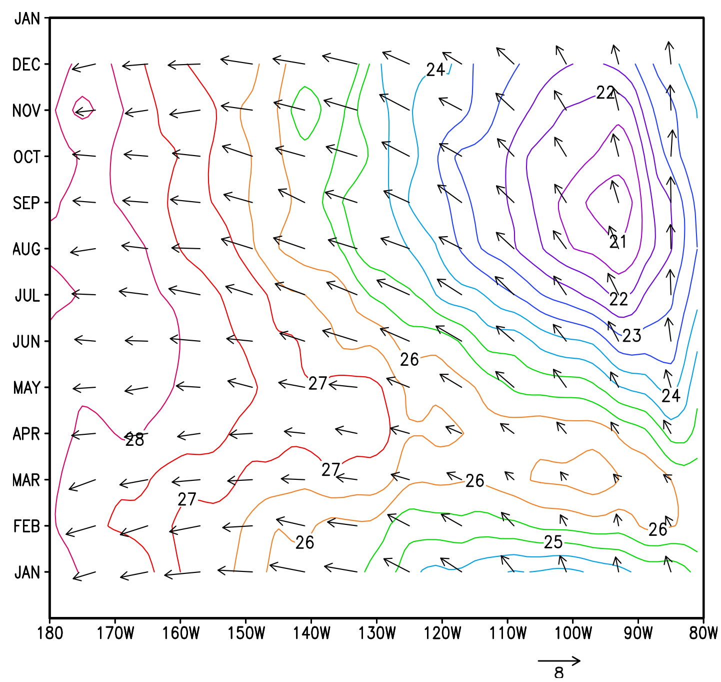

Figure 2. Seasonal cycle in the equatorial Pacific: SST (contours in ^{o}C) and surface wind velocity (vectors in m/s).

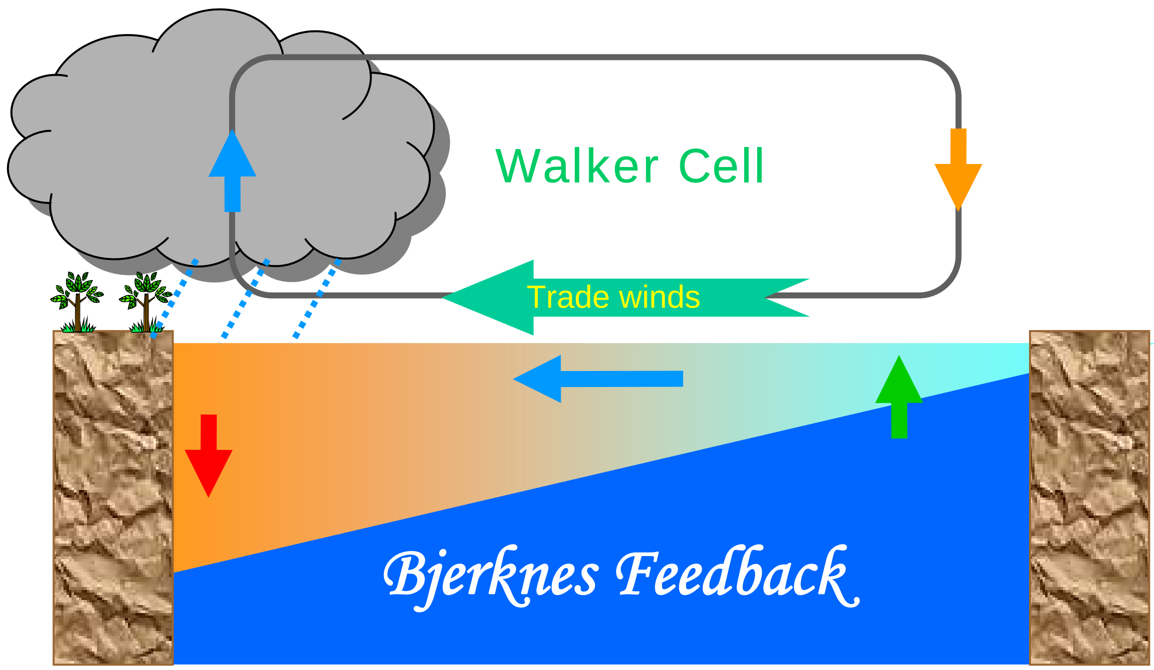

Figure 3. Schematic for Bjerknes feedback.

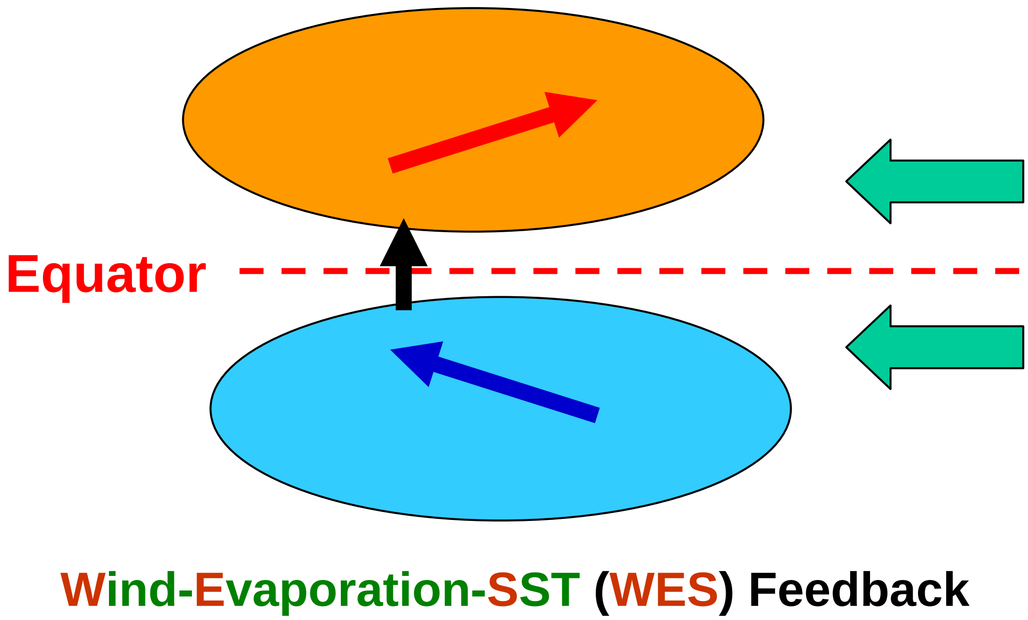

Figure 4. Schematic for wind-evaporation-SST feedback.

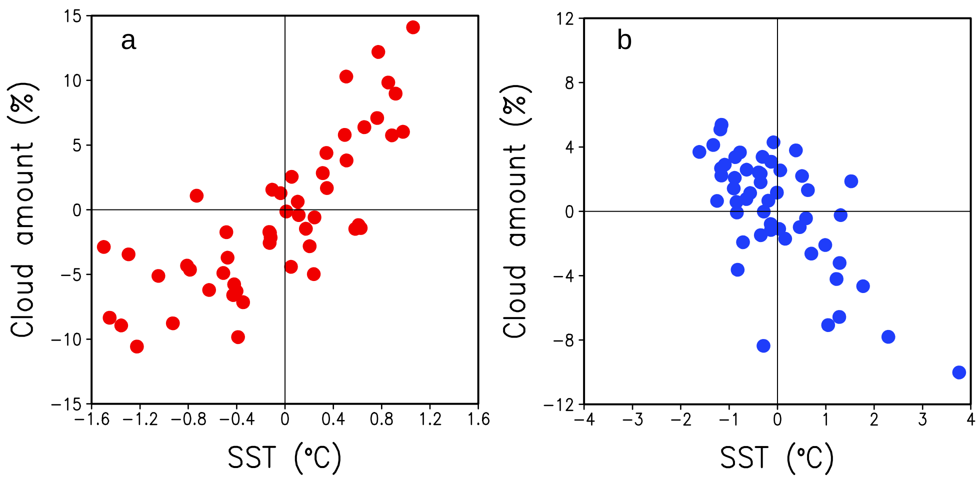

Figure 5. Scatter diagrams between SST and cloudiness for (a) the central equatorial Pacific (150^{o}E – 140^{o}W, 5^{o}S-5^{o}N) and October-January; and (b) the southeastern tropical Pacific (100 – 80^{o}W, 10^{o}S-0) and June-November, based on ship observations for 1950-2000. Deep convective and low clouds prevail in the respective regions. Correlations are 0.79 for (a) and -0.69 for (b).

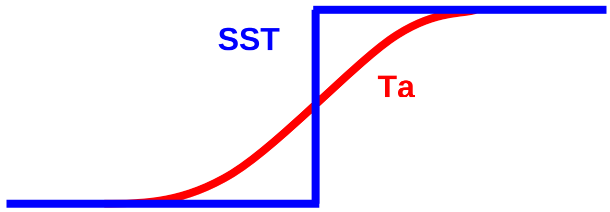

Figure 6. Schematic for surface air temperature (Ta; red) adjustment to a sharp SST front (blue).

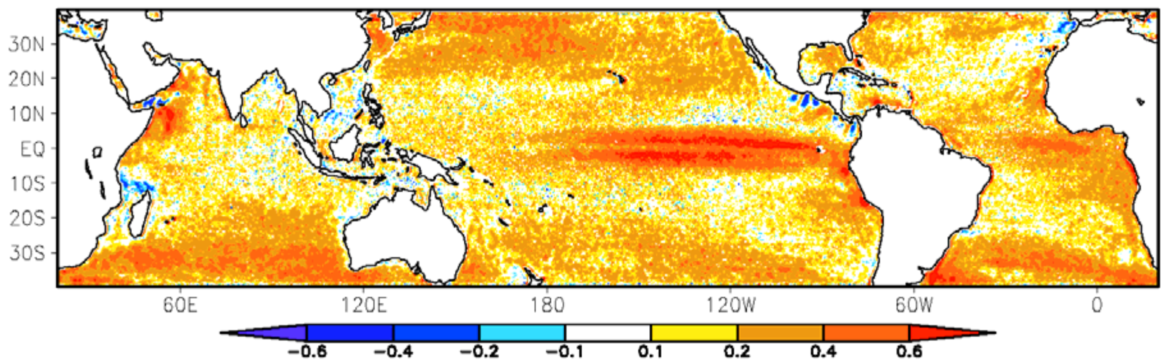

Figure 7. SST-wind speed correlation based on measurements by the Tropical Rain Measuring Mission satellite. Data has been filtered to extract variations of zonal scales shorter than 1000 km. The correlation is predominantly positive along major ocean fronts.