Effects of Seasonal Solar Forcing on Latitudinal Asymmetry of the ITCZ

SHANG-PING XIE

Graduate School of Environmental Earth Science, Hokkaido University, Sapporo, Japan

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ABSTRACT

A coupled ocean–atmosphere model is used to investigate the effects of seasonal variation in solar radiation on the configuration of the intertropical convergence zone. The model maintains a Northern Hemispheric ITCZ under annual mean insolation, with convection being suppressed in the Southern Hemisphere. In the presence of seasonal variations, a Southern Hemispheric ITCZ develops in boreal winter and spring in response to the seasonal rise in local solar radiation. As a result, the equatorial asymmetry of the annual-mean model climatology is reduced. The latitudinal asymmetry of the model climate is thus determined by a balance between the symmetry-breaking land forcing and the symmetry-restoring seasonal solar forcing.

1. Introduction

Solar radiation is the ultimate driving force for oceanic and atmospheric motions. Although the distribution of the annual-mean solar radiation at the top of the atmosphere is predominantly symmetric about the equator, the response of the earth climate to this external forcing is far from symmetric. Strong hemispheric asymmetries are seen in atmospheric polar vortexes, midlatitude westerly jets, and ocean deep water formation, to name but a few. This note concerns the intertropical convergence zone, which is a zonal band of precipitating clouds in the Tropics, and the rising branch of the global Hadley circulation. Over the Atlantic and central and eastern Pacific, the annual-mean precipitation climatology displays pronounced asymmetry with respect to the equator, with the ITCZ centered around 10°N.

Hemispheric asymmetry of continental geometry has long been speculated to be the cause of the Northern Hemispheric (NH) position of the Pacific ITCZ. How land forcing is transmitted to longitudes thousands of kilometers away has not been known, given that the zonal scale of the direct atmospheric response is an order of magnitude smaller than that of the NH ITCZ. Recently Xie (1996, X96 hereafter) proposed a transmitter mechanism that involves an interaction of wind speed, evaporation at the ocean surface, and sea surface temperature. Coupled ocean–atmosphere waves carry the effects of land forcing westward, lowering SST and suppressing convection south of the equator.

In addition to land forcing, the seasonal cycle of insolation is another factor likely to affect the configuration of the ITCZ (Mitchell and Wallace 1992; Giese and Carton 1994). Previous model studies tend to give contradictory results on the role of seasonal solar forcing, however. Pike (1972) argues that the ITCZ should follow the seasonal migration of the sun, whereas Xie and Philander (1994) report that ocean–atmosphere feedback could keep the ITCZ to one side of the equator against seasonal solar forcing. Reflecting the lack of understanding of the processes involved, the simulation of the Pacific NH ITCZ remains a challenge for state-of-the-art coupled general circulation models (CGCMs, Mechoso et al. 1995).

This study attempts to address the following issues associated with the seasonal variation of solar radiation. Does it enhance or reduce equatorial asymmetry of the ITCZ, and what role does the seasonal component of land forcing play? A coupled ocean–atmosphere model is used, which consists of a shallow water atmosphere forced by latent heating, and an ocean mixed-layer forced by radiative and latent heat fluxes (appendix). Continental forcing is a complicated issue that atmospheric GCMs still have difficulty handling (Navarra et al. 1994). Here, land forcing is not explicitly treated, but is represented by its effect on atmospheric circulation; cross-equatorial southerly winds are imposed on the eastern boundary of the model, following X96. Standard equatorial wave theory suggests that much of these southerlies is due to latitudinal asymmetries in land geometry. Under the long-wave approximation, equatorial *asymmetries* in the SST excite only westward propagating Rossby waves (see Fig. 3 of Gill 1980) and thus have no effects on atmospheric flow on the eastern boundary of the ocean. Representing land forcing as an eastern boundary condition is therefore

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Corresponding author address: Dr. Shang-Ping Xie, Graduate School of Environmental Earth Science, Hokkaido University, Sapporo 060, Japan.
E-mail: xie@eoas.hokudai.ac.jp

self-consistent in the present model context. Attention will be restricted to the ocean–atmospheric response to this prescribed boundary forcing.[1] It will be shown that this boundary forcing, together with ocean–atmospheric coupling and seasonal solar forcing, determines the configuration of the model ITCZ.

2. Weak seasonal forcing

For small departures from equatorial symmetry, a linear one-dimensional model of zonal variations can be derived from the full model. Latitudinal asymmetry of the system is represented by a single variable of meridional wind speed at the equator V, or equivalently the hemispheric SST difference Θ. The atmospheric response to a latitudinal asymmetric SST forcing is described by a low-order version of the Matsuno–Gill model (appendix)

$$\left(1 - \frac{\partial}{\partial x}\right)V = \Theta, \qquad (1)$$

which features a damped long Rossby wave trapped to the west of the forcing. The Kelvin wave response is absent here, as only equatorial asymmetric heating is considered. Changes in latitudinal asymmetry of the model SST are caused by hemispheric differences in wind speed/evaporation and insolation:

$$\left(\frac{\partial}{\partial t} + 1\right)\Theta = \sigma V + S. \qquad (2)$$

Here, x is the zonal coordinate nondimensionalized with the e-folding scale of the damped atmospheric Rossby wave (~1000 km), t is the time nondimensionalized with the oceanic Newtonian cooling rate (~1 yr^{-1}), and σ is the coupling coefficient. The full model with $\sigma \sim 1$ produces a zonal profile of V that resembles observations in the eastern Pacific (section 3, Fig. 3). The model is forced by an annual harmonic (angular frequency Ω) of insolation, $S = \tilde{S}e^{i\Omega t}$, while the semiannual insolation cycle is not included, as it is small and equatorially symmetric. The eastern boundary condition is

$$V|_{x=0} = V_E + \tilde{V}_E e^{i\Omega t}, \qquad (3)$$

where V_E and $\tilde{V}_E$ are the annual mean and seasonal parts of asymmetric atmospheric flow on the eastern boundary that are assumed to be maintained by land forcing. Although further work is needed to understand the maintenance of the annual mean southerlies on the South American coast, the seasonal part of the boundary forcing, $\tilde{V}_E e^{i\Omega t}$, is presumably due to seasonal solar radiation over land.

An equation for V is obtained from (1) and (2):

$$\left(\frac{\partial}{\partial t} + 1\right)\left(1 - \frac{\partial}{\partial x}\right)V = \sigma V + \tilde{S}e^{i\Omega t}. \qquad (4)$$

Equation (4) contains westward-propagating free-wave modes for which the ocean–atmosphere feedback is the restoring force (X96). Similar equatorial asymmetric coupled modes seem to exist in some CGCMs (M. Kimoto and X. Shen 1995, personal communication; C. Ma 1995, personal communication). Here, we focus on the forced problem with $\tilde{S} \neq 0$. The solution that satisfies the boundary condition (3) is

$$V = V_E e^{-(\sigma-1)x} + \tilde{V}_S e^{i\Omega t} + (\tilde{V}_E - \tilde{V}_S)e^{x/L}e^{i\Omega[(\sigma/1+\Omega^2)x+t]}, \quad x < 0, \qquad (5)$$

where

$$\tilde{V}_S = \frac{\tilde{S}}{i\Omega - (\sigma - 1)}. \qquad (6)$$

The first term on the rhs of (5) is the steady-state solution, and the second term is the zonally symmetric response to solar forcing. The third term is the response to seasonal boundary forcing, which decays toward the west with an e-folding scale

$$L = \frac{1 + \Omega^2}{\Omega^2 - (\sigma - 1)} \qquad (7)$$

(negative values denote westward amplification). The influence radius of the boundary forcing as measured by L is plotted in Fig. 1, which is a decreasing function of frequency, reaching the maximum at $\Omega = 0$. For $\sigma \approx 1$ that is typical of the eastern Pacific, time-mean land forcing over the Americas can affect a large, if not

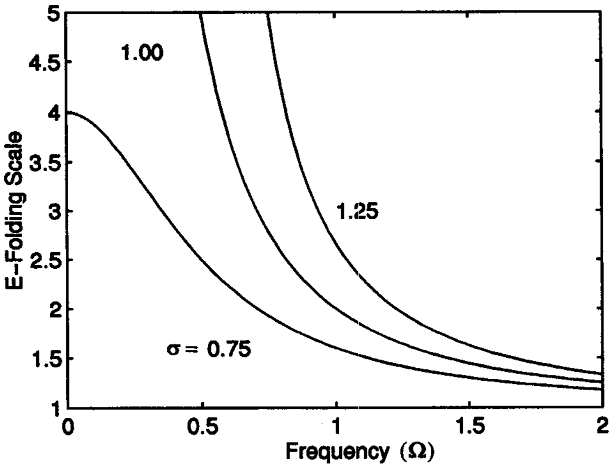

FIG. 1. The e-folding zonal scale of land-forced latitudinal asymmetry as a function of frequency for different coupling coefficients.

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[1] Similar type of studies of the atmospheric response to prescribed SST forcing (e.g., Gill 1980) were instrumental to subsequent leaps in our understanding of ocean–atmosphere interactions involved in El Niño/Southern Oscillation.

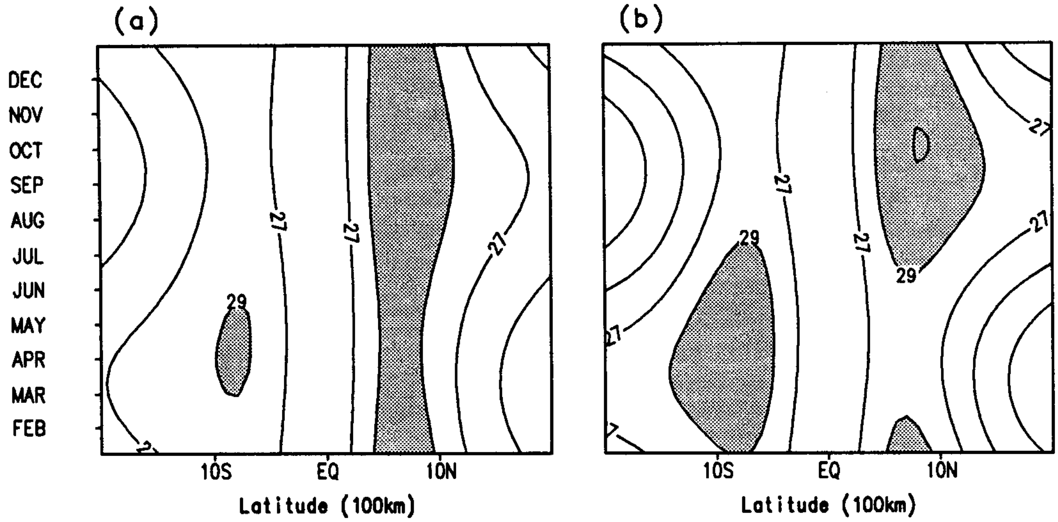

FIG. 2. Time–latitude evolution of sea surface temperature at $x = -5000$ km under seasonal solar radiation, with $\tilde{S}^* =$ (a) 200 and (b) 400 W m^{-2}. Values higher than 29°C are shaded.

infinite, distance to the west. At the annual frequency $\Omega \approx 2$, the e-folding scale $L \approx 1$, or 1000 km, in dimension. It follows that the influence of land forcing is tightly trapped near the coast on the seasonal timescale. The coastal trapping of the coupled response to seasonal boundary forcing is confirmed in the full model (not shown).

In boreal winter, Atlantic air mass often breaks into the Pacific through Central America, cooling the ocean surface and producing the so-called Costa Rica Dome (Umatani and Yamagata 1991). Both the NH ITCZ and the Costa Rica Dome may be viewed as the response of the Pacific ocean–atmosphere system to forcing exerted on its eastern boundary. The SST signature of the Costa Rica Dome is coastally trapped with a zonal scale of 1000 km, in contrast to the basinwide response of the NH ITCZ to time-mean forcing.

The perennial NH ITCZ has a profound consequence on the equatorial seasonal cycle. Superposed on the annual-mean southerlies, the annual variation in meridional wind, which is nearly zonally uniform both in (5) and in observations, forces an annual cycle in equatorial SST (Mitchell and Wallace 1992). Because of the eastward shoaling of the thermocline, the annual cycle in SST has larger amplitudes in the east. The theory presented above is concerned with latitudinally asymmetric modes of a large meridional scale, but it says nothing about the equatorially symmetric modes. Xie (1994b) shows that a different kind of ocean–atmospheric wave mode can cause the phase of annual signals in SST and zonal wind to propagate westward along the equator. This may explain why the westward propagating annual signals in the ocean are largely symmetric about the equator (Minobe and Takeuchi). Liu and Xie (1994) further suggest that this type of ocean–atmospheric wave carry annual signals from the extratropics onto the equator. It is interesting to note that Liu and Xie's wave equation (4.6) bears some similarities to (4).

3. Nonlinear symmetry-restoring effect

In the limit of weak seasonal cycle, the annual-mean state is not affected by the seasonal forcing, as is evident by averaging (5) over a year. This is not true for large seasonal forcing, however. This section describes the response of the full nonlinear model to an annual harmonic of solar radiation,

$$S^* = \tilde{S}^* \sin\theta \sin(t^* - 82), \qquad (8)$$

where θ is the latitude and t^* the time in days, with $\tilde{S}^* = 160$ W m^{-2} being typical over tropical oceans. The seasonal component of land forcing is not included, as its effect is limited to a small coastal region. With $\tilde{S}^* = 0$, the model ITCZ is located in the Northern Hemisphere and convection is suppressed south of the equator.

The effect of the annual solar cycle can be best illustrated when it has large amplitudes. Figure 2b shows the time–latitude section of the SST in the central basin for a large forcing amplitude, $\tilde{S}^* = 400$ W m^{-2}. In boreal winter, the seasonal rise in insolation warms the ocean surface and establishes an ITCZ in the Southern Hemisphere. In the Northern Hemisphere, on the other hand, both the decrease in solar radiation and the high winds induced by the Southern Hemispheric (SH) ITCZ lower the SST, leading to suppressed convection. Thus, strong seasonal solar forcing overpowers the symmetry-breaking boundary forcing, leaving a double ITCZ structure in the annual-mean climatology. Figure

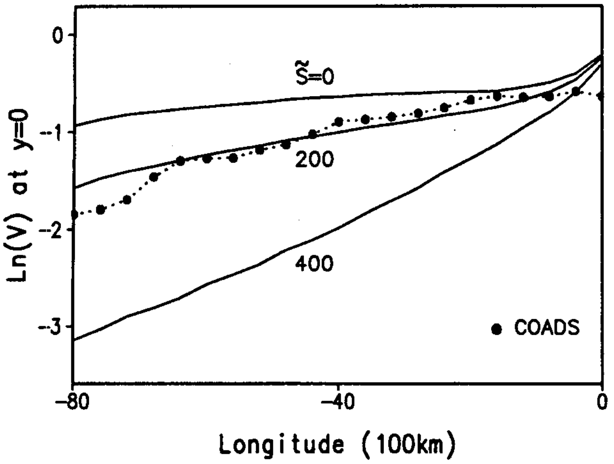

FIG. 3. Logarithm of annual-mean meridional wind speed at the equator as a function of longitude for different amplitudes of annual cycle in solar radiation. COADS data are rescaled and plotted in solid dots, with the west coast of South America at $x = 0$.

3 compares the cases of large and zero seasonal forcing in terms of annual-mean meridional wind speed at the equator, which is taken as a measure of equatorial asymmetry. Seasonal forcing of large amplitudes greatly reduces equatorial asymmetry, causing it to decay rapidly off the eastern boundary. In the central ocean, the cross-equatorial wind is an order of magnitude weaker than that in the steady forcing case. Key to the reduction of equatorial asymmetry is the nonlinear SST threshold for convection, which limits SST forcing to a warm ocean surface. With an intermediate forcing amplitude $\tilde{S}^* = 200$ W m^{-2}, the model features a perennial NH ITCZ and a seasonal SH ITCZ in boreal spring (Fig. 2a). Such a boreal spring double ITCZ is observed in the eastern Pacific. Also plotted in Fig. 3 are observed meridional wind speeds based on the Comprehensive Ocean–Atmosphere Data Set (COADS) (Slutz et al. 1985), which decrease westward and scatter around the $\tilde{S}^* = 200$ curve.

4. Discussion

The position of the Pacific ITCZ north of the equator is a result of land forcing over the Americas. Not all land forcing can be felt over a large zonal distance, however. The influence radius of a symmetry-breaking force that acts on the eastern boundary of the ocean is a decreasing function of forcing frequency. The ocean–atmospheric response to a seasonal boundary forcing is strongly trapped near the coast. This theoretical finding is consistent with the observed contrast in zonal scale between the Pacific NH ITCZ and the Costa Rica Dome forced by the seasonal Atlantic air outbreaks.

Annual-mean land forcing can penetrate deep into the west in the model. Seasonal solar forcing at the ocean surface acts to reduce the influence radius of steady land forcing, causing a basinwide reduction in the latitudinal asymmetry of the annual-mean climate (Fig. 3). A tendency for latitudinal asymmetry of the ITCZ to decrease with increasing effects of annual forcing can be seen in Giese and Carton's (1994) CGCM experiments. Figure 4 quantifies this tendency in this model, showing the meridional wind speed at the equator as a function of annual solar harmonic. The open circle at $S^* = 200$ W m^{-2} features a seasonal double ITCZ that resembles observations in the eastern Pacific (Fig. 2a). Under the same seasonal forcing at $S^* = 200$ W m^{-2}, a reduction in the mean boundary forcing (solid circle) moves the model into the symmetric regime characterized by a double ITCZ in the annual-mean climatology. It follows that the key region for the basin-scale NH ITCZ is the eastern Pacific, a region whose climatological features the current GCMs have difficulty simulating. Along the west coast of South America, OGCMs tend to underestimate the intensity of the cold tongue, while atmospheric general circulation models fail to produce the strong southeasterly trades, a defect often attributed to the poor representation of the Andes. These seemingly local deficiencies of the uncoupled GCMs are likely to have a great impact over the whole Pacific and could be a cause of the poor simulation of the Pacific NH ITCZ by the current coupled GCMs.

It is not well understood what maintains the strong cross-equatorial southerlies on the west coast of South America. While the effect of the South American monsoon is blocked by the Andes, the eastern Pacific is under the influence of the Northern Hemispheric monsoon. This contributes to the annual-mean southerly winds on the South American coast. Without similar topographic features on Africa, the prevailing southerlies in the eastern Atlantic seem to be due to the large

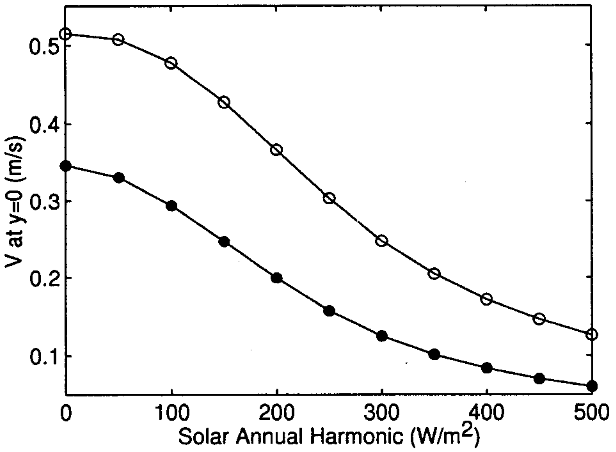

FIG. 4. Annual-mean meridional wind speed at the equator averaged over the eastern half of the ocean as a function of annual harmonic of solar radiation. The lower curve is obtained by halving the time-mean boundary forcing.

land mass to the north of the equator. The southeast-to-northwest sloping of the eastern boundaries of the Pacific and the Atlantic, which favors coastal upwelling south of the equator in an environment of easterly winds, has also been proposed as being crucial for initiating hemispheric asymmetry. The individual and collective effects of these continental asymmetries need to be investigated with an atmospheric model capable of handling land processes.

Despite all the difficulty in pinning down the responsible continental features, the long-held notion that land forcing causes the Pacific ITCZ to form in the Northern Hemisphere is confirmed by this study. It further suggests that the zonal distribution of latitudinal asymmetry of the tropical climate is determined by ocean–atmosphere coupling and is affected by seasonal solar forcing. The role of ocean–atmosphere interaction in establishing the basin-wide NH ITCZ has just begun to be investigated. In addition to the wind-evaporation–SST feedback illustrated here, both coastal upwelling and low-level stratiform clouds contribute significantly to the low SST off the west coast of South America. The temperature inversion at the top of the surface boundary layer is crucial for the formation of stratus clouds, with its strength controlled by both the downdraft aloft and the cold SST below (Klein and Hartmann 1993). Thus, over a cold ocean surface, a positive feedback is possible between SST and low-level stratus clouds (Philander et al. 1996, manuscript submitted to *J. Climate*). These interaction mechanisms are complementary to each other, both contributing to maintaining the Pacific NH ITCZ.

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APPENDIX

Model Equations

The atmosphere is the Matsuno–Gill-type model

$$\epsilon U + Y\mathbf{k} \times \mathbf{U} = -\nabla\Phi, \qquad \text{(A1)}$$

$$\epsilon\Phi + \nabla \cdot \mathbf{U} = -Q, \qquad \text{(A2)}$$

where (U, V) are the wind velocity vector, Φ the geopotential, Y the nondimensional meridional coordinate, and ϵ the nondimensional damping rate. Heating by precipitating convection is a nonlinear function of SST with a threshold at $T_c = 27.5°\text{C}$,

$$Q = K_Q(T - T_c)H(T - T_c), \qquad \text{(A3)}$$

where $H(x)$ is the Heaviside function and K_Q the thermal coupling coefficient.

The ocean extends from 30°S to 30°N with a longitudinal extent of 130°. The mixed-layer temperature T is governed by

$$\frac{\partial T}{\partial t} = \frac{Q_0 - Q_w - C_E^* W q(T)}{\rho c_p h} + \kappa \nabla^2 T, \qquad \text{(A4)}$$

where h is the mixed-layer depth, ρ and c_p are the water density and specific heat, κ is the diffusivity, Q_0 the radiative flux, C_E^* the drag coefficient, and q the saturated specific humidity given by the Clasius–Clapeyron equation. The wind speed is

$$W = \max(4, |\mathbf{U} + \mathbf{U}_0|), \qquad \text{(A5)}$$

where a constant easterly wind, $U_0 = (-4 \text{ m s}^{-1}, 0)$, is imposed to mimic the effect of the equatorward eddy momentum transport, and a minimum wind speed of 4 m s^{-1} accounts for the effect of high-frequency atmospheric disturbances on evaporation. The cooling by the equatorial upwelling is parameterized as an equatorially trapped, symmetric function of latitude Q_w. As a result, surface winds can affect the ocean only through evaporation. For simplicity, all the external fields are set to be zonally uniform and typical of the eastern Pacific. Since the dominant coupled mode has a westward phase propagation [see (4)], zonal variations in parameters have little effect on the solutions upstream in the eastern basin. The coupling coefficient K_Q is so chosen so that the meridional wind speed at the equator is nearly zonally uniform in response to a small boundary forcing, corresponding to $\sigma = 1$ in (2).

Large meridional variation in the mixed-layer depth is observed in the eastern Pacific, with a thermocline ridge below the ITCZ and much deeper thermocline on the opposite side of the equator. The change in h does not affect the steady-state solution of (A4) if the diffusivity is small. The insensitivity of hemispheric SST differences to thermocline depth is demonstrated by Xie (1994a), using OGCM. For simplicity, a spatially uniform value of $h = 50$ m is used in this study. The meridional distribution of h will have a quantitative effect on the mixed-layer response to seasonally varying solar radiation, however.

Under the long-wave approximation, an equation for V can be derived from (A1) and (A2):

$$\frac{\partial^2 V}{\partial Y^2} + \left(\frac{1}{\epsilon}\frac{\partial}{\partial X} - Y^2\right)V = -\frac{\partial Q}{\partial Y} + \frac{Y}{\epsilon}\frac{\partial Q}{\partial X}. \qquad \text{(A6)}$$

For SST disturbances antisymmetric about the equator and confined to $|Y| < Y_P$, valid boundary conditions are $V|_{Y=\pm Y_P} \approx 0$, where $Y = \pm Y_P/2$ are the latitudes of the off-equatorial SST maxima. Applying a center difference to (A6) at the equator, with the SST's grid points at $Y = \pm Y_P/2$, leads to (1).

See Xie and Philander (1994) and X96 for the derivation of the model equations.

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