Global climate change and ENSO: a theoretical framework

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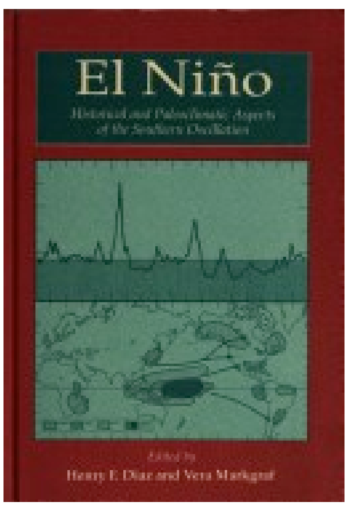


Abstract

To better understand what drives El Niño, an analytical model of the coupled ocean-atmosphere system over the equatorial Pacific is constructed. The equatorial atmosphere is approximated as a linear feedback system whose surface winds are driven by sea surface temperature (SST) gradients and whose thermal effect is to restore the entire equatorial SST to its maximum value-the SST of the warm-pool. The upper ocean is represented by a shallow water model capped by a mixed layer with a constant depth. The zonal mean stratification of the thermocline is maintained by upwelling from the deep ocean. The model captures the oscillatory behavior of the present tropical Pacific climate—the El Niño–Southern Oscillation. The main features of the oscillation in the model agree well with the observed El Niño including the period of the oscillation and the phase relationship between the variations of SST and the variations in the depth of the thermocline. Moreover, the model predicts that the climate of the eastern tropical Pacific has two regimes: one is warm and steady, and the other is cold and oscillating, consistent with the inference from geoarcheological data that El Niño did not exist during the early-to-mid-Holocene when the global and regional climate was warmer than today. The transition from the steady climate to the oscillating climate takes place when the temperature contrast between the surface warm-pool and the deep ocean exceeds a critical value. A stability analysis reveals that the zonal SST contrast and the accompanying wind-driven currents have to be sufficiently strong to become oscillatory and that requires a sufficiently large difference between the temperature of the warm-pool and the temperature of the deep ocean. On the time-scale of millennia, a sufficiently cold equatorial deep ocean implies a sufficiently cold high latitudes, a condition which is met by the present climate, but possibly not by the climate of the early-to-mid Holocene. In the oscillating regime, the magnitude of El Niño is found to increase monotonically with increases in the difference between the temperature of the warm pool and the temperature of the deep ocean. The increase in the magnitude of El Niño is accompanied with an increase in the zonal SST contrast in the equatorial region. The implication of these results for the response of El Niño to an increase in the greenhouse effect is discussed.

Introduction

Global ramifications of El Niño-Southern Oscillation (ENSO) has been well recognized (Glantz et al 1991; Rasmussen and Wallace 1983). What's much less studied is the dependence of this oscillatory behavior of the tropical Pacific climate on the basic parameters that characterize the global climate. What will happen to ENSO after significant global warming takes place? Was ENSO during the last ice age more energetic or weakened? These are outstanding questions in the study of climate change and yet we have just began to study them. Pivotal to addressing these questions is to have a clear understanding of what drives ENSO. The purpose of this article is to put forward a simple hypothesis on what drives ENSO.

The existence of ENSO has been attributed to the dynamic coupling between the atmosphere and ocean in the equatorial Pacific region (Neelin et al 1997, Jin 1996, Neelin 1991, Suarez and Schopf 1988, Battisti 1989, Zebiak and Cane 1987, Bjerknes 1969). The dynamic coupling refers to the positive feedback loop between surface wind stress, SST gradients, and oceanic upwelling. Over the tropical Pacific Ocean, the surface winds are driven by SST gradients (Lindzen and Nigam 1987). As a consequence, changes in SST gradients affect the strength of surface winds. Because upwelling is driven by the surface winds, changes in the strength of the surface winds affects the strength of the upwelling, which in return affects the SST gradients. The prevailing theory of El Niño, the delayed oscillator hypothesis, states that the positive feedback loop between surface wind stress, SST gradients, and oceanic upwelling is responsible for the growth of a SST anomaly, while the phase lag between upper ocean heat content and surface winds provides the mechanism for the phase transition. The theory also links the period of the oscillation to the memory of the subsurface ocean.

The dynamic coupling between the atmosphere and ocean in the equatorial Pacific, however, may not be sufficient for the existence of El Niño. Recent geoarcheological and geomorphic evidence suggests that El Niño may not have been present prior to about 5000 years ago, when the global and regional climate were warmer than today (Sandweiss et al 1996, 1997; Keefer et al 1998). The picture that emerges from this suggestion is that the eastern tropicalPacific may have two distinct regimes: one is warm and steady, the other is colder and oscillating. Moreover, the former regime is accompanied by a warmer global climate and the latter is associated with a colder global climate. The early-to-mid-Holocene was only slightly warmer than today, and as usual the warming was most noticeable at higher latitudes (the polar region was about 2-3 °C warmer than today (Klimanov 1982; Velichko 1984; Bozenkova and Zubakov 1984). The suggested connection to the higher latitude temperature is particularly interesting because it suggests a potential role of the temperature of the equatorial deep ocean in determining the state of the surface ocean. The property of deep water of the Pacific ocean is connected to the surface condition in the high latitude oceans through the conveyor-belt-like thermohaline circulation (Broecker 1987). Paleo-temperature records indicate that the deep ocean temperature of the tropical oceans responds sensitively to coolings in the high latitudes (Hoffert 1990; Berger 1981).

The question is then whether a significantly warmer equatorial deep ocean temperature can result in a warmer eastern Pacific that has no El Niño. The lack of a ready answer from existing paradigms of El Niño to this very specific question highlights a deficiency in the reigning paradigms, which is a clear exposition of where El Niño derives its energy, or what drives El Niño (Neelin et al 1998). Though the existing paradigms of El Niño suggest that El Niño depends on the climatological state, which aspect of the climatological state serves as the main driving force of El Niño is not clear. We will attempt to isolate the main thermal forcing that drives El Niño and examine the behavior of the tropical SST as a function of the strength of the thermal forcing. We will suggest that El Niño is a thermally forced oscillation: it is fundamentally driven by the contrast between the tropical maximum SST and the temperature of the equatorial deep ocean.

A heuristic model

We begin with some heuristic arguments on what determines the magnitude of El Niño. It has long been recognized that El Niño is associated with a zonal redistribution of warm water in the equatorial Pacific (Wyrtki 1985). To help fix this picture in mind, we may divide the surface equatorial pacific into two boxes: the western Pacific box with temperature T_w and the eastern Pacific with temperature T_E (Fig. 1a). T_w and T_E may be regarded respectively as the characteristic climatological value of the warm-pool and cold-tongue. To the extent that the El Niño warming is due to an eastward displacement of the warm-pool (Fig. 1b), we may equate the magnitude of El Niño, $\tilde{T}$, to the difference between T_w and T_E.

$$\tilde{T} \sim T_w - T_E \qquad (1)$$

The normal or climatological condition of the equatorial Pacific may be safely regarded as the average between the El Niño and the La Niña condition. It follows that the climatological SST in the eastern Pacific may be equated to the average of the SST in the eastern Pacific during El Niño ($\mathrm{T_E(Ni\tilde{n}o)}$) and La Niña ($\mathrm{T_E(Ni\tilde{n}a)}$. During La Niña, the thermocline is almost exposed to the surface in the eastern Pacific. For our purpose here, we will make an approximation and assume that the thermocline does expose to the surface during La Niña (Fig. 1c). This gives $\mathrm{T_E(Ni\tilde{n}a)} \sim T_{sub}$, where T_{sub} is the subsurface temperature. Further noting $\mathrm{T_E(Ni\tilde{n}o)} \sim \mathrm{T_w}$, we have

$$T_E \sim \frac{1}{2}(T_w + T_{sub}) \qquad (2)$$

Combining Eq. (1) with Eq. (2), we have

$$\tilde{T} \sim \frac{1}{2}(T_w - T_{sub}) \qquad (3)$$

The value of T_{sub} depends on the east-west slope of the thermocline as well as the vertical stratification of the thermocline (Fig. 1 c). The slope of the thermocline is balanced gravitationally by the zonal wind stress and thus is largely determined by the zonal SST contrast, or T_w–T_{sub}. Now we see that the value of T_{sub} depends on a factor that in turn depends on the value of T_{sub}—a nonlinearity that is responsible for the self-organization that we will introduce later. To the leading order, the vertical stratification is given by the difference between T_w and the deep ocean temperature T_c. This raises the role of temperature of the deep ocean T_c in determining the value of T_{sub} and thereby the magnitude of El Niño. The next section is to describe these heuristic arguments using differential equations so that the analysis can be carried out quantitatively. A quantitative analysis will allow us to see that the difference between T_w and T_c needs to be sufficiently large to enable the coupled equatorial ocean- atmosphere to find itself in an oscillatory state, a point that can be brought out only when the nonlinearity and the effect of thermal and mechanical damping are considered quantitatively.

A more rigorous model

Focusing on the two fundamental features of the tropical Pacific climate, the zonal SST contrast and ENSO, we may reduce the coupled ocean- atmosphere to a low-order system through spatial truncation (Sun 1997a, b; Sun and Liu 1996; Neelin 1991). The resulting model is schematically illustrated in Fig. 2a. The upper ocean model is a box model version of the model of Cane (1979), a shallow water model for the equatorial upper ocean embedding a mixed layer with a fixed depth. The zonal mean vertical stratification is maintained by a balance between surface heating from above and upwelling of cold water from the equatorial deep ocean. The equatorial deep ocean is connected to the high latitude oceans through the global thermohaline circulation (Fig. 2b). Following Sun (1997 a,b), the surface ocean is divided into two regions with equal areas: the western surface ocean (120°E–155°W) and the eastern surface ocean (155°W-70°W). They are represented respectively by two boxes with temperature T_1 and T_2. The atmosphere is approximated by a linear feedback system whose surface winds are driven by SST gradients and whose thermal effect is to force the entire equatorial SST towards its maximum value—the SST of the warm-pool (Sun and Liu 1996; Neelin 1991).

The heat budget of the surface ocean may be formulated as follows:

$$\frac{dT_1}{dt} = \frac{Q_1 A_1}{C_p \rho V_1} + sq(T_2 - T_1) \tag{4}$$

$$\frac{dT_2}{dt} = \frac{Q_2 A_2}{C_p \rho V_1} + q(T_{sub} - T_2) \tag{5}$$

where Q_1A_1 and Q_2A_2 are respectively the net heat flux into the western surface ocean and the eastern surface ocean with A_1 and A_2 being respectively the surface area of the two regions. V_1 and V_2 are the volume of the two boxes for the surface ocean. C_p is the specific heat and ρ is the density of water. We assume that $V_1 = V_2$ and $A_1 = A_2$. The second term on the right side of Eqs. (4) and (5) represents the heat transport by the ocean currents; $q = w/H_1$, where w is the upwelling velocity, and H_1 is the depth of the surface ocean; $sq = u/L_x$, with u and L_x being respectively the zonal velocity and the half zonal width of the tropical Pacific ocean. Thus $s = uH_1/wL_x$, the ratio between the east-west flow and the total upwelling. It assumes a constant value here. There is observational evidence suggesting that zonal advection may play an important role in establishing the time scale of El Niño (Picaut et al 1996). Note that the inflow of box 1 comes from box 2 and thus carries temperature T_2. Correspondingly, the heat moved from box 2 to box 1 is $u(V_1/L_x)\rho C_p T_2$ where $u(V_1/L_x)$ is the flow rate. At the same time, through the outflow of the same rate, the heat moved out of box 1 is $u(V_1/L_x)\rho C_p T_1$. Consequently, the net heat gained by box 1 from the ocean currents is $u(V_1/L_x)\rho C_p(T_2\text{-}T_1)$. Dividing this term by $V_1 \rho C_p$, one obtains the second term on the right hand of Eq. (4). The same consideration was employed to obtain the corresponding terms in Eq. (5). Such a formulation of the oceanic transport may also result from an upstream differencing of the advection terms in the temperature equation of the surface ocean. T_{sub} in Eq. (5) is the temperature of the water upwelled into the mixed layer.

The surface heat flux into the ocean is assumed to be proportional to the difference between the local SST and the SST of the warm-pool.

$$Q_1 = C_p \rho H_1 c (T_w - T_1) \tag{6}$$

$$Q_2 = C_p \rho H_1 c (T_w - T_2) \tag{7}$$

where T_w is the SST of the warm-pool, H_1 is the depth of the mixed layer, C_p is the specific heat, and c is a constant whose reciprocal has the unit of time ($1/c$ is the characteristic time for the surface flux to remove a SST anomaly). Such a formulation is consistent with the observation that the net heat flux over the warm-pool is nearly zero (Ramanathan et al 1995). The formulation produces zonal distribution of heat flux that is consistent with that of Esbensen and Kushnir (1981). The formulation also implies the surface flux is negatively correlated with the SST variations in the equatorial region, which is also consistent with the observed relationship between the interannual variations of surface flux and the SST (Sun and Trenberth 1998).

As demonstrated by previous studies, the positive feedback loop between surface winds, ocean currents, and zonal SST gradients is crucial for the occurrence of El Niño. We capture the essence of this feedback loop by assuming that the strength of the upwelling is proportional to the strength of the surface wind and that the strength of the surface wind is proportional to the SST gradients. This results in the following equation for q,

$$q = \frac{\alpha}{a}(T_1 - T_2) \tag{8}$$

where α measures the sensitivity of wind-stress to changes in the SST gradients, and a defines the adjustment time scale of the ocean currents to changes in the surface winds.

Using $\phi(z)$ to represent the zonal mean temperature profile of the subsurface upper ocean and assuming that the effect of the zonal variation of the depth of the upper ocean is simply to displace the profile $\phi(z)$ vertically (Zebiak and Cane 1987), we have

$$T_{sub} = \Phi(-H_1 + h_2') \tag{9}$$

where h_2' *is* the deviation of the depth of the upper eastern Pacific ocean from its reference value H. H may be regarded as the zonal mean depth of the upper ocean. $\Phi(z)$ may be parameterized as follows,

$$\Phi(z) = T_w - \frac{T_w - T_c}{2}\left(1 - \tanh\left(\frac{z + z_0}{H^*}\right)\right) \tag{10}$$

where T_c is temperature below the upper ocean. We will loosely call T_c the temperature of the equatorial deep ocean. H^* and z_0 are constants that have the unit of depth. The basic physics embodied in Eq. (10) is that the vertical temperature gradients in the equatorial ocean result from atmospheric heating from above and upwelling of cold water from below. The exponential form is based on observations (Zebiak and Cane 1987) and is also consistent with results from theoretical and numerical models of the thermocline (Munk 1966; Verdiere 1988). The variation of the depth of the upper ocean (or the thermocline) are governed by the following equations,

$$h_2' - h_1' = -\frac{H_1}{H_2} H \frac{\alpha}{b^2}(T_1 - T_2) \tag{11}$$

$$\frac{1}{r}\frac{dh_1'}{dt} = -h_1' + \frac{H_1}{2H_2} H \frac{\alpha}{b^2}(T_1 - T_2) \tag{12}$$

Eq. (11) represents the balance between the zonal pressure gradients and the zonal wind stress. h_1' is the deviation of the depth of the upper ocean in the western Pacific from its reference value H. $H_2 = H - H_1$. $b=c_k/L_x$ where c_k is the speed of the first baroclinic Kevin wave and L_x is the half width of the basin. Eq. *(12)* is an approximate way to represents the slow adjustment of the depth of the thermocline to its equilibrium value determined by the surface wind-stress and mass conservation; r defines the time scale of this slow adjustment (Jin 1996).

With Eqs. (6) and (7), we may rewrite Eqs. (4) and (5) as

$$\frac{dT_1}{dt} = c(T_w - T_1) + sq(T_2 - T_1) \tag{13}$$

$$\frac{dT_2}{dt} = c(T_w - T_2) + q(T_{sub} - T_2) \tag{14}$$

Equilibrium SST as a function of T_c

Fig. 3 shows the equilibrium SST of the coupled system as a function of T_c. The value of T_w is fixed at 29.5°C. The use of a constant T_w is based on paleotemperature records indicating that the value of T_w was strongly regulated (Crowley and North 1991). Starting from a very warm value for T_c, the equatorial Pacific is characterized by a very warm SST without zonal SST contrast. Zonal SST gradients are developed when T_c is below 21.5°C. The SST gradients increase quickly with further decreases in T_c. When T_c is below approximately 17.5°C, the coupled system starts to oscillate. Oscillations at T_c = 17.3°C are plotted in Fig. 4. The oscillations have a period of about 4 years with a slight westward phase propagation (Fig. 4a). The variations of the depth of the thermocline in the eastern half of the ocean also lead slightly the variations of SST in that region (Fig. 4b). All these features agree well with those of observed El Niño (Wang and Fang 1996, Rasmusson and Carpenter 1982). In the oscillating regime, the magnitude of the oscillation increases with further decreases in T_c (Fig. 5). The typical amplitude of ENSO anomaly in the present climate is about 2 °C. Fig. 5 raises the possibility that the present climate may be close to the critical point. The early-to-mid Holocene was about 2-3°C warmer in the polar latitudes (Klimanov 1982, Velichko 1984, Bozenkova and Zubakov 1984, R. Webb, personal communication). Since the temperature of the equatorial deep ocean has to equilibrate with the temperature of the polar oceans through the global circulation, it is plausible that in the early mid-

Holocene, the value of T_c may have been significantly warmer than the critical value than that is needed to sustain an oscillatory surface climate.

The Physics of El Niño

To understand the physics of the two bifurcations in Fig. 3, we replace Eq. (10) by a linear profile, so that Eq. (9) can be written as,

$$T_{\text{sub}} = T_{s0} + \gamma h_2' \qquad (15)$$

where $T_{s0} = \lambda T_w + (1-\lambda)T_c$ and $\gamma = \gamma^*(1-\lambda)\,(T_w - T_c)/H_2$. λ and γ^* are numerical constants that are related to z_0 and H^* in Eq. (10). Eq. (15) may be regarded as a first order approximation of Eq. (10) with T_{s0} and γ being respectively the characteristic temperature and the lapse rate of the upper subsurface ocean. With Eq. (15) for T_{sub}, the coupled system has qualitatively the same behavior as the original system (Fig. 6).

A schematic illustration of the re-organization of the system as the deep ocean temperature becomes systematically colder is given in Fig. 7. In state 1, the heat transfer is a balance between downward background diffusion and upwelling of the background circulation. In this stage, the deep ocean has minimum control on the surface temperature and therefore minimum control on the surface climate. Because of the positive feedback loop between SST, surface winds, and oceanic upwelling, this state is potentially unstable. When the deep ocean temperature becomes sufficiently cold and the vertical stratification becomes sufficiently large, the system possesses more potential energy and the positive feedback becomes strong enough to overcome the effect of thermal and mechanical damping. This enables the system to create zonal SST contrast, winds, and currents. The system enters a state with steady zonal SST contrast and steady currents. In this state, the deep ocean has more control on the surface climate, and the warm-pool retreats to the west. The zonal SST contrast and the strength of the currents increases with further decreases in the deep ocean temperature. When the strength of the currents becomes so strong that “overshoot” occurs, the system enters an oscillatory state. The situation is analogous to a water-wheel (Strogatz 1994).

A more quantitative analysis may be obtained by nondimensionalizing Eqs. (8), (11), (12), (13), (14), and (15) (appendix 1). After non-dimensionalization, we find that the dynamic behavior of the system is determined by four non-dimensional parameters: $R = \alpha(T_w - T_{s0})/ac$, $\Lambda = p\kappa\gamma^*$ (where $p = H_1/2H_2(1+H_1/H_2)$ and $\kappa = ac/b^2$), $\sigma = r/c$, and s is defined above.

The non-dimensional parameter R is analogous to the Rayleigh number in the classical problem of Rayleigh-Bénard convection (Rayleigh 1916). It measures how hard the system is driven, relative to the dissipation. In the present case, the driving force for the circulation is from the dynamic tension between the warmth that the greenhouse effect and the solar radiation tend to create (T_w) and the coolness that the subsurface ocean tries to retain (T_{s0}). The dissipation is from the mechanical and thermal damping. The thermal damping depends on the radiative feedbacks (Sun and Liu 1996, Sun 1997a). It is readily shown that SST gradients start to develop when R=1 (the first bifurcation in Fig. 3). This is easy to understand because the driving force has to be strong enough to overcome dissipation in order to create and maintain a steady circulation.

The oscillation regime is switched on when $R = R_c \cong C/\Lambda$, where C is a parameter whose value depends only on the values of s and σ (Fig. 8). s measures how strong the eastern and the western Pacific is coupled through the surface current. σ measures the memory of the subsurface ocean. For $s = 1/3$, $\sigma = 1/2$ (the values used in Fig. 4) C is about 1. C increases with increases in either s or δ. As long as $s \neq 0$, C remains a finite value even when the ocean has no memory (i.e., $r = \infty$ *or* $\delta = \infty$). For example, with $s = 1/3$, C=1.5 for $\sigma = \infty$. It is easy to show that $\partial T_{sub}/\partial T_2 = \Lambda R$. Therefore, when the circulation in equilibrium is perturbed by a cooling in T_2^* in the amount of $-\delta T_2^*$, the corresponding change in T_{sub}^* will be $-\Lambda R \delta T_2^*$. A decrease in T_{sub} enhances the cooling effect of the upwelling upon T_2. Opposing the cooling effect is the accompanying increase in the surface heat flux. When ΛR is sufficiently large, the enhanced cooling from the upwelling is able to overcome the opposing effect from the accompanying increase in the surface heat flux and consequently the initial cooling in T_2 will be further amplified. The instability is an oscillatory one because after some time, the decrease in T_{sub} is slowed down and eventually stopped by the adjustment of T_1 to changes in the cooling from the zonal advection and by the adjustment of h_l' to changes in the zonal SST contrast. The corresponding saturation and decrease in the cooling from the upwelling allows the surface heating to catch up, stop, and eventually reverse the cooling in T_2. In short, critical for the instability of the steady circulation is that the temperature of the upwelled water depends strongly on the strength of the flow rate (i.e., a large Λ), and that the flow rate (or the thermal forcing that drives the flow) is sufficiently large relative to the thermal and mechanical damping (i.e., a large R). A similar mechanism is responsible for the onset of oscillation in the Lorenz system (Lorenz 1963) which is a low-order approximation of Rayleigh-Benard convection. Thus, fundamentally, El Niño arises from an intensified competition between the warming effects from the atmosphere and the cooling effects from the subsurface ocean. The regime transitions discussed above are further illustrated schematically in Fig. 7.

Note also that $\Lambda \sim b^{-2} = (L_x/C_k)^2$. Therefore the critical value of R or T_w for the onset of oscillation (C/Λ) is inversely proportional to the width of the basin ($T_w - T_c \sim 1/\alpha(C_k/L_x)^2$). This explains the absence of self-sustaining El Niño-like phenomena in the tropical Atlantic ocean whose width is only one third that of the tropical Pacific ocean.

At the critical point $R = R_c = C/\Lambda$ where the oscillation takes place, the period of the oscillation decreases with increases in σ (Fig. 9b). This is consistent with early analysis of El Niño (Neelin 1991, Jin and Neelin 1993a, Neelin and Jin 1993b). The period also depends significantly on the parameter *s (Fig. 9a)*. For $\Lambda = 0.8$, $\sigma = 1/2$, for the period corresponding to $s = 1/3$ is about 11⁄c while the period corresponding to $s = 1$ is reduced to 6⁄c. This provides a theoretical basis for a suggestion by a previous study (Picaut et., al 1996) that zonal advection may play an important role in determining the timescale of ENSO. For a fixed s or σ, the period also increases quickly with increases in Λ (Fig. 9). Doubling Λ can result in a doubling of the period. This reveals the importance of the width of the basin and the speed of the Kelvin wave in setting the time scale of the oscillation (recall $\Lambda \sim (L_x/C_k)^2$). With basin width, s, and σ fixed, changes in T_c or R results in no significant changes in the period of oscillation.

The role of the warm pool SST

The non-dimensional analysis also reveals that El Niño is thermally driven by the difference between Tw and Tc, implying increasing the value of Tw has the same effect on the magnitude of El Niño as decreasing the value of Tc. Fig. 10 and Fig. 11 show respectively the dependence of the equilibrium SST and the magnitude of El Niño on the value of Tw with the value of Tc being fixed. Comparing Fig. 10 with Fig. 3, and Fig. 11 with Fig. 5 confirms the results from the non-dimensional analysis. Again, the equilibrium zonal SST contrast has to be sufficiently large to bring the system to an osccilatory state. With a fixed Tc, this requires that the warm pool SST has to be sufficiently high. The requirement for a sufficiently high warm pool SST implies that El Niños may play a fundamental role in regulating the heat budget of the tropical ocean. The result of Sun and Trenberth (1998) supports this inference. They found that during 1986-87 El Niño, there was a large increase in the poleward heat transport in the equatorial region in both the atmosphere and ocean. Sun (1998) further shows that most of the heat removed by El Niño from the equatorial region to the subtropical ocean does not return to the equatorial ocean, but is eventually transported to extratropics.

Discussion

The focus of this article has been to provide a simple theoretical framework for understanding ENSO variability on the millennia and longer time-scales. For conceptual clarity, we adopted a simple parametrization for the zonal mean stratification of the equatorial ocean which emphasizes the role of the global thermohaline circulation and the implied connection between the equatorial subsurface temperature and the SST of high latitude oceans. It has been suggested that changes in the subtropical SST may also affect the subsurface temperature of the equatorial ocean through the wind-driven subtropical cell (McCreary and Lu 1994, Gu and Philander 1996). Because on the time-scale of millennia or longer, the subtropical temperature usually changes in the same direction as the polar temperature (Budyko and Izael 1991), the qualitative results we derived here for the dependence of ENSO on global climate should not be affected by the simplification we adopted. On the decadal time-scale, changes in the value of the subsurface temperature may be dominated by changes in the subtropical SST. Accordingly, when applying the present theory (i.e., ENSO is driven by the difference Between T_w and T_c) to the decadal variability of El Niño, the value of T_c should be linked to the subtropical SST. Independent of the relative importance of the subtropical cell and the global thermohaline circulation in determining the temperature of the subsurface ocean, the present theory predicts that an increase in the magnitude of El Niño should be accompanied with an increase in the SST contrast between the western Pacific warm pool and the SST in the eastern Pacific cold-tongue.

The suggestion that El Niño is driven by the difference between the warm pool SST and the temperature of the subsurface ocean is qualitatively consistent with the result of Zebiak and Cane (1987). Using an intermediate coupled model they found that reducing the mean vertical stratification of the upper ocean thermocline results in a less energetic oscillation. The prediction of a positive correlation between changes in the zonal SST contrast in the equatorial region and the magnitude of El Niño is qualitatively consistent with the finding of Knutson et al (1997). Using a coupled GCM, they found in their CO_2 experiments that a reduction in the magnitude of El Niño is accompanied with a reduction in the zonal SST contrast in the time mean state. Knutson et al (1997) did not show the corresponding changes in the equatorial subsurface temperature. Judging from the spatial distribution of the SST changes in the CO_2 sensitivity experiments, it is likely that the subsurface temperature in their model increased more than the SST in the warm pool. It needs to be emphasized, however, whether the coupled GCM used by Knutson et al (1997) has correctly predicted the response of warm pool SST and the subsurface temperature to the increases in the greenhouse effect is controversial (Sun 1997b, Cane et al 1997, Clement et al 1996, Sun and Liu 1996). Sun and Liu (1996) suggested that when the dynamical coupling between the atmosphere and ocean is sufficiently strong relative to the thermodynamical feedbacks, the zonal SST contrast may actually increase in response to an increase in the heating. Similar result was obtained by

Clement et al (1996). There is evidence suggesting that the dynamical coupling in the low resolution coupled GCMs may be too week (Cane et al 1997). Cane et al (1997) further showed that the observed zonal SST contrast in the equatorial region has a positive trend over the period of this century. The present theory does not resolve this controversy, but highlights the need to better understand what determines the SST in the warm-pool and what determines the temperature of the subsurface ocean.

Through the difference between T_w *and* T_c provides the major thermal forcing, the non-dimensional analysis reveals that the magnitude of the El Niño also depends on other parameters. Specifically, the analysis shows that an increase in the coupling strength between the atmosphere and ocean has the same effect on the magnitude of El Niño as an increase in the difference between T_w and T_c. The other potentially important parameter is γ^*, which measures the lapse rate of the upper-most part of the thermocline scaled by the mean vertical stratification of the entire upper ocean (Recall that $\Lambda \sim \gamma^*$). Though there is no obvious reason to suspect these two parameters will change significantly in response to global warming, recognizing their role may be important in understanding differences in the simulated El Niño by different climate models. Note that in anomaly models of El Niño such as the model Zebiak and Cane (1987), the mean thermocline is fixed. Correspondingly, γ^* decreases as T_w increases. The difference between the present results and those from Clement et. al., (1996) on the response of El Niño to an increase in T_w are likely due to the differences in treating the response of the thermocline structure. In the present model, the value of γ^*, not the value of γ remains constant as the value of T_w increases.

Appendix 1: Nondimensional analysis

Introducing $\tau = ct$, $q^* = q/c$, $T_1^* = T_1\text{-}T_w/T_w\text{-}T_{s0}$, $T_2^* = T_2\text{-}T_w/T_w\text{-}T_{s0}$, $T_s^* = T_{sub}\text{-}T_w/T_w\text{-}T_{s0}$, $h_2'^* = h_2'/H_2$, and $h_1' = h_1'/H_2$, we have

$$\frac{dT_1^*}{d\tau} = -T_1^* + sq^*\left(T_2^* - T_1^*\right) \quad (16)$$

$$\frac{dT_2^*}{d\tau} = -T_2^* + q^*\left(T_s^* - T_2^*\right) \quad (17)$$

$$\frac{dh_1^*}{d\tau} = -\sigma\left[h_1^* - \mathrm{p}\kappa R(T_1^* - T_2^*)\right] \quad (18)$$

$$q^* = R(T_1^* - T_2^*) \quad (19)$$

$$T_s^* = -1 + \Upsilon^* h_2^* \quad (20)$$

$$h_2^* = h_1^* - 2\mathrm{p}\kappa R(T_1^* - T_2^*) \quad (21)$$

where $R = \alpha(T_w\text{-}T_{s0})/ac$, $\kappa = ac/b^2$, $\sigma = r/c$, *and* $\mathrm{p} = H_1/2H_2(1+H_1/H_2)$. Plugging Eqs. (19), (20), and (21) into Eqs. (16), (17), and (18), and replacing $\gamma^* h_1^*$ by η^*, we have

$$\frac{dT_1^*}{d\tau} = -T_1^* - s\mathrm{R}\left(T_2^* - T_1^*\right)^2 \quad (22)$$

$$\frac{dT_2^*}{d\tau} = -T_2^* + R\left(T_1^* - T_2^*\right)\left[-1 + \eta^* - 2\Lambda \mathrm{R}(\mathrm{T}_1^* - \mathrm{T}_2^*) - T_2^*\right] \quad (23)$$

$$\frac{d\eta^*}{d\tau} = -\sigma\left[\eta^* - \Lambda R\left(T_1^* - T_2^*\right)\right] \quad (24)$$

where $\Lambda = \mathrm{p}\kappa\gamma^*$. After such a manipulation, we see that the dynamic behavior of the coupled system is determined by for non-dimensional parameters: R, Λ, σ and s.

Appendix 2: Definition of symbols used in this study

T_w: warm-pool SST (tropical maximum SST)

T_c: deep ocean temperature

T_1: equatorial western Pacific SST (120E-155W)

T_2: equatorial eastern Pacific SST (155W-70W)

T_{sub}: subsurface temperature

H_1: depth of the mixed layer

H: zonal mean depth of the upper ocean

$H_2 = H - H_1$

L_x: half width of the basin.

u: zonal velocity

w: upwelling velocity

q: w/H_1

s: uH_1/wL_x

1/a: adjustment time scale of the surface ocean currents to changes in the surface winds.

1/c: time scale of removing a SST anomaly by surface fluxes

1/r: time scale of the slow adjustment in the ocean

α: sensitivity of the surface wind stress to changes in the SST gradients

h_1': deviation of the depth of the upper ocean in the western Pacific from its reference value H.

h_2' deviation of the depth of the upper ocean in the eastern Pacific from its reference value H.

c_k: phase speed of the first baroclinic Kelvin wave

b: c_k/L_x

p: $H_1/2H_2(1+H_1/H_2)$

k: ac/b^2

σ: r/c

T_{s0}: characteristic value of the subsurface temperature

λ: $(T_{s0}-T_c)/(T_w-T_c)$

γ: lapse rate of the subsurface ocean

γ_0: $(T_w-T_{s0})/H_2$

γ^*: γ/γ_0

R: $\alpha(T_w-T_{s0})/ac$

Λ: $p\kappa\gamma^*$

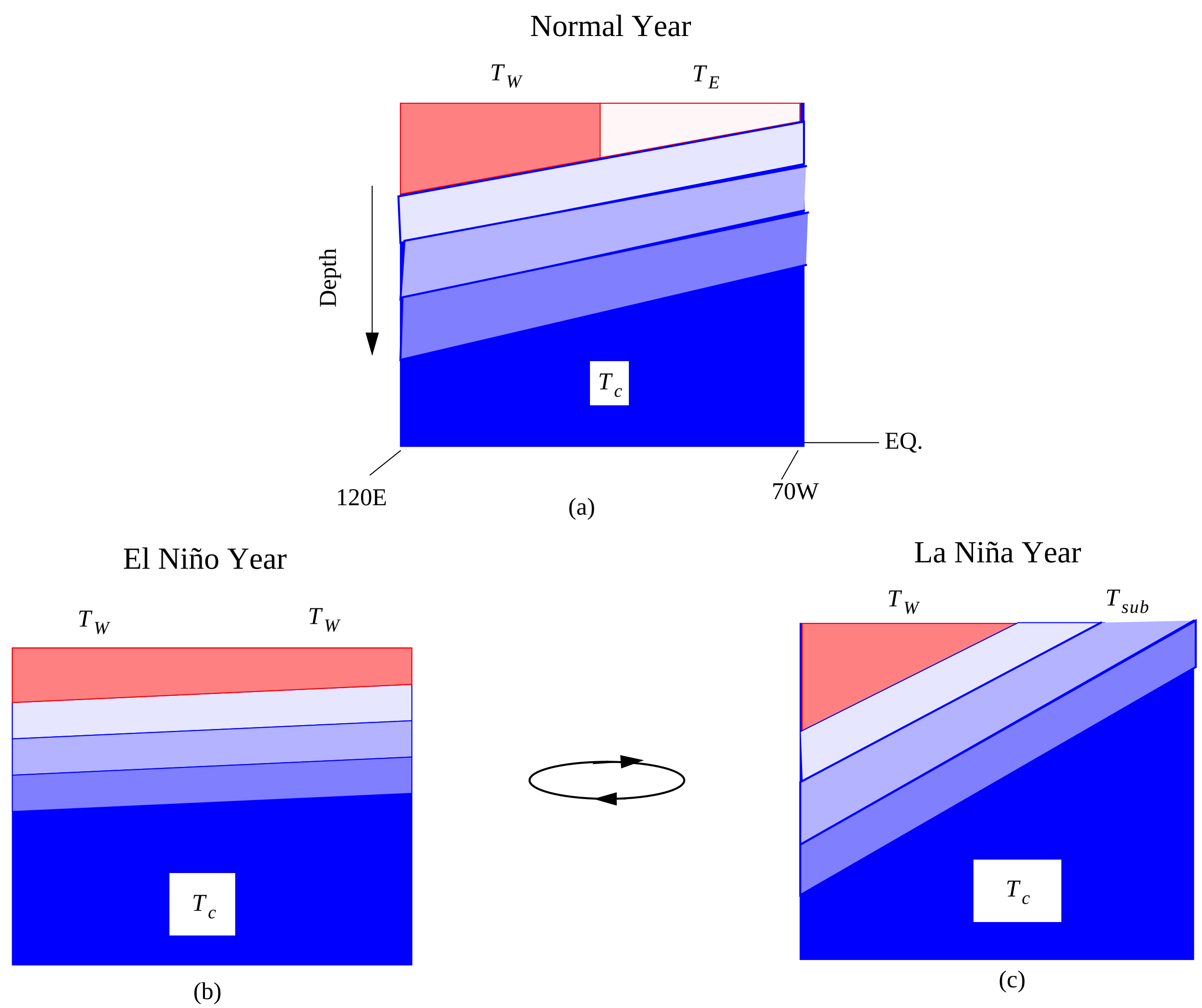


Fig. 1 A schematic illustration of the thermal structure of the equatorial Pacific during a normal year (a), an El Niño year (b), and a La Niña year.

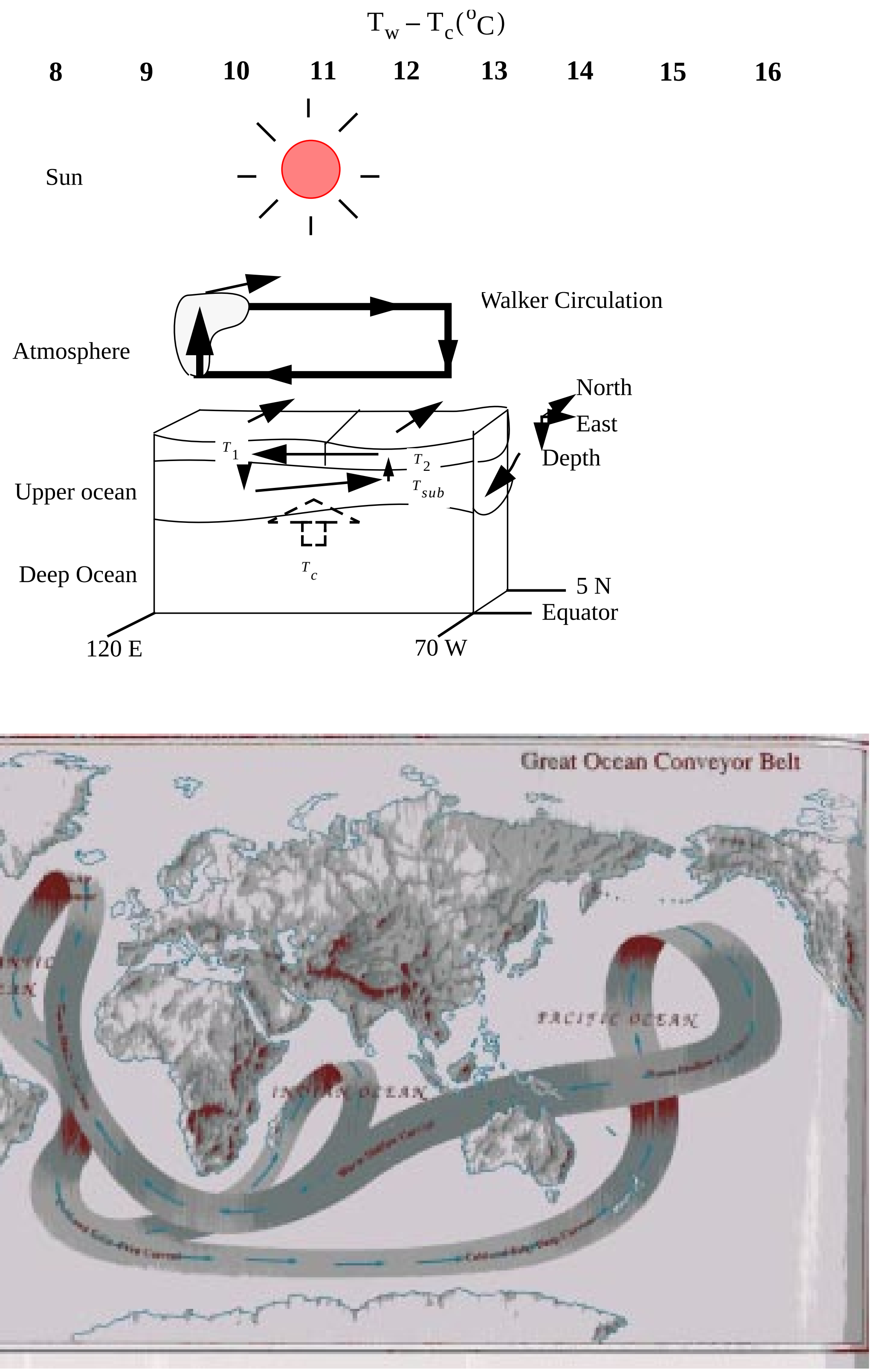


Fig. 2 A schematic diagram of the coupled model. Solid arrows represent the Walker circulation in the atmosphere and the equatorial wind-driven currents in the upper ocean. The latter are constituted by the upwelling in the east Pacific, the westward surface drift, the poleward surface drift, and the equatorial undercurrent. The large dashed arrow represents the background thermohaline circulation, which may be regarded as the Pacific branch of the global conveyor belt (shown at the bottom of the figure).

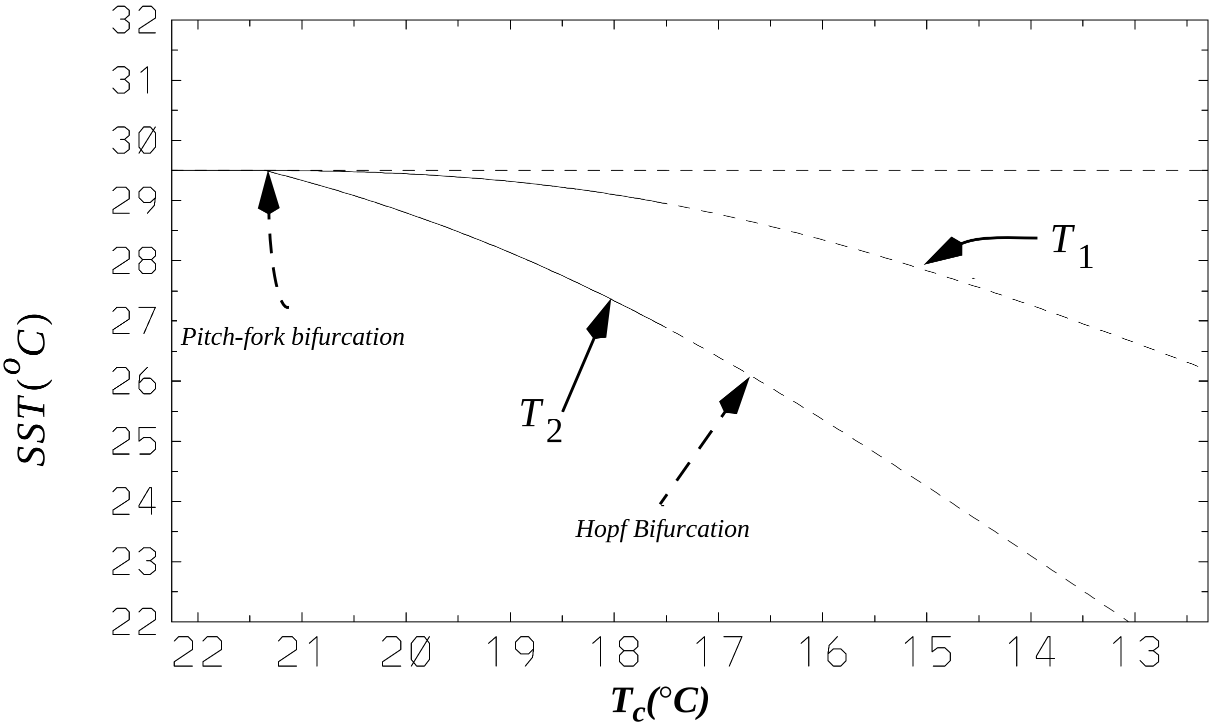
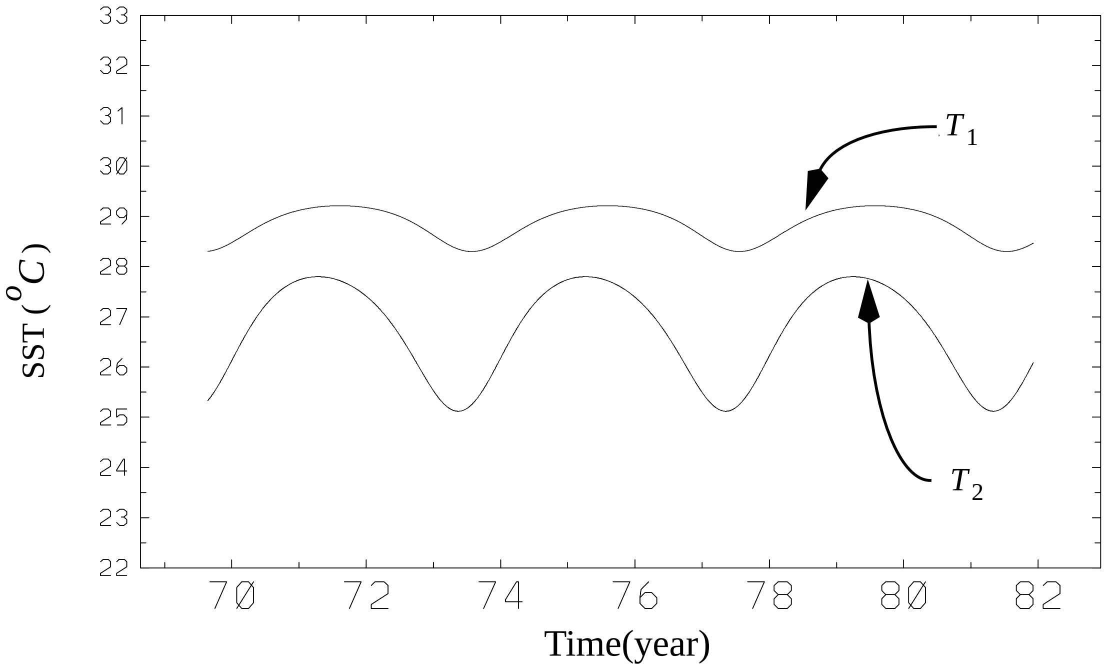


Fig. 3 Equilibrium SST of the coupled system as a function of deep ocean temperature T_c. Warm pool temperature T_w is fixed at 29.5°C. Dashed lines indicate that the solution exists, but unstable. After the Hopf Bifurcation, the system starts to oscillate. Parameters used are: $1/c$ = 150 days, $1/r$ = 300 days, $\alpha = 3.0 \times 10^{-8}\ K^{-1}s^{-1}$, $\alpha/b^2\ HH_1/(2H_2) = 11.5\ mK^{-1}$, $H^* = 65$ m, $H_1 = 50$ m, and $z_0 = 75$m.

8 9 10 11 12 13 14 15 16

(a)



(b)

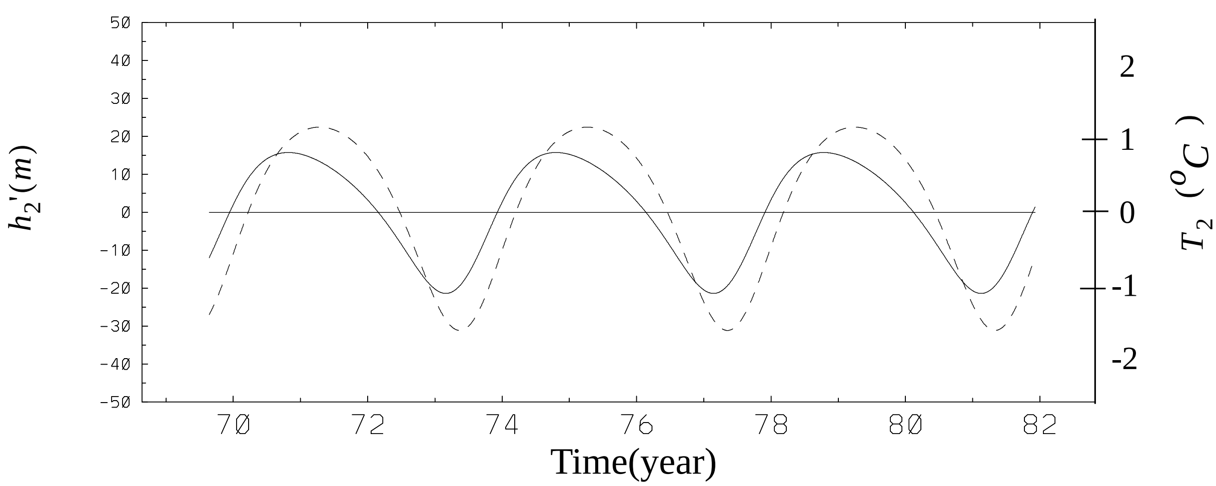


Fig. 4 Oscillations with T_c = 17.3°C. a: Variations of T_1 and T_2. b: Variations of h_2' (anomalies, solid line). Dashed line denotes anomalies of T_2.

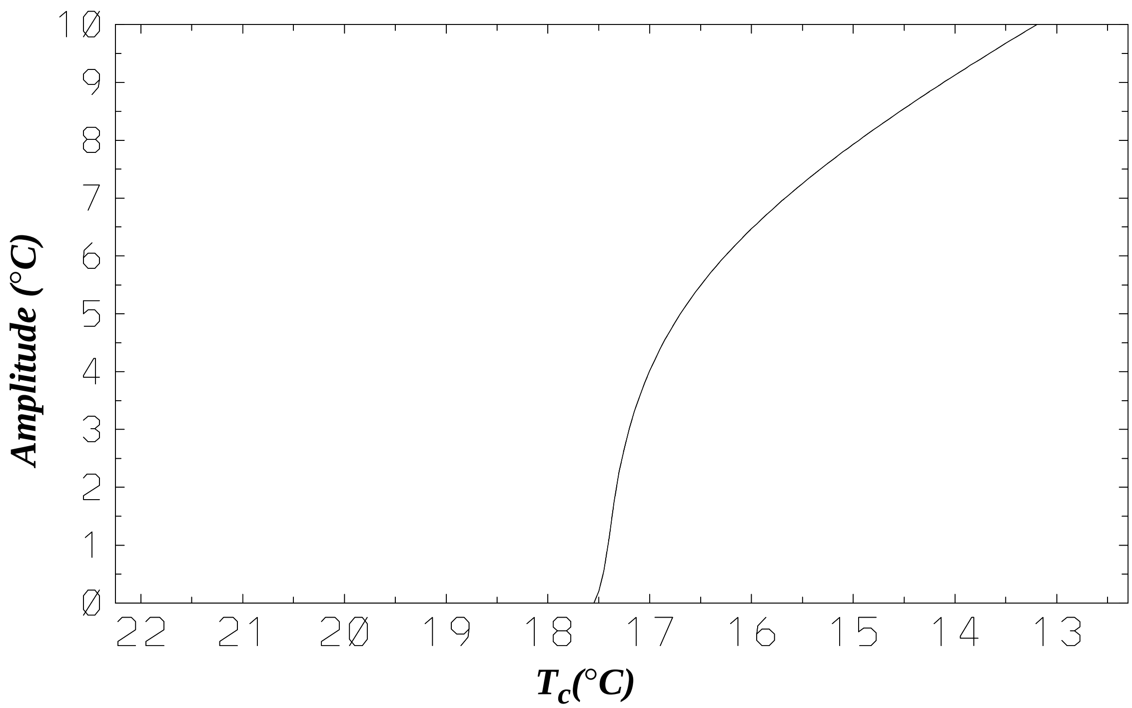


Fig. 5 Amplitude of the oscillation as a function of T_c. The amplitude is defined here as the half value of the difference between the maximum and minimum value of T_2. The value of T_w is fixed at 29.5°C. Other parameters are the same as in Figure 4.

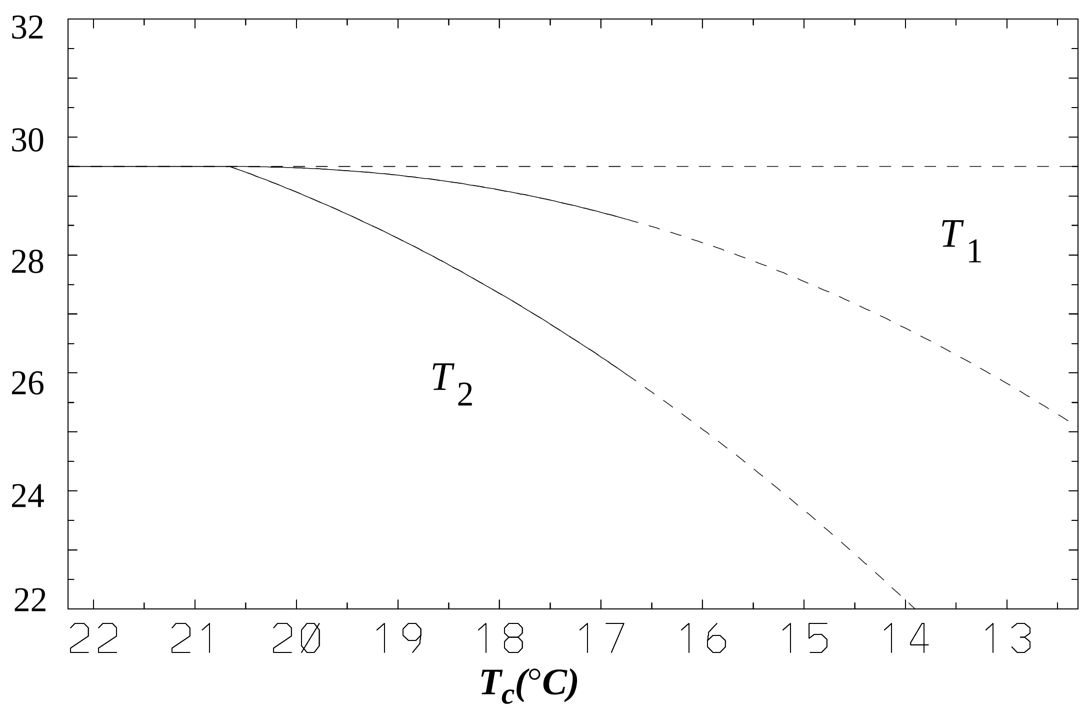


Fig. 6 Equilibrium SST of the coupled system as a function of T_c. Dashed lines indicate that the solution exists, but unstable. Different from Figure 3, Eq. (15) is used for calculating the value of T_{sub}. T_w *is fixed at 29.5*°C. Other parameters are: $1/c$ = 150 days, $1/r$ = 300 days, $\alpha = 3.0 \times 10^8 K^{-1}s^{-1}$, $\alpha/b^2 H_1/2H_2(1+H_1/H_2) = 12.5\ mK^{-1}$, $\Lambda = 17/24$, and $\gamma^*/H_2 = 1/150m^{-1}$.

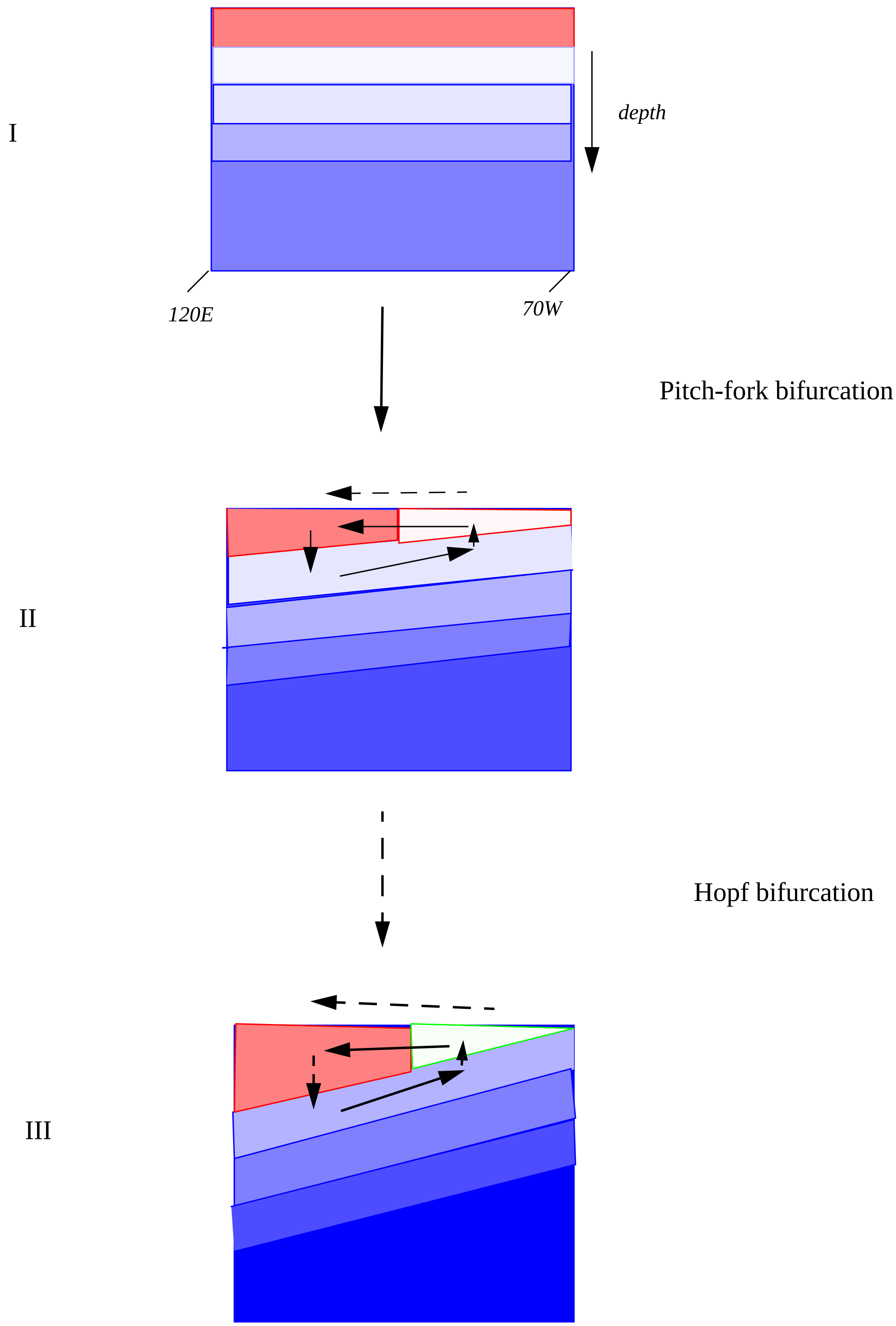


Fig. 7 A schematic illustration of the re-organization of the system as the deep ocean temperature becomes systematically colder. The dashed arrow represents the zonal wind. The solid arrows stand for the zonal branch of the wind-driven circulation.

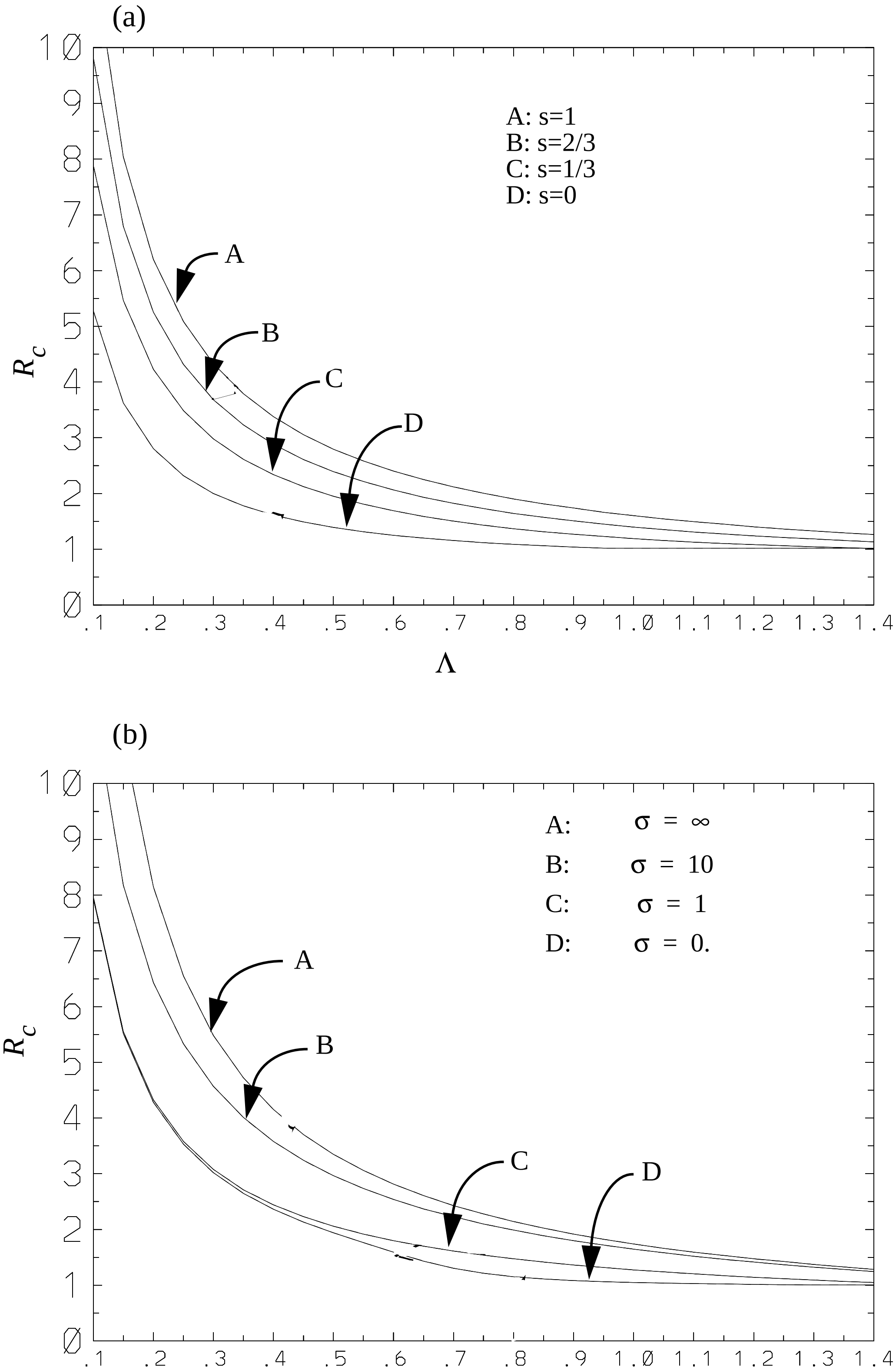
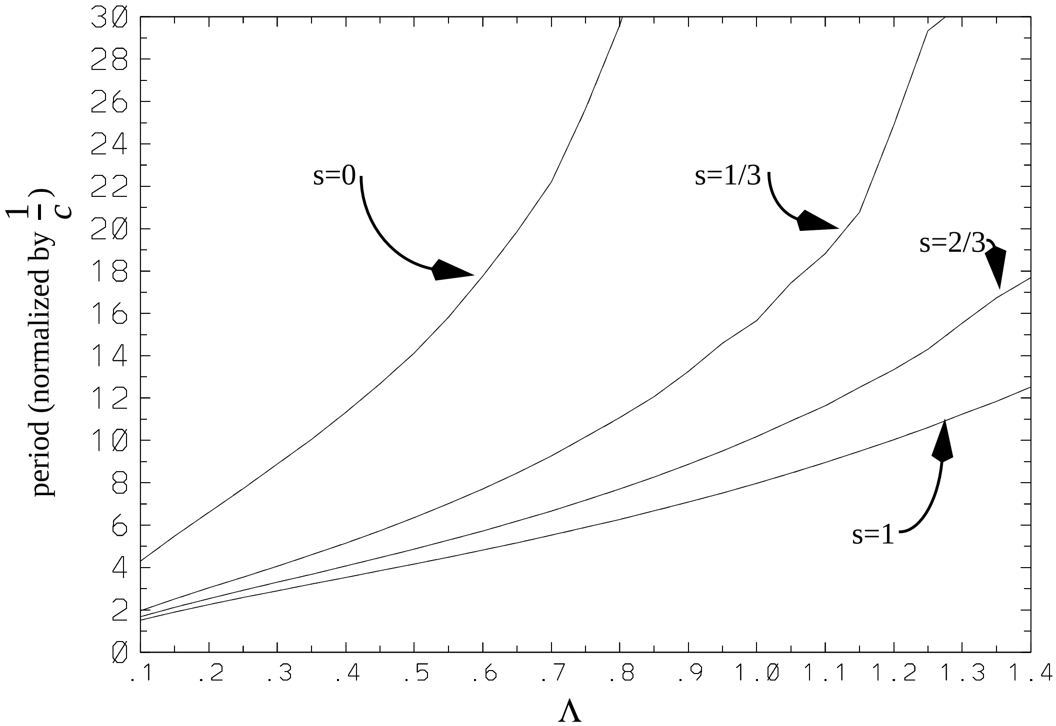


Fig. 8 (a) Dependence of the critical value of R (at which the Hopf bifurcation takes place) on the values of Λ and s. The value of σ is fixed at 1/2. (b) Dependence of the critical value of R (at which the Hopf bifurcation takes place) on the values of Λ and σ. The value of s is fixed at 1/3

(a)



(b)

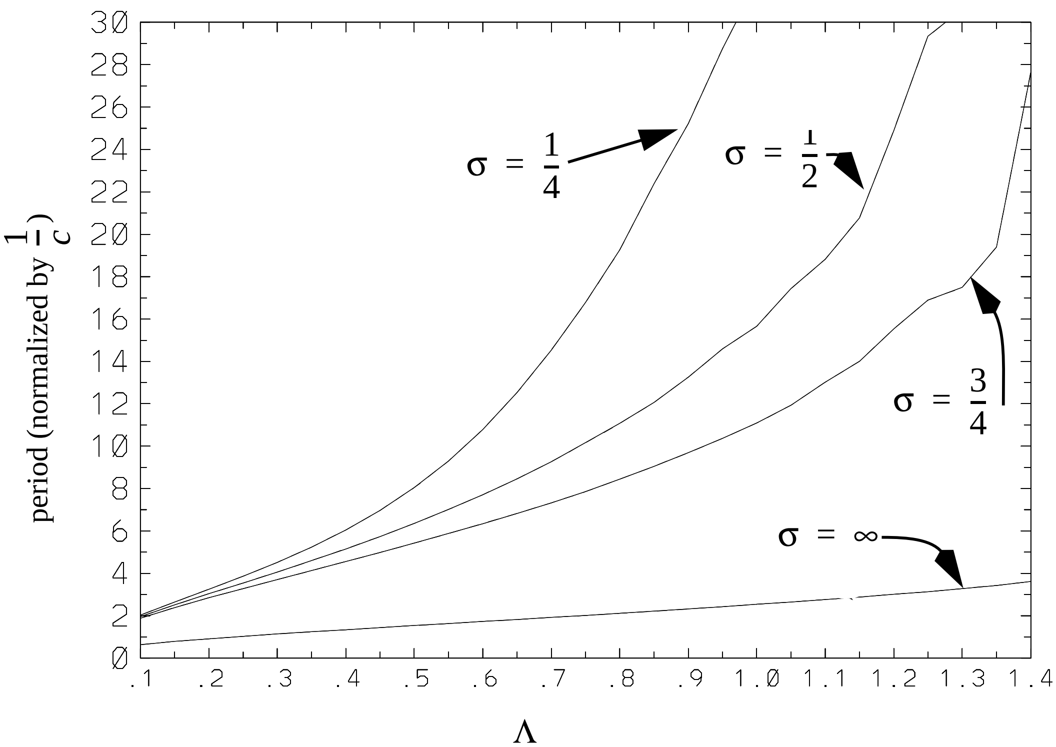


Fig. 9 (a) Dependence of the period of the oscillation on Λ and s. The value of σ is fixed at 1/2. (b) Dependence of the period of the oscillation on Λ and σ. The value of s is fixed at 1/3.

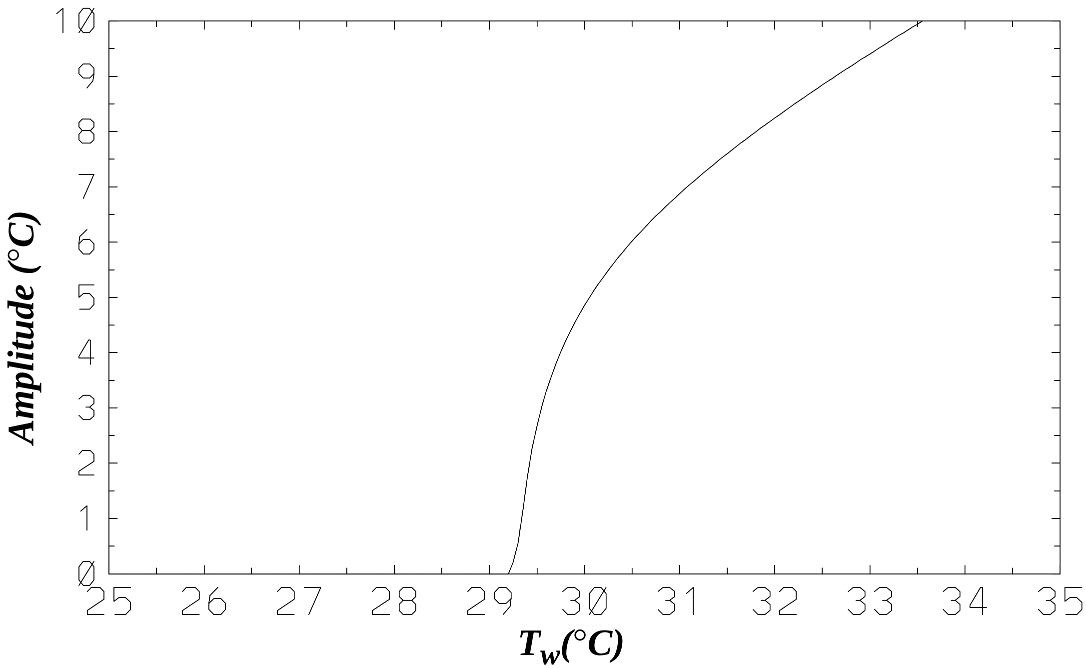


Fig. 10 Amplitude of the oscillation as a function of T_w. The amplitude is defined here as the half value of the difference between the maximum and minimum value of T_2. The value of T_c is fixed at 17.3°C. Other parameters are the same as in Figure 4.

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