Characteristics of the vertical baroclinic mode in the equatorial Pacific and its impact on the interannual and decadal variabilities

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Abstract

The characteristics of equatorial baroclinic modes are found in ocean assimilation data and the impacts of baroclinic modes on the variability of SST anomalies are analyzed in interannual and decadal time scales. The author developed a reliable air-sea coupled general circulation model (CGCM). The CGCM shows the observation-like mean field of sea surface temperature (SST), velocity, precipitation and the variance of interannual SST anomalies. The strategy for upgrading the performance of CGCM is discussed in terms of oceanic vertical baroclinic mode.

The baroclinic contributions to surface zonal current and sea level anomalies are investigated. The first three modes contribute mainly to spatial distributions in equatorial wave guide. The first mode exhibits the largest variability in the western Pacific, while the second and third modes have peaks in the central-eastern part of the basin. The reason for these modal distributions is because the thermocline slope has a strong zonal gradient in equatorial Pacific. The wind stress forcings are projected onto the each baroclinic mode with the increased stratification in the upper ocean: the increased (decreased) stability leads to well excitation of higher (gravest) baroclinic modes.

Since the characteristics of vertical modes, which are associated with subsurface change, vary in the tropical Pacific with time, it is a very important task to investigate whether such changes can be related to the variability of the ENSO. After the study of equatorial wave propagation for 1997 – 1998 and 1986 – 1988 El Niño
events in terms of baroclinic modes, it is suggested that the second baroclinic mode is essential to determine the amplitude and duration of warm event. In addition, at the transition time of 1997 – 1998 El Niño, the second mode also contributes to initiation of cold event. These results motivated the further study to predict better the interannual variability in terms of vertical modes.

The author investigated the change in baroclinic mode contributions at low frequencies. The vertical stratification increases substantially at upper levels in the tropical Pacific after the late 1970s, which is eventually associated with an increase of contributions for the higher-order baroclinic modes. Consequently, the dominant ENSO oscillation period increased from 2 – 3 years during 1960 – 1975 to 4 – 6 years after the late 1970s.

A newly coupled general circulation model is developed and the preliminary results are analyzed with respect to vertical baroclinic modes. The model simulates well the Pacific mean fields, particularly oceanic thermal structures which are important to capture better interannual variability. The simulated data shows that the variability of SST anomalies of CGCM are dominated by the second baroclinic mode that is due to the too diffused thermocline. We may, therefore, reasonably conclude that the more mixing processes are needed at upper levels to simulate well thermocline.

**Key words**: Vertical Baroclinic Mode, ENSO, Interannual Variability, Decadal Variability, Coupled General Circulation Model, Climate Change

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Chapter 1

Introduction

1.1 ENSO

The El Niño-Southern Oscillation (ENSO) is a dominant interannual variation in the tropical Pacific and also is the primary source of global climate variability [Philander, 1990; Ropelewski and Halpert, 1987]. El Niño is now recognized as one phase of a natural mode of oscillation that results from unstable interactions between the tropical Pacific ocean and the atmosphere; the other complementary phase is called La Niña. The typical cycles of ENSO have periodicity on an order of 2 – 6 years.

During El Niño, there is a dynamic adjustment of heat and mass between the western and eastern tropical Pacifics, producing a positive sea surface temperature (SST) anomaly in the eastern Pacific. In this situation, the trade winds are weak, and heavy rainfall is displaced eastward from the warm pool region toward the date line so that the west coastal zones of South America have severe floods. During La Niña,
the eastern tropical Pacific SST anomaly is negative, the trade winds strengthen and heavy rains fall mainly over the far western tropical Pacific. The Southern Oscillation (SO) describes a seesaw of sea level pressure in the Southern Hemisphere associated with the SST anomaly in the eastern tropical Pacific. When the warm (cold) SST anomaly exists in eastern tropical Pacific, unusually high (low) atmospheric sea level pressures develop in the western tropical Pacific and Indian Ocean regions, and unusually low (high) pressure develops in the southeastern tropical Pacific.

_Bjerknes_ [1969] first recognized that the SST anomaly in eastern equatorial Pacific could be amplified by air-sea interaction. He noticed that the decreases of equatorial easterlies associated with the decreased speed of Walker Circulation weakens the equatorial upwelling, thereby the eastern equatorial Pacific becomes warmer near the sea surface and also supplies heat to the atmosphere so that the Walker Circulation slows down. The cold SST anomaly also can be developed in the reversed way. While the concept of a positive feedback between SST and atmospheric circulation seems well supported by the evidence, it does not provide an explanation for the fundamental question of how the coupled system evolves from the warm state to the cold state and vice versa.

His work _[Bjerknes, 1969]_ stimulated numerous studies to understand the coupled dynamics of ENSO _[Wyrtki, 1975; Rasmusson and Carpenter, 1982; Anderson and McCreary, 1985; Cane and Zebiak, 1985; Hirst, 1986, 1988; Jin and Neelin, 1993; Wakata and Sarachik, 1991; Jin and Neelin, 1993; Kang and An, 1998]. Wyrtki [1975] realized that prior to El Niño, there was a rise (depression) of sea level (thermocline) in the western Pacific warm pool because the trade winds strengthened.
When the trade winds weakened, the piled-up warm water in the western Pacific would collapse and propagate eastward in the form of a Kelvin wave which leads to initiate El Niño. Wyrtki [1975] emphasized the evolution of the warm phase of ENSO and considered that El Niño is an isolated and sporadic occurrence.

Through more understanding about the important roles of oceanic waves in thermocline adjustment, a number of phase transition mechanisms were suggested [Schopf and Suarez, 1988; Suarez and Schopf, 1988; Battisti and Hirst, 1989; Jin, 1997a; Weisberg and Wang, 1997; Picaut et al., 1997].

According to the theory of the delayed action oscillator [Schopf and Suarez, 1988; Suarez and Schopf, 1988; Battisti and Hirst, 1989], the delayed negative feedback involving the reflection of oceanic equatorial waves at the western boundary is responsible for the phase reversal of the ENSO cycle, these waves being generated by the atmosphere response to increased or decreased SST in the eastern and central Pacific. A westerly zonal wind stress anomaly in the central Pacific produces a downwelling equatorial Kelvin wave that propagates rapidly to the eastern Pacific. In the eastern Pacific where the climatological thermocline is shallow, the propagated downwelling Kelvin wave produces a positive SST anomaly, giving rise to local air-sea coupling. A westerly zonal wind stress anomaly also makes the upwelling Rossby waves in the off-equatorial region that somewhat slowly propagate to the west. These Rossby waves propagate westward from the forcing region in the central Pacific and reflect at the western boundary, forming an upwelling Kelvin wave. This upwelling Kelvin wave propagates eastward to the eastern Pacific and leads to the termination of the warm phase and the initiation of the cold phase.
There also has been a recent revival of interest in the recharge-discharge of the equatorial heat content [Wyrtki, 1975, 1985; Cane and Zebiak, 1985; Jin, 1997a]. Jin [1997a] suggested the recharge-discharge process to describe the nature of the ENSO mechanism by using a simple conceptual model. This process emphasizes the importance of the buildup and release of zonal mean ocean heat content in the equatorial band for the phase reversal. The charge and discharge are generated by the non-equilibrium between the zonal mean ocean heat content and the wind stress forcing. For example, an initial positive SST anomaly induces a westerly wind forcing in the central Pacific. This wind stress promptly sets up the equatorial thermocline slope in the equatorial Pacific. The warm SST anomaly associated with deepening the thermocline in the eastern Pacific are growing by the positive feedback process. At the same time, the equatorial thermocline depth gradually reduces as the result of the geostrophic balance between zonal gradient of thermocline and meridional transport. This process can be viewed as the discharge of the zonal mean equatorial heat content. This discharge also gradually reduces the thermocline depth in the eastern Pacific and eventually leads to a cold event. This conceptual model can be easily tested by examining the equatorial mass or heat content budgets without the details of forced and free equatorial waves and their reflections.

Consistent with observation, Weisberg and Wang [1997] constructed an analogical oscillator model for ENSO, i.e., a western Pacific oscillator. This model emphasizes the importance of off-equator sea surface temperature and sea level pressure variations west of the date line for termination of growing SST anomalies. For instance, warm SST anomalies in the eastern Pacific produce the heating
in the west-central Pacific. This heating induces a pair of off-equator cyclones with westerly wind anomalies on the equator. These westerly winds act as a positive feedback for anomaly growth. On the other hand, the off-equator cyclones raise the thermocline there by Ekman pumping. Thus, a shallow off-equator thermocline anomaly expands in the western Pacific leading to a decrease in SST and an increase in sea level pressure. During the mature phase of El Niño, the off-equator high sea level pressure initiates equatorial easterly winds over the far western Pacific. These easterly winds cause upwelling Kelvin waves that propagate eastward as a forced ocean response providing a negative feedback. The western Pacific oscillator differs from the delayed oscillator, in which the wave reflection at the western boundary is not necessary for the negative feedback.

Both the above delayed action and the recharge-discharge oscillator emphasize the thermocline feedback associated with the vertical advection of anomalous subsurface temperature by the mean upwelling. However, the recent observational evidence indicated that the zonal advective process is also important for SST anomalies of interannual variability [McPhaden and Picaut, 1990; Picaut and Delcroix, 1995; Picaut et al., 1996]. The relative importance of these two processes (thermocline and advective processes) in the SST thermodynamics on the interannual time scale has been revisited through a heat budget analysis [Kang et al., 2001; Jin and An, 1999; An et al., 1999].

In addition to the advective process, the observational evidence that the eastern boundary reflection cannot be neglected for the ENSO mechanism leads to the advective-reflective model for the ENSO mechanism [Picaut et al., 1997] which can
be considered as a significant modification of the delayed action oscillator. It was based on the finding of an oceanic zone of convergence on the eastern edge of the western Pacific warm pool, which moves over thousands of kilometers along the equator in phase with ENSO [Picaut and Delcroix, 1995].

In this model, the equatorial wave reflection on the eastern boundary is more important than on the western boundary and zonal advection is more effective overall than vertical advection. The eastern boundary reflection of a downwelling Kelvin wave into a downwelling Rossby wave may be responsible for the shift of the El Niño into the La Niña by pushing the eastern edge of the warm pool westward. This pushing back attributes to the anomalous equatorial westward currents which are from the first meridional Rossby waves.

In addition to above mentioned phase transition mechanisms, the succession of the westerly wind bursts are fundamental to the onset and development of El Niño [McPhaden, 1999]. McPhaden and Yu [1999] also suggested that the rapid onset and strong strength of the 1997 – 1998 El Niño may have been related to the westerly winds associated with the Madden-Julian Oscillation (MJO) [Madden and Julian, 1971, 1972] originating from the Indian Ocean, which exited downwelling Kelvin waves an order of magnitude larger than those generated by western boundary reflections. As pointed out by Picaut et al. [2002], there is however not only one mechanism at work during the development and transition of El Niño.

Although many robust features of ENSO, i.e., spatial structure and temporal behavior, have been known [Rasmusson and Carpenter, 1982], the characteristics of each ENSO seem to be quite different from one to another, and the periodic time
between events can vary from 2 to 6 years. The behavior of the ENSO also seems to change from one decade to another [Gu and Philander, 1995; Wang and Wang, 1996; Torrence and Compo, 1998]. Multidecadal and century-long changes also have been found in proxy ENSO records [Cole et al., 1993; Diaz and Markgraf, 1992]. For instance, the two most strongest El Niño during the last century occurred in the past two decades, in 1982 and 1997.

There is no clear understanding of the mechanisms that lead to ENSO irregularity. The three hypothesis as sources for irregularity are (1) deterministic chaos within the nonlinear dynamics of the coupled system [Münnich et al., 1991; Tziperman et al., 1994; Chang et al., 1995], (2) uncoupled atmospheric random disturbances [Blanke et al., 1997; Kleeman and Moore, 1997; Kirtman and Schopf, 1998], and (3) changes in the background climate state [Gu and Philander, 1997; An and Wang, 2000; Fedorov and Philander, 2000; Pierce et al., 2000; Wang and An, 2001, 2002]. Without fully excluding the first two, the last one is seductive because there was a simultaneous occurrence of the Pacific decadal variation (PDV) and decadal variation of ENSO (EDV).

In late 1970s, the tropical Pacific experienced an abrupt warming [Nitta and Yamada, 1989; Zhang et al., 1997]. Going through that, the dominant period of the ENSO increased from 2–3 years during 1960–1975 to 4–5 years during 1980–1995 and ENSO became stronger [An and Wang, 2000]. It is also noticed that the SST anomalies associated with El Niño before 1980 showed westward phase propagation; since then, eastward phase propagation has been more common [Wang, 1995].

In the intermediate coupled model, the ENSO is sensitive to changes in the
background conditions of ocean [Kleeman et al., 1999], but how do the changes lead to modulate the ENSO has been a subject of debate. Fedorov and Philander [2000] also suggested that the changes in thermocline and zonal wind stress are the main cause of modulating behavior of ENSO.

The possibilities emphasizing the role of wind are suggested as the mechanism of ENSO modulation [An and Wang, 2000; Karspeck and Cane, 2002; Pierce et al., 2000; Wang and An, 2001, 2002]. An and Wang [2000] found that the longitudinal shift of wind forcing is the primary factor contributing to the change of ENSO. For instance, an eastward (westward) shift of the maximum westerlies in the equatorial Pacific results in a longer (shorter) period and larger (weaker) amplitude of ENSO. It is also suggested that the broadening (narrowing) of the meridional scale of wind stress anomalies leads to a longer (shorter) period.

Based on equatorial wave dynamics, zonal shift and meridional-scale changes in wind stress can change the basin wide thermocline adjustment times. In delayed action oscillator and recharge-discharge context, the time for Rossby waves reaching the western boundary is crucial for determining the period and amplitude of ENSO [Kirtman, 1997; Wang and An, 2001; An and Kang, 2001].

Wang and An [2001] illustrated that the Pacific climate shift may have altered ENSO behavior by changing background tropical winds and associated equatorial upwelling. The changes in equatorial wind may be connected to midlatitude SST anomalies. Pierce et al. [2000] showed that midlatitude SST anomalies are strongly correlated with changes in equatorial zonal wind stress in a coupled general circulation model. The changed trade wind system may alter the east-west slope of the
tropical thermocline and affect the decadal properties of ENSO.

1.2 Oceanic baroclinic mode

Recent observational and modeling studies have indicated that more than one baroclinic mode is necessary to simulate low frequency variability in the tropical Pacific [Kessler and McCreary, 1993; Dewitte et al., 1999; Delcroix et al., 2000; Shu and Clarke, 2002]. For instance, the 1997 – 1998 El Niño, the best observed so far [McPhaden, 1999], has a very strong sea level signal during its mature phase (~ 35 cm in the eastern Pacific) which cannot be explained by just the first baroclinic mode Kelvin waves. The single mode simulation forced by observed wind stress shows an amplitude for sea level anomalies on the order of 20 cm, but three-mode one exhibited the reasonable amplitude [Dewitte et al., 2002].

Shu and Clarke [2002] found that the first two vertical mode are needed to model the observed oceanic fields. When forced by observed wind stress, the ocean model showed that the two lowest baroclinic modes are necessary to successfully capture the equatorial thermocline depth, zonal current and SST anomalies in both central and eastern equatorial Pacific.

The inclusion of high-order vertical modes results in an increase of the contributions of the forced Kelvin waves to sea level anomalies. Because the higher the order of the mode has the finer the meridional scale, it has the larger the amplitude for the forced Kelvin wave for the same input of wind stress forcing [Dewitte, 2000]. This leads to a better agreement between the observation [McPhaden, 1999] and
linear simulations in equatorial Pacific [Dewitte et al., 2002; Shu and Clarke, 2002].

It is also showed that the predictive skill of the intermediate model is increased by including the high-order baroclinic mode [Dewitte et al., 2002]. Dewitte et al. [2002] argued that the model showed the larger forecast skill with three modes for the 1997 – 1998 El Niño because (1) the inclusion of higher-order mode produces more realistic initial conditions for the prediction. In particular, the sea level and current anomalies are better represented than in a one-mode simulation over off-equator region [Dewitte et al., 1999]. (2) the second mode with slower phase speed contributed to better capture the memory of the system [Dewitte et al., 2003] and it is more efficient in initiating the warming than a first baroclinic mode [Dewitte, 2000].

We would like to lay special emphasis on Dewitte [2000] and Yeh et al. [2001] in which high-order ocean baroclinic modes can set up a slower variability through the complex air-sea coupled instabilities. As we shall see later in the following chapters, these two studies helped to suggest a new theory about ENSO decadal modulation emphasizing the role of the baroclinic energy distribution associated with subsurface temperature variability.

Dewitte [2000] first showed that the mean vertical structure determines the relative contribution of zonal advection feedbacks and thermocline feedbacks which are responsible for changes in SST anomalies through the coupled instabilities [Hirst, 1986, 1988; Hao et al., 1993; Jin and Neelin, 1993; An and Jin, 2000, 2001; An and Kang, 2001; Kang et al., 2001]. In intermediate ocean-atmosphere coupled model including three baroclinic modes coupled to Gill-type atmosphere [Gill, 1980], the
first mode favors the free propagating Rossby waves and enhances zonal advection over vertical advection. Whereas the high-order modes tend to generate warm anomalies in the central Pacific and set an equilibrium between the atmosphere and vertical advection that leads to an unstable Kelvin mode propagating slowly eastward. So the more energy in high-order modes leads to strong El Niño as shown in 1997–1998 [McPhaden, 1999; Dewitte et al., 2003] because there are better realistic Kelvin waves.

It is also shown that the oceanic vertical structure is responsible for the ENSO period in the model [Dewitte, 2000; Yeh et al., 2001]. High-order modes, which propagate more slowly, are necessary in intermediate mode to get oscillation at a 3–4 yr period [Dewitte, 2000]. Only the gravest mode gives rise to biennial (2–3 yr period) oscillation. In a simple coupled model, the exclusive contribution of second mode and third one induce interannual (3–5 yr) and interdecadal (∼10 yr), respectively [Yeh et al., 2001].

Yeh et al. [2001] first investigated the impact of variability modes in the atmosphere on the oceanic baroclinic modes by using a statistical atmospheric model based on the singular value decomposition [Cassou and Perigaud, 2000; Kang and Kug, 2000]. The central mode favors the energetic second baroclinic mode in coupled process, which will then tend to set the dominant period of oscillation between 4 and 5 years. The off-equatorial mode, however, leads to larger contribution of the first baroclinic mode over the second one through the contribution of Rossby waves. The reader is invited to refer to Yeh et al. [2001] for definition of these two atmospheric modes.
Dewitte [2000] and Yeh et al. [2001] have stimulated study to evaluate the role of vertical mode on equatorial variability by using observational data. Their discussions have raised the issue on whether the oceanic subsurface variability is responsible for changes in ENSO, particularly in period and amplitude. Although the characteristics of the vertical modes vary in the tropical Pacific with time [Dewitte et al., 1999], little attention has been given to the point that the decadal variation of vertical structure can impact on ENSO variability particularly before and after the Pacific climate shift of the 1970s. Thus, it is noteworthy to test the role of baroclinic modes on ENSO variability in high-quality ocean assimilation data.

1.3 Coupled GCM

The coupled general circulation model (CGCM) is a useful tool to study the interaction between atmosphere and ocean in global scale. Essentially, this is model based on the primitive equations, and it include the basic physics required to reproduce the large scale motion in both atmosphere and ocean. Manabe and Bryan [1969] used a very simple ocean-continent configuration to simulate climate in a CGCM. CGCMs with more realistic geometry were developed in the 1970s [Manabe et al., 1975; Bryan et al., 1975].

Much of the early progress in coupled model were motivated by the desire to understand, and eventually forecast the ENSO, and great progress has been made in this area under the Tropical Ocean Global Atmosphere (TOGA) programme and its successor, CLIVAR-GOALS [Delecluse et al., 1998, for a review]. Many of
CGCMs have major errors in the mean temperature of the equatorial Pacific and its mean zonal gradient. Nevertheless, some of those models produced some interannual variability that resembled some aspects of the Southern Oscillation [Sperber et al., 1987; Meehl, 1990b]. ENSO-like interannual oscillation in a number of fully global models was reported with realistic spatial pattern and time scales. Some models have included enhanced resolution in order to resolve the equatorial wave guide [Roeckner et al., 1996; Guilyardi and Madec, 1997; Barthelet et al., 1998]. Surprisingly, even some coarse resolution models still exhibit quite realistic SST variability [Tett, 1995; Knutson et al., 1997; Timmermann et al., 1999a; Collins, 2000]. This may be due to some physical mechanisms other than equatorial waves are at play in ENSO, and such mechanism are working in the coarse resolution models [Neelin et al., 1998, for a review].

Though there is no clear connection between climate drift and the performance of a model to simulate interannual variability [Neelin et al., 1992], the comparison suggested that a correct simulation of tropical Pacific climatology is important to generate the best interannual oscillation. In a recent CGCM intercomparison, Mechoso et al. [1995] concentrated the analysis on the mean state in the equatorial Pacific and the mean seasonal cycle. The simulated equatorial cold tongue generally tends to be too strong, too narrow, and extend too far west in 11 models. There are generally too warm SSTs along the western coast of South America. This is accompanied in models by an unrealistic double intertropical convergence zone (ITCZ) straddling the equator over the eastern Pacific. Some models simulate the annual cycle quite realistically, although the seasonal cold tongue tends to appear
prematurely. Others overestimate the amplitude of the semiannual harmonic. The some models exhibit a mean trade wind stress that is too weak and shifted westward relative to the observations [Philander et al., 1992; Latif et al., 1993].

In addition to wind stress errors, many atmospheric GCMs also misrepresent the surface heat fluxes. The strong cold tongue can occur because the incoming solar radiation may be too weak at the bottom of atmosphere, the cloud cover can be too dense, or oceanic upwelling can be to high associated with strong currents. The warm bias in eastern Pacific SST, another common error in CGCMs, is partly related to the excess of solar heating received in the subtropics. Several sensitivity studies have tried to identify the response of the coupled interaction to the radiative scheme used of the atmospheric model [Ma et al., 1994]. In order to reduce excessive solar radiation in the subtropics in the eastern Pacific, Ma et al. [1996] prescribed a stratus cloud cover the ocean off the Peruvian coast. In this experiment, the SST beneath the prescribed cloud deck are reduced by up to 5 °K for decrease of solar radiation reaching the surface. The decrease in SST also results in stronger and more realistic asymmetries in SST distribution over eastern Pacific.

Because of the above biases in CGCMs, simulating realistic amplitude of variability in the tropical Pacific ocean has been a challenge for some time [Delecluse et al., 1998; AchutaRao and Sperber, 2002]. The simulated interannual behavior ranged from realistic to very low SST variability, with both propagating and standing modes occurring. The results from CGCMs provided evidence for different oscillation mechanisms and for the sensitivity of coupled models to apparently small background mean changes. The hypothesis drawn from the analysis of simplified
models must be confirmed with more comprehensive climate model, that is CGCM.

Although every model has different configurations from each other, most models show that ENSO SST variability is too weak and the center of variance is shift toward western Pacific. There are too weakly phase-locked to the annual cycle compared to the observation and the period of interannual is short to biennial [Robertson et al., 1995; Latif et al., 2001; AchutaRao and Sperber, 2002]. One possible reason for the weak of variability in CGCMs is that air-sea coupling strength is too weak; this could be due to a strong cold bias, weak atmosphere response to SST anomalies.

The equatorial mean thermocline plays a crucial role for determining the amplitude of the El Niño oscillation in coupled model experiments [Zebiak and Cane, 1987; Latif et al., 1993]. The sharpness of the equatorial thermocline is of course related in part to the vertical mixing process [Wilson, 2000, 2002]. Thus, the vertical diffusion is a possible candidate for differences between the El Niño simulated by coupled models. Yu and Schopf [1997] has explored how vertical mixing parameterizations influence the features of zonal currents in the equatorial Pacific using an isopycnal ocean model. They tested four different mixing schemes: PP scheme [Pacanowski and Philander, 1981], PGT scheme [Peters et al., 1988], CRB scheme [Chen et al., 1994] and Step scheme [Yu and Schopf, 1997]. Their results confirmed the need to get high surface mixing in the upper ocean layer, controlled by current shear and wind stress, and much lower mixing under the mixed layer. Recently, Meehl et al. [2001] demonstrated that the lower the value of the ocean background vertical diffusivity, the greater the amplitude of El Niño variability which is related primarily to a sharper equatorial thermocline.
Another alternative has recently been proposed by Noh and Kim [1999] in view of the observation of the microstructure of the upper ocean. The observation shows that turbulent kinetic energy (TKE) near the sea surface exhibits $\sim 2$ order larger magnitude than that expected from the wall boundary layer [Drennan et al., 1996; Terray et al., 1996]. Noh and Kim [1999] developed a new ocean mixed layer model taking these observational evidences into consideration. The model is a second-order turbulence closure model using eddy diffusivity, but it produces a well-mixed layer even under the stabilizing net flux, consistent with the observation. The well-known Mellor-Yamada model [Mellor and Yamada, 1982], however, results in excessively high SST under heating and leads to strong stratification in upper ocean [Rosati and Miyakoda, 1988]. Noh et al. [2002] tested the new ocean mixed model in an ocean general circulation model (OGCM), and they showed that SST simulation could be improved by new vertical mixing process compared to constant vertical mixing and the scheme of Pacanowski and Philander [1981]. Because the OGCM forced by prescribed wind forcing and restoring heat flux, there are no air-sea coupled processes by which the model may tend to drift, so further investigation of simulation experiment with a new vertical scheme [Noh et al., 2002] in fully coupled GCM are required.

1.4 The purpose of this study

The purpose of this dissertation is to find the characteristics of vertical baroclinic mode in equatorial Pacific and to investigate if ENSO variability can be associated
with the variability of the vertical baroclinic modes. A study of the oceanic baroclinic mode in a newly developed CGCM is also performed.

To do this, we first decomposed an ocean assimilation data and analyzed the vertical structures of the variability in the equatorial Pacific. We also suggested simple wave dynamical concepts that help us to understand the equatorial low frequency variability. The reasons for the different spatial distribution of baroclinic modes are suggested with consideration on zonal gradient of equatorial density field.

The Kelvin and first-meridional Rossby waves are derived for the first two more energetic baroclinic modes and found the relative roles of baroclinic mode on the initiation, development, and reversal of warm event during 1997 – 1998 El Niño. The impact of air-sea coupling on the interannual variability also is discussed in terms of baroclinic mode. We investigated the 1986 – 1988 El Niño of long duration, and compared these results with 1997 – 1998 event of short duration.

We also investigated the relationship between the variability of baroclinic modal energy distribution associated with change in stratification and the ENSO characteristics in decadal time scale. A special focus is on the climate shift in the tropical Pacific ocean around the mid-1970s. To approach to this, we first analyze the results from vertical decomposition of assimilation data in tropical Pacific. Then, an intermediate ocean and statistical atmosphere coupled model is used to test the hypothesis which we have from the analysis.

A CGCM composed of Ocean GCM and Atmosphere GCM is developed to study the characteristics and impact of the low frequency variability of the background stratification on the interannual variability. In despite of some common
biases in our CGCM, this model simulates the realistic interannual variability and gives the evidences for the impact vertical baroclinic mode on interannual variability.

The organization of this study is as follows. Chapter 2 presents the climatological features of equatorial Pacific in terms of vertical mode as well as some basic theoretical views. Chapter 3 draws upon these results to evolution of El Niño. Two El Niño events, 1986 – 1988 and 1997 – 1998, are analyzed in terms of vertical baroclinic mode. The relationship between the ENSO decadal change and variability in vertical baroclinic mode is demonstrated in Chapter 4. This chapter has already been reported in Moon et al. [2004]. Chapter 5 is devoted to the development of a coupled general circulation model and simulated features of the model. The features of baroclinic modes are also described. Summary and discussion are presented in last chapter.
Chapter 2

Vertical structure of an assimilation data in 1950-1997

2.1 Introduction

Most of simple ocean models with only one baroclinic mode are capable of describing the low frequency variability of the tropical Pacific when coupled to simple atmospheric modes [Zebiak and Cane, 1987; Battisti, 1988]. They retain the essential physics of the more complicated general circulation model (GCM) and allow for rapid calculation. In these models, the ocean is composed of two immiscible layers, each of constant density. The upper layer is warmer rather than deeper one and this two layers are separated by thermocline.

Recently, an attempt has been performed to include more vertical modes into the oceanic component in order to better capture the equatorial low frequency [Chen
et al., 1995; Dewitte, 2000; McPhaden and Yu, 1999]. This is driven by observational and modeling studies which suggest that the tropical Pacific vertical structure cannot be reduced to a first baroclinic mode [McCreary, 1981; McPhaden, 1999; Delcroix et al., 2000]. For instance, Boulanger and Fu [1996] showed that the Rossby wave amplitudes are decreasing while propagating westward from the eastern Pacific in TOPEX/POSEIDON data. This feature can be explained by vertical energy propagation of Rossby waves and suggesting the contribution of higher mode to wave energy distribution [Kessler and McCreary, 1993; Dewitte and Reverdin, 2000]. There is also larger amplitude of sea level height during El Niño mature phase than that is simulated from model with only first baroclinic mode [McPhaden, 1999; Dewitte et al., 2002]. The contribution of higher modes resulted in an increase of the forced Kelvin wave and larger amplitude of sea level height. For these reasons, investigating the role of long equatorial waves associated with their vertical structure remains of great interest.

In this chapter, we perform the decomposition into vertical modes of the assimilation data at each grid point and time step. Dewitte et al. [1999] have also studied the various aspects of the surface features of the baroclinic contributions to the sea level and zonal current anomalies in ocean general circulation model (OGCM). They showed that the zonal current variability have more energy in gravest mode over western Pacific, whereas it is largest in the eastern Pacific for the second and third baroclinic mode contribution. We follow Dewitte et al. [1999], but use the more reliable and period extended oceanic data.

The vertical structure can be decomposed into an infinite series of vertical
normal modes corresponding to eigenfunctions of a Sturm-Liouville equation. The associated eigenvalues determine the phase speed, or equivalently the deformation radius of each mode. A good place to start is investigating the relative importance of baroclinic modes associated with a rich structure to the density field at equator. We focus on the spatial structure of the variability for the different vertical modes and how this could be influenced by spatial variability in the density structures. The zonal slope of the equatorial thermocline is the most prominent feature and the associated density field can affect on modal variability of equator. We also investigate the interannual and decadal variation of the vertical baroclinic modes. Although the subsurface structure vary in the tropical Pacific with time [Dewitte et al., 1999], little attention has been given to this point.

The chapter is organized as follows. Section 2.2 presents the description of the Simple Ocean Data Assimilation (SODA) data. Section 2.3 is devoted to the method of decomposition the vertical structure. The modal features of vertical structure for zonal current anomalies are described in section 2.4. Section 2.5 gives the descriptions for phase speed and projection coefficient. We also performed the analysis for idealized profiles of buoyancy frequency. We described these results in section 2.6. A summary and a discussion are then given in last section.

2.2 SODA data

We used ocean data (temperature, salinity, sea level and currents) from Simple Ocean Data Assimilation (SODA) system for the period of 1950 – 1997. The
basic algorithm is described in two papers [Carton et al., 2000a,b]. SODA uses an ocean model based on Geophysical Fluid Dynamics Laboratory MOM2 physics [Pacanowski, 1995] with a resolution of $2.5^\circ \times 0.5^\circ$ (lon.-lat.) horizontal resolution in the Tropics, expanding to $2.5^\circ \times 1.5^\circ$ resolution at midlatitude ($62^\circ$N and $62^\circ$S). Bottom topography is included. The model has 20 levels in the vertical, with 15-m resolution in the upper 150 m (Table 2.1).

The main datasets to constrain the model forecast are the hydrographic data contained in the World Ocean Atlas [Levitus et al., 1994]. SODA data have average temperature and salinity errors in the upper 500 m of 0.70$^\circ$C and 0.092 psu with comparison to a series of global hydrographic sections.

### 2.3 Vertical mode decomposition

The ocean and atmosphere are thin sheets of fluid so that their horizontal scale of motions is very much larger than vertical scale. We can easily know this, in atmosphere, as the vertical component of wind is very weak relative to the horizontal wind [Holton, 1992]. For such reason, there are certain simplifications that can be made, i.e., separation of variables technique; the solution of motion can be expressed as a sum of normal modes, each of which has a vertical structure and behaves in the horizontal dimension and in time in the same way as does a homogeneous fluid. We derived this idea of normal modes for continuously stratified flat-bottomed ocean in Appendix A with assumption of no-rotating frame. This decomposition can be applied even when rotation effects or forcing are introduced, and also thus provides a
Table 2.1: Vertical grids of data from the Simple Ocean Data Assimilation system

<table>
<thead>
<tr>
<th>Level</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>7.5</td>
</tr>
<tr>
<td>2</td>
<td>22.5</td>
</tr>
<tr>
<td>3</td>
<td>37.5</td>
</tr>
<tr>
<td>4</td>
<td>52.5</td>
</tr>
<tr>
<td>5</td>
<td>67.5</td>
</tr>
<tr>
<td>6</td>
<td>82.5</td>
</tr>
<tr>
<td>7</td>
<td>97.5</td>
</tr>
<tr>
<td>8</td>
<td>112.5</td>
</tr>
<tr>
<td>9</td>
<td>127.5</td>
</tr>
<tr>
<td>10</td>
<td>142.5</td>
</tr>
<tr>
<td>11</td>
<td>157.5</td>
</tr>
<tr>
<td>12</td>
<td>190.61</td>
</tr>
<tr>
<td>13</td>
<td>276.27</td>
</tr>
<tr>
<td>14</td>
<td>443.79</td>
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<td>15</td>
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<td>1099.46</td>
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<tr>
<td>17</td>
<td>1598.79</td>
</tr>
<tr>
<td>18</td>
<td>2201.30</td>
</tr>
<tr>
<td>19</td>
<td>2885.60</td>
</tr>
<tr>
<td>20</td>
<td>3622.50</td>
</tr>
</tbody>
</table>
useful information of low frequency variability in equatorial Pacific [Lighthill, 1969; Cane, 1984; McCreary, 1985; Philander, 1978; Shu and Clarke, 2002].

To find solutions over vertical modes, we use the formalism developed in Cane [1984]. If the vertical dependence is separable, then the variables describing the oceanic motions may be written as a sum of vertical standing modes (with ignoring the barotropic mode $n = 0$),

$$\begin{bmatrix} u(x, y, z, t), v(x, y, z, t), \rho^{-1} P(x, y, z, t) \end{bmatrix} = \sum_{n=1}^{\infty} \begin{bmatrix} u'_n(x, y, t), v'_n(x, y, t), gh'_n(x, y, t) \end{bmatrix} A_n(z)$$  \hfill (2.1)

The vertical structure functions $A_n(z)$ are the solutions to an eigenvalue-eigenfunction problem that depends on the stratification $N(z)$ and on the top and bottom boundary conditions, namely $A_n(z) = \partial_z G_n(z)$ and

$$\partial_{zz} G_n(z) + \frac{N^2(z)}{c_n^2} G_n(z) = 0$$  \hfill (2.2)

Here, $N(z) = \left( -g \rho^{-1} \partial_z \rho \right)^{1/2}$ is the Brunt-Väisälä frequency and $c_n$ are the wave phase speeds. $G_n(z)$ are the vertical structures for the vertical velocity. The rigid-lid and flat-bottom boundary conditions applicable to the baroclinic solutions of interest here are that the vertical velocity vanish at the sea surface and ocean bottom; that is,

$$G_n = 0 \quad \text{at} \quad z = 0$$  \hfill (2.3)

$$G_n = 0 \quad \text{at} \quad z = -D,$$  \hfill (2.4)

where $D$ is the ocean depth. Equation (2.2) is differentiated by forward scheme and the system can be solved as a eigenvalue problem [Pryce, 1993].
For each \( u'_n, v'_n, \) and \( h'_n \) satisfy the shallow water equations for an ocean of depth \( H_n \) (equivalent depth for \( n \)th mode defined by \( c^2_n/g \)). By nondimensionalizing the shallow water system by distance and time scales as shown in Table 2.2, we obtain the same formal equations for each mode

\[
\frac{\partial u_n}{\partial t} - yv_n + \frac{\partial h_n}{\partial x} = \tau^x \tag{2.5}
\]

\[
\frac{\partial v_n}{\partial t} + yu_n + \frac{\partial h_n}{\partial y} = \tau^y \tag{2.6}
\]

\[
\frac{\partial h_n}{\partial t} + \frac{\partial u_n}{\partial x} + \frac{\partial v_n}{\partial y} = 0 \tag{2.7}
\]

The (2.5), (2.6) and (2.7) describe the horizontal structure of the motion. No stratification dependent parameters appear in these equations but since the length and time scales depend on the \( c_n \), the response will be different for each mode. It is known that \( H_n \) is related to the vertical scale of the motion in terms of \( c_n \), which therefore related to the horizontal scales such as the radius of deformation: the larger the vertical scale, the wider the equatorial waveguide.

The set \( (A_n) \) is orthogonal so that expressions in (2.1) is indeed possible. To determine the oceanic response to given forcing it is then necessary to solve, for each mode \( A_n \), the shallow-water equations with the appropriate value for the equivalent depth and with the appropriate projection of the forcing function onto that vertical mode. The wind stress forcing in shallow equation system can be expressed as a sum of vertical modes if it acts as a body force. Suppose the ocean circulation is forced by a wind stress divergence \( \partial_z \tau(x, y, z, t) \); then \( n \)th shallow water system is
driven by a body force

\[ \tau_n(x, y, t) = \frac{1}{D} \int_{-D}^{0} \partial_z \tau A_n(z) dz \]  

and \( A_n(z) \) has been normalized so that

\[ \int_{-D}^{0} A_n^2(z) dz = D \]  

Assuming the stress is confined to a well mixed surface layer (equivalent to the \( A_n(z) \) are constant in this layer) and the wind stress vanished below the mixed layer, then (2.8) reduces to

\[ \tau_n(x, y, t) = \tau(x, y, 0, t) A_n(0) D^{-1} \]  

where \( \tau(x, y, 0, t) \) is the surface wind stress. (2.10) suggests the contribution each mode makes to the ocean circulation depends strongly on the surface value of the horizontal structure function, \( A_n(0) \).

We shall now look more carefully into the results from vertical decomposition of oceanic data. Figure 2.1a presents the vertical profiles of temperature and salinity in the equatorial Pacific (180°, 0°) at January 1990. The corresponding vertical distribution of stratification is also given in Figure 2.1b. The equation of state defined by the Joint Panel on Oceanographic Tables and Standards is used to get the density for Brunt-Väisälä frequency [Gill, 1982; UNESCO, 1983]. There is a rapid drop in temperature at upper ocean (\( \sim 150m \)). Whereas the salinity shows the slightly change only at this level, which supporting the Boussinesq approximation in equatorial Pacific (see the Appendix A). Below the 400m, temperature has a nearly constant value representing the cold deep water. Figure 2.1b shows the low
Figure 2.1: Vertical profiles of the (a) temperature (solid) and salinity (dashed) in the equatorial Pacific (180°, 0°) at January 1990. (b) The corresponding squared Brunt-Väisälä from temperature and salinity profiles. We used the profiles from SODA.
amplitude of $N^2$ at upper-most ocean indicating the well mixed surface layer. There is a marked increase in $N^2$ between 150 m and 300 m. This indicates the location of the core of the thermocline in the central Pacific. $N^2$ shows a rapid drop and has almost constant magnitude below 350 m, which is in consistent with the profile of temperature.

Figure 2.2 shows the first three eigenfunctions $A_n(z)$ corresponding to the $N(z)^2$ profile of Figure 2.1b. The phase speed, $c_n$, gives us more information of properties of equatorial wave. Table 2.2 gives values of equivalent depth $H_n$, surface amplitude $A_n(0)$, the gravity wave speeds $c_n = (gH_n)^{1/2}$, the equatorial inertial time $(\beta c_n)^{-1/2}$, and the equatorial radii of deformation $(c_n/\beta)^{1/2}$ from decomposition.

As the vertical mode number $n$ increases, the number of nodes of the vertical mode increase so that $n$ can be regarded as a vertical wave number (Figure 2.2). It is evident from this figure that vertical mode is related to the vertical scale of the motion, which is therefore related to the horizontal scaled such as the radius of deformation (Table 2.2). For example, the larger the vertical scale, the wider the equatorial waveguide. Note that $A_n(z)$ is almost constant above 150 m representing the mixed layer again. The wave speed $c_n$ decreases with mode number as shown in Table 2.2 and other parameters are changed in consistent with $c_n$. We also easily find this behavior of $c_n$ in mathematical solution (see B.15) derived with help of WKB theory [Morse and Feshbach, 1953]. The mode $n = 0$, which we ignored, known as the barotropic mode, has an equivalent depth equal to the actual depth of the ocean. The barotropic mode is unaffected by the stratification of the ocean and that its horizontal velocity components are independent of depth. The radius of
Figure 2.2: The eigenfunctions \((n = 1, 2, 3)\) for baroclinic waves when buoyancy frequency \(N\) varies with \(z\) as shown in Figure 2.1b. See Table 2.2 for corresponding eigenvalues and other parameters.
Table 2.2: The values for various parameters of the modal problem using a mean temperature and salinity profile from SODA (January 1990 and 180°, 0°).

<table>
<thead>
<tr>
<th>n</th>
<th>Equivalent depth $H_n$(cm)</th>
<th>Surface amplitude $A_n(0)$</th>
<th>Wave speed $c_n = (gH_n)^{1/2}$(m s$^{-1}$)</th>
<th>Time scale $(\beta c_n)^{-1/2}$ (days)</th>
<th>Rossby radius $(c_n/\beta)^{1/2}$(km)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>4</td>
<td>5</td>
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<td>1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>3</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

30
deformation of this mode is so larger than those of baroclinic modes. For example, if the total depth of ocean is 4 km, then the barotropic wave speed is approximately 200 m s$^{-1}$ and the corresponding length scale is 3000 km.

Because we are interested in interannual variability of zonal current, zonal current ($u$) and pressure ($p$) anomalies are estimated relative to the climatological monthly mean of the 1950–1997 data. These field are the projected on the vertical structure functions $A_n(x,0,z,t)$. We do not consider the meridional change in $A_n$ as can be seen in Dewitte et al. [1999]. This can be expressed by the formula:

$$< q(x,y,z,t) | A_n(x,z,t) > = \int_0^H q(x,y,z,t) A_n(x,z,t) dz.$$  

The method is detailed by [Dewitte et al., 1999].

### 2.4 Zonal current anomalies

In this section, we discuss the contribution of the different vertical modes to the zonal current anomalies (ZCA). The root mean square (rms) variability of surface zonal current anomalies over the 48-yr (1950–1997) is present in Figure 2.3. Large values are confined within 5°S – 5°N, with maximum variability located in western Pacific and eastern Pacific. The variability of wind stress has a maximum value at off-equatorial central Pacific, so that rms peak in surface zonal current anomalies over western Pacific is due to local forcing of Kelvin and Rossby waves (see Figure 4.2j). It may be safely stated that the variability of ZCA over equatorial eastern Pacific is reflection of equatorial long ocean waves. Note that there is little local wind forcing over the eastern Pacific.
Figure 2.3: The spatial distribution of variability (rms) of surface zonal current anomalies for the period of 1950-1997. Units are cm s$^{-1}$, contour interval is 6 cm s$^{-1}$, and shading is above 18 cm s$^{-1}$.
Figure 2.4 shows the contribution of the first three modes to ZCA. The meridional scales for each mode are different. For mode 1 large values, for example greater than 6 cm s\(^{-1}\), extend meridionally to 5\(^\circ\)S - 5\(^\circ\)N, whereas variance of mode 2 is more confined near the equator. This behavior is most prominent in mode 3. No doubt, decreasing of meridional scale as increasing mode number is associated with Rossby radius of deformation (Table 2.2). In this study, we will concentrate on the first three mode because these three modes explain 85.4% of the total surface current variance over the eastern Pacific (90\(^\circ\)W - 150\(^\circ\)W) (Figure 2.5). This ratio is decreasing to 72.1% over western Pacific (140\(^\circ\)E - 180\(^\circ\)E).

It is also interesting that the spatial distributions are different from mode to another. For mode 1, maximum variance is centered in the western and central equatorial Pacific, whereas the contribution of the second baroclinic mode is more important in the eastern equatorial Pacific. This features of the vertical modes agree to Dewitte et al. [1999], in which data from OGCM simulation are used. We must draw attention to strong zonal gradient in equatorial thermocline depth as the reason of modal distribution. Later we shall try to give a more discussion of this problem.

It is desirable to show the results of pressure from decomposition before moving on to the next. Figure 2.6 displays the maps of rms for the baroclinic mode contribution to sea level anomalies. As discussed in the case of ZCA, the variability for the first mode is largest in the western Pacific, the second and third mode is more active in the central and eastern Pacific. As we shall see later in the next chapter, this modal distribution can be associated with the evolution of El Niño. It also
Figure 2.4: The rms variances of zonal current anomalies for the contribution of (a) the first mode, (b) the second mode and (c) the third mode for the period of 1950-1997. Units are cm s$^{-1}$, contour interval is 3 cm s$^{-1}$.
Figure 2.5: The rms variances along the equator for the total zonal current anomalies (solid), mode 1, mode 2, mode 3 contribution, and the sum of the first three mode (dashed). The ratios, sum(1+2+3)/total, are 85.4% (eastern Pacific) and 72.1% (western Pacific).
seems reasonable to say from Figure 2.6 that the gravest mode is associated with Rossby waves and the high-order modes are related to Kelvin waves. We described this point in Chapter 4, Appendix C and D again.

Figure 2.7 shows the power spectrum of ZCA of the first three modes contributions to ZCA, all quantities being averaged over NINO3 (150°W – 90°W, 5°S – 5°N) and normalized by mean and standard deviation. All three modes present low-frequency around 50 months peak. This low-frequency may be associated with ENSO. Obviously, mode 1 has relatively more variance at higher frequency (between 7 and 10 months). Evidences for the existence of near-annual mode in the tropical climate system was in fact found in a number of studies. For examples, in Zebiak and Cane [1987] a fast mode, called the mobile mode [Mantua and Battisti, 1995; Perigaud and Dewitte, 1996], can be identified by 9 – 12 month period. Recently, Jin et al. [2003] and Kang et al. [2004] suggested that this fast mode can be understood in terms of a coupled Pacific basin mode and that it has important implications in understanding and predicting ENSO. Figure 2.7 indicates that the fast mode is associated with the first baroclinic mode in the equatorial Pacific ocean. It also tells us that the co-existence of this fast mode and the relative slow ENSO mode. The longitude-period distribution of power spectrum also supports this features (Figure 2.8).
Figure 2.6: The same as Figure 2.4 except for the sea level height anomalies. Units are cm. Note that the contour intervals are different from each mode. The intervals of mode 1, mode 2 and mode 3 are 1 cm, 1 cm and 0.2 cm, respectively.
**Figure 2.7:** Energy spectrum of NINO3 (150°W − 90°W, 5°S − 5°N) averaged zonal current anomalies of mode 1 (solid), mode 2 (dashed) and mode 3 (dotted). Each zonal current anomalies is normalized by mean and standard deviation.
Figure 2.8: The longitude-period distribution of power spectrum for the baroclinic contribution to zonal current anomalies (5°S – 5°N). Upper-left panel is for total zonal current anomalies and upper-right one is for first baroclinic zonal current anomalies. Lower-left and lower-right ones are for second and third baroclinic zonal current anomalies, respectively.
2.5 Phase speed and projection coefficient

Low-frequency linear waves have the propagating phase speed which can be given by eigenvalue $c_n$ in (2.2). Figure 2.9 shows the averaged phase speed at equator for the 1950–1997 period as a function of longitude for the first three baroclinic modes. The mean phase speeds at central Pacific (160°W, 0°) for mode 1, mode 2 and mode 3 are 2.63 cm s$^{-1}$, 1.60 cm s$^{-1}$, and 0.98 cm s$^{-1}$, respectively. Note that these phase speeds are lightly differences from those of Table 2.2, which represents the subsurface changes in thermal structure. Most of all, there are sharp discontinuities in phase speed over western Pacific because of a realistic bottom topography. These discontinuities are disappeared when a uniform 2200 m ocean depth is used (squared lines in Fiugre 2.9).

There are also difference in phase speed between the central Pacific and the eastern Pacific that is due to zonal gradient in stratification. The effects of longitudinal variations of the stratification are manifest as the decrease of $c_1$ from about 2.8 m s$^{-1}$ near the western boundary to 2.4 m s$^{-1}$ near the eastern boundary. This east-west asymmetry is an indication of stronger stratification in the western basins suggested by the WKB approximation (B.15). The shorter-scale variations (near 160°E and 100°W) of $c_n$ in Figure 2.9 represent the topographic effects also noted from the WKB solution; the gravity wave phase speed decreases over shallow water. (B.15) also imply that the phase speed is depend on the vertical integrated buoyancy frequency. Theory suggests that if $N$ does not change with oceanic depth then $c_n$ is defined as $NH/n\pi$. This relationship is also presented in (6.11.1) of Gill [1982].
Figure 2.9: Averaged phase speeds for the 1950 – 1997 period as a function of longitude for the first three baroclinic modes. Solid line is for mode 1, dashed line is for mode 2 and dotted is for mode 3. Squared markers indicate that phase speed with a flat topography 2200 m depth is assumed. Units are m s$^{-1}$. 
Therefore, the phase speed is usually increasing (decreasing) from its average value during a warm (cold) event as expected from the temporal changes in stratification of the upper ocean. In this way, we would like to predict the higher phase speed in equatorial eastern Pacific than in western Pacific, because the equatorial eastern Pacific has sharp thermocline (Figure 2.10). But Figure 2.9 shows the opposite distribution of phase speeds between western Pacific and eastern Pacific. We can find the answer of this problem in vertical structures of buoyancy frequency from western and eastern equatorial Pacific (Figure 2.11). As mentioned above, the total integrated buoyancy frequency determine the equatorial wave speed. Clearly, the vertical summed \( N \) has a larger amplitude at western Pacific than eastern Pacific that is supporting the results of Figure 2.9. Note that the sharper thermocline is over eastern Pacific, the broader one is western Pacific.

At this point, we should consider the WKB solution of vertical baroclinic mode with its failure; WKB solution can be from (B.14). We illustrated the accuracy of WKB approximation in Figure 2.12 for the first three baroclinic modes from long-term averaged density profiles at equatorial Pacific (184.5°E, 0°). It is shown that the difference between the WKB solutions and numerical eigenfunctions is largest in the upper ocean where the vertical scale of \( N(z) \) is short. This is consistent with the WKB assumptions that restrict the validity of the approximation to regions of the water column where the scaled rate of change of the environment is small compared with the scale vertical wave number of the eigenfunction. These assumptions are not valid in the upper ocean where \( N(z) \) increases rapidly from a small value near the surface to a largest at the thermocline and then decreases quickly below the
Figure 2.10: The longitude-depth distribution for averaged temperature along the equator during 1950–1997. Units are °C. Temperatures between 18°C – 22°C are shaded for clear thermocline.
Figure 2.11: The vertical profiles of buoyancy frequency ($N^2$) averaged during 1950 – 1997 at two equatorial Pacific regions. Solid represents the value at western Pacific (170°E, 0) and dashed does at eastern Pacific (100°W, 0). The profiles are represented only above 500 m for comparison.
thermocline. The WKB eigenvalues are less sensitive than the WKB eigenfunctions to the validity of the assumptions of the WKB approximation.

There is another thing to note when we consider the vertical structure in mind, that is projection coefficient [Lighthill, 1969] defined as

\[ P_n = \frac{150}{\int_{-H}^{0} A_n^2(z)\,dz} \]  

This \( P_n \), qualitatively same as \( A_n(0) \) (see 2.10), is the non-dimensional coefficient representing the projection of the wind forcing onto the \( n \)th mode. Figure 2.13 shows mean values of \( P_n \) along the equator for the first three modes. Not surprisingly, \( P_n \) is high in the western Pacific for first baroclinic mode, whereas the higher modes (mode 1 and 2) have larger values than mode 1 near the western boundary. Although zonal wind forcing are usually weak there, mode 2 and 3 may be strongly forced over the eastern Pacific.

We test this in terms of a forced numerical experiment. We use a simple ocean model similar to the one of Yeh et al. [2001] and Moon et al. [2004]. The atmospheric wind stress are used from SODA during 1950 – 1997. Two experiments (runRef and runPx) are performed by forced wind stress to show the impact of the zonal change in projection coefficient. Experiments runRef and runPx are the same each other except the runPx uses the zonal varied projection coefficients shown in Figure 2.13 (Table 2.3). Although the ocean model forced by the same observed wind stress, the rms of zonal current anomalies from each simulation are different form each other (Figure 2.14 and Figure 2.15). The third baroclinic mode has a largest variance in eastern Pacific in runPx but runRef shows the maximum values in western Pacific.
Figure 2.12: Vertical profiles of $A_n$ from numerical calculation (no marker) and WKB approximation (squared marker). Solid, dashed and dotted lines represents first, second and third baroclinic mode, respectively. Only above 1500 m distributions of vertical structure are shown.
Figure 2.13: The mean projection coefficient (see text for definition) for the period of 1950 – 1997 along the equator, which are derived from SODA vertical mode decomposition. Solid, dashed and dotted lines are for first, second and third baroclinic mode, respectively.
**Figure 2.14:** The rms distributions from forced numerical model experiments. (a) is for total zonal current anomalies. (b), (c) and (d) are for the first, second and third baroclinic mode contribution to zonal current anomalies, respectively. The forced wind stress are from the SODA during 1950 – 1997, but the rms are from the results of 1960 – 1997. The projection coefficients used in simulation of first, second and third modes are 0.655, 0.435 and 0.14, respectively.
Figure 2.15: The same as Figure 2.14 except the projection coefficients are from Figure 2.13.
Table 2.3: Parameter values for runRef and runPx.

<table>
<thead>
<tr>
<th></th>
<th>runRef</th>
<th>runPx</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n=1</td>
<td>n=2</td>
</tr>
<tr>
<td>Projection coefficient:</td>
<td>$P_n = \frac{150}{\int_H^0 A_n^2(z)dz}$</td>
<td>0.655</td>
</tr>
</tbody>
</table>

Note also the projection coefficient of third baroclinic mode larger over the eastern Pacific. All these things make it clear that the different spatial distribution of baroclinic modes is resulted from the difference in projection coefficients, eventually from the zonal gradient of density field. We will describe this with theoretical point of view in next section.

2.6 Idealized profile works

We demonstrated that the first three baroclinic mode present different spatial distribution in equatorial Pacific in previous sections. Mode 1 variability is the largest in the western Pacific, whereas mode 2 and mode 3 is more active in the central and eastern Pacific, respectively. This is because of the zonal gradient of the density field showing the strong stratification of eastern Pacific. Considering this relatively strong stratification in eastern Pacific than western Pacific (Figure 2.10), we would like to predict the increase of stratification results in more exciting the higher mode.
Table 2.4: Values for idealized buoyancy profiles for Figure 2.16a and corresponding phase speeds \((n = 1, 2)\).

<table>
<thead>
<tr>
<th>(N_1^2)</th>
<th>(N_2^2)</th>
<th>(N_3^2)</th>
<th>(z_1)</th>
<th>(z_2)</th>
<th>(c_1)</th>
<th>(c_2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1 \times 10^{-5})</td>
<td>(5 \times 10^{-4})</td>
<td>(1 \times 10^{-5})</td>
<td>50</td>
<td>150</td>
<td>3.59</td>
<td>2.11</td>
</tr>
<tr>
<td>(1 \times 10^{-5})</td>
<td>(3 \times 10^{-4})</td>
<td>(1 \times 10^{-5})</td>
<td>50</td>
<td>150</td>
<td>3.56</td>
<td>1.91</td>
</tr>
<tr>
<td>(1 \times 10^{-5})</td>
<td>(1 \times 10^{-4})</td>
<td>(1 \times 10^{-5})</td>
<td>50</td>
<td>150</td>
<td>3.54</td>
<td>1.83</td>
</tr>
</tbody>
</table>

Units for buoyancy frequency, \(z\) and speed are \(s^{-2}\), \(m\) and \(m\ \text{s}^{-1}\), respectively.

Therefore, we should answer the question, *why the higher modes are well forced or excited when the vertical stratification increases?* In this section, we address this question from an analysis of idealized buoyancy frequency.

We now illustrated the sensitivity of the vertical structure to idealized profiles of buoyancy frequency. \(N^2(z)\) profile assumed by

\[
N^2(z) = \begin{cases} 
N_1^2, & z_1 < z \\
N_2^2, & z_2 \leq z \leq z_1 \\
N_3^2, & z < z_2 
\end{cases}
\]  

(2.12)

This piecewise form of the buoyancy profile is used by Gent and Luyten [1985] to address the vertical propagation of surface energy flux. All profiles are piecewise that is constant value. The total depth of ocean is set as 3500 m. First, the thermocline thickness is kept constant at 100 m, but its strength varies with maximum \(N^2\) values of 5, 3 and \(1 \times 10^{-4}\ \text{s}^{-1}\) (Figure 2.16a). The related parameters and the first three
Figure 2.16: (a) Idealized $N^2$ profiles with thermocline peaks of 5 (solid), 3 (dashed) and 1 (dotted) ($\times 10^{-4}$ s$^{-1}$), thickness 100 m and down to 3500 m. (b) Corresponding vertical structures of horizontal velocity from (a) idealized $N^2$ profiles. Left-three lines are for the first baroclinic modes and right-three lines are for the second modes.
Table 2.5: The same as Table 2.4 except for Figure 2.17a.

<table>
<thead>
<tr>
<th>$N_1^2$</th>
<th>$N_2^2$</th>
<th>$N_3^2$</th>
<th>$z_1$</th>
<th>$z_2$</th>
<th>$c_1$</th>
<th>$c_2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$1 \times 10^{-5}$</td>
<td>$3 \times 10^{-4}$</td>
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<td>50</td>
<td>200</td>
<td>3.61</td>
<td>2.18</td>
</tr>
<tr>
<td>$1 \times 10^{-5}$</td>
<td>$3 \times 10^{-4}$</td>
<td>$1 \times 10^{-5}$</td>
<td>50</td>
<td>150</td>
<td>3.56</td>
<td>1.91</td>
</tr>
<tr>
<td>$1 \times 10^{-5}$</td>
<td>$3 \times 10^{-4}$</td>
<td>$1 \times 10^{-5}$</td>
<td>50</td>
<td>100</td>
<td>3.54</td>
<td>1.82</td>
</tr>
</tbody>
</table>

Units for buoyancy frequency, $z$ and speed are s$^{-2}$, m and m s$^{-1}$, respectively.

Phase speeds from decomposition are given in Table 2.4. Figure 2.16b shows that as the stratification increases, the second vertical structures change shape so that the variability in the upper layers project more onto the second mode. Whereas, the first mode shows a little change. As we discussed, $A_n(0)$ determines the amount of wind stress acting on each baroclinic mode. Note also that the phase speed is increase as the stratification increase and the change is more prominent in the mode 2. This is resulted from the change of the vertical integrated buoyancy frequency that is predicted by (B.15).

We can address above results more intuitively, an increased stratification in equatorial ocean tend to trap more energy above the thermocline, which has to have much contribution on the higher-order modes, in particular through the equatorial Kelvin waves (Remember that the higher-order modes have small vertical and meridional scales).

Similarly, we could predict the effect of thickness change of buoyancy frequency on vertical mode. If the thickness of thermocline are large, then the downward
Figure 2.17:  (a) Idealized $N^2$ profiles with thermocline peaks $3 \times 10^{-4} \text{ s}^{-1}$, thickness 150 m (solid), 100 m (dashed) and 50 m (dotted) and down to 3500 m.  (b) Corresponding vertical structures of horizontal velocity from idealized $N^2$ profiles. Left-three lines are for the first baroclinic modes and right-three lines are for the second modes.
energy flux could not penetrated though it so that the higher-order modes are more excited. Figure 2.17 and Table 2.5 illustrated this. In Figure 2.17b, the thicker the thermocline, the second mode is more excited.

It is noteworthy that Gent and Luyten [1985] demonstrated the similar results from a theoretical point of view. They calculated the percentage of the downward surface energy flux that reaches the deep equatorial oceans which vary widely depending on the $N(z)$. As shown in our study, the thicker and stronger thermocline do not allow the penetration of downward energy flux through it.

## 2.7 Decadal change in vertical structure

The averaged features are addressed in the previous sections. The equatorial Pacific density field presents a rich structure with temporal variations in thermocline depth [Meyers, 1979]. We present the time-vertical distribution of buoyancy frequency in Figure 2.18. $N^2(z)$ has a variability of not only interannual but also decadal time scales. In particular the buoyancy frequency profiles reflect the persistent warming of upper ocean after late 1970s [Zhang et al., 1997] associated with increasing stratification. This subsurface variation of temperature is supported by Gu and Philander [1997] and Zhang et al. [1998], which suggest that the changes in the characteristics of the El Niño events on decadal timescales [Wang and Wang, 1996] may be due to the influx of thermocline water with an anomalous temperature from higher latitudes.

But it is rather difficult to figure out that what is cause and what is result? In
Figure 2.18: Vertical distribution of the squared buoyancy frequency as a function of time in the first 500 m at 180°, 0°. Units are $10^{-4}\text{s}^{-2}$. Contour interval is $1 \times 10^{-4}\text{s}^{-2}$. 
other words, there is a possibility that the changes in El Niño, resulted from other forcing, may cause the subsurface temperature change as shown in Figure 2.18. Some relevant studies are in order. Recent observational studies indicated preferred pathways for subsurface temperature anomalies to propagate from the off-equatorial region to the equator, which has the potential to modify the equatorial structure toward a increase of stratification. The more convincing study of this kind is due to Luo and Yamagata [2001] which showed a source for the decadal variability of the thermocline in the South western and central Pacific around 12°S. Note also that sensitivity experiments to the mean thermocline depth fluctuations with the simple coupled mode indicate that the larger the period of the mean thermocline depth fluctuation, the larger the impact on the ENSO mode [Dewitte, 2000], which is consistent with the oceanic tunnel hypothesis (considering the long time scales under consideration).

As point by previous sections, we should expect that the contribution of higher mode has been increased after pacific climate shift corresponding to increase of stratification. Thus, the mechanism proposed in Dewitte [2000] and Yeh et al. [2001] can be a reliable candidate to modulation of ENSO. We will strongly discuss this problems in the following.

2.8 Discussion

In this chapter we represented the mean characteristics of the equatorial baroclinic modes from results of vertical mode decomposition by using SODA data. The modal
distribution of surface zonal current anomalies are different from each other. The first baroclinic mode is more energetic in western Pacific, whereas the second and third mode have larger energy in central-eastern equatorial Pacific. Because the thermocline slope has a strong zonal gradient with stronger stability in eastern Pacific, the wind stress forcing are projected onto the each baroclinic mode in different way and eventually leads to the above modal distribution. Simple forced ocean model experiments indicated when atmospheric forcing projected onto baroclinic mode in zonal different manner, the rms of baroclinic zonal current anomalies show the similar distribution to observation.

There are two parameters that determine the relative contribution of the baroclinic mode: (i) the strength of the stratification and (ii) the thermocline depth. Analysis of idealized buoyancy frequency profiles supported that the stronger stratification and thicker thermocline lead to increased role of higher modes. The wind stress forcing projected onto ocean mode is depends on the surface values of the horizontal structure function, \( A_n(0) \). This simple parameter is powerful to diagnose the vertical state of equatorial ocean. In other words, long term values of \( A_n(0) \) for each mode can give us answers of the questions, “What is the energetic mode excited by wind stress at this time?” or “How the contribution of each baroclinic mode is change with time?” Figure 2.18 indicates that the role of baroclinic modes on the interannual variability has been changed with time through the changes in vertical stratification. We discuss this problem in following chapters.
Chapter 3

The baroclinic mode contributions to evolution of ENSO

3.1 Introduction

There are increasing evidences that the vertical baroclinic mode of the equatorial variability cannot be reduced to the first baroclinic mode [Kessler and McCreary, 1993; Dewitte and Reverdin, 2000; Shu and Clarke, 2002]. Dewitte et al. [1999] demonstrated that the zonal current anomaly in the eastern Pacific is prominently associated with the second baroclinic mode whereas the first baroclinic mode is most energetic in the western Pacific. From a forced simple model simulation showed that the amplitude of sea level anomalies are better captured when the higher modes are
included in model [Delcroix et al., 2000; Dewitte et al., 2002]. The first baroclinic mode simulation has only 22 cm during 1997 – 1998 El Niño, while the sea level anomaly peaks at 30 cm in this time by observation.

In a tropical Pacific coupled system, Dewitte [2000] and Yeh et al. [2001] illustrated that the baroclinic modal energy distribution could determine the timescale of the coupled variability of equatorial pacific. In their model, the relative contribution of the baroclinic modes controls the characteristics of the evolution of SST anomalies and therefore the duration and amplitude of the El Nino. Recently, Dewitte et al. [2003] derived the Kelvin and first-meridional Rossby waves for the first two baroclinic mode in the OGCM simulation. He illustrated that Kelvin waves of both mode constructively contribute to the initiation of 1997 El Nino. But the two modes have some different roles. The mature phase of the event is concurrent with an increased contribution of the second mode which becomes dominant over the first baroclinic mode.

We follow the Dewitte et al. [2003] but rather different analysis is performed in this section. The ENSO events are categorized into two types in terms of the duration, i.e., relative shorter events and longer ones [Tomita and Yasunari, 1993]. Recently, Boo et al. [2004] demonstrated that these two types of ENSO have major difference in the intensity of the low-level circulation over the western north Pacific. The shorter events are driven by the strengthened 2 – 3 year quasi-biennial signal, but the 3 – 6 interannual signal is almost the same between the two event types.

In this chapter, we will analyze the results of the 1997–1998 and 1986–1988 El Niño in terms of vertical baroclinic mode. The former was a short and strong event;
the latter was a weak and long event. Thereafter, we will identify the contributions of the vertical baroclinic modes to evolution of El Niño and suggest the relationship between the duration of El Niño and vertical mode.

The chapter is organized as follows. Section 3.2 is devoted to describe the method to decompose the variable into meridional waves. Section 3.3 presents the roles of vertical modes in terms of equatorial waves for 1997 – 1998 El Niño. Then the relationship between the vertical modes and the duration of El Niño is presented in section 3.4. Section 3.5 is devoted to concluding remarks.

### 3.2 Method: Meridional decomposition

The method of *Boulanger and Menkes* [1995] is used to project the sea level into the modes of long equatorial wave as follows:

\[ h(x, y, t) = \sum_{n=0}^{N} r_n(x, t) R_h^n(y) \]  

where \( r_n \) are the coefficients to be calculated by the method and \( R_h^n(y) \) are long equatorial wave sea level structures (\( n = 0 \) refers to Kelvin wave, \( n \) greater than 0 denotes to the corresponding equatorial long Rossby waves). The sum is assumed to be finite (\( N = 20 \)); the reader is referred to *Boulanger and Menkes* [1995] for a discussion on the value of \( N \). The decomposition of zonal current is expressed as similar way.

\[ u(x, y, t) = \sum_{n=0}^{N} r'_n(x, t) R_u^n(y) \]  

The sea level \( R_h^n(y) \) and and zonal current \( R_u^n(y) \) components are in geostrophic equilibrium and can be written as combinations of Hermite functions as follows: let
\( \psi_n \) be the normalized Hermite functions,

\[
R_0 = \frac{1}{\sqrt{2}} \begin{pmatrix} \psi_0 \\ \psi_0 \end{pmatrix} \tag{3.3}
\]

\[
R_n = \sqrt{\frac{n(n+1)}{2(2n+1)}} \begin{pmatrix} \frac{\psi_{n+1}}{\sqrt{n+1}} - \frac{\psi_{n-1}}{\sqrt{n}} \\ \frac{\psi_{n+1}}{\sqrt{n+1}} + \frac{\psi_{n-1}}{\sqrt{n}} \end{pmatrix} \tag{3.4}
\]

where \( R_n = (R_n^u \ R_n^h)^T \). For a given longitude and time, meridional sea level can be expressed by a set of long equatorial waves:

\[
h(y) = \sum_{0}^{N} r_n R_n^h(y) \tag{3.5}
\]

Then, the projection onto the wave \( j \) of sea level height is

\[
b_j = \int_{Y_S}^{Y_N} h(y) R_j^h(y) dy = \sum_{0}^{N} r_n \alpha_{n,j} \tag{3.6}
\]

where

\[
\alpha_{n,j} = \int_{Y_S}^{Y_N} R_n^h(y) R_j^h(y) dy \tag{3.7}
\]

The wave coefficient \( r_n \) is found by solving the linear system \( A \cdot \mathbf{r} = \mathbf{b} \), where \( A = (\alpha_{n,j}) \) is the \((N+1) \times (N+1)\) scalar product matrix and \( \mathbf{b} \) is the projection vector \( \mathbf{b} = (b_n) \). Because \( R_n^h \) does not constitute a complete set of functions in a bounded region \([Y_S, Y_N]\), the determinant of the matrix \( A \) is zero so that \( A \) cannot be inverted. Therefore, a solution have to be found by singular value decomposition [Press et al., 1992; Boulanger and Menkes, 1995].

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3.3 Equatorial waves for 1997 – 1998 El Niño

We first concentrate on 1997 – 1998 El Niño event. Boulanger and Menkes [1999] analyzed the wave sequence during entire period of 1997 – 1998 El Niño using the TOPEX/POSEIDON sea level data. They found that two theories potentially important for ENSO; the delayed action oscillator [Schopf and Suarez, 1988] and zonal advective-reflective model [Picaut et al., 1997]. Recently, Picaut et al. [2002] analyzed all the available satellite data during the 1997 – 1998 and found the relevance of the existing ENSO theories. They suggested that various process of ENSO were at work at various stages of the events. Figure 3.1 shows the SST anomalies and zonal wind stress anomalies along the equator. The development of El Nino SST anomalies was significantly influenced by zonal wind stress of western Pacific. The initiation of the warm event was driven by westerly anomalous forcing in December 1996 and March 1997. These westerly winds excited a series of equatorial downwelling Kelvin waves that propagated toward the east and stopped the equatorial upwelling and resulted in positive SST anomalies (Figure 3.2). It has been hypothesized that the 1997 El Niño needed the action of the westerly wind bursts to occur with such strength [McPhaden, 1999]. Picaut et al. [2002] showed that the western Pacific warm pool that is surface water greater than about 29°C migrated eastward with the collapse of the trade winds.

In June 1997 a third westerly wind burst was clearly detectable. In this time, the largest positive sea level anomaly were resulted from the forced downwelling Kelvin waves so that the SST anomalies were developed to a mature stage (Fig-
Figure 3.1: The longitude-time distribution of (a) SST anomalies (NOAA) and (b) zonal wind stress anomalies (FSU) along equator during January 1996 – July 1996. Units are °C and dyne/cm², respectively.
As the westerly wind forcing moved toward the central part of the basin equatorial upwelling Rossby waves were generated to the west (Figure 3.3). This upwelling Rossby waves reflected to upwelling Kelvin waves and resulted in initiation of La Niña phase in mid-1988. There are also easterly wind stress that forced the reflected upwelling Rossby Kelvin waves in the west of the basin (Figure 3.1). Note that the impact of the wind-forced equatorial upwelling Kelvin waves was gradually strengthened by the forcing of the easterly wind anomalies from early 1998.

Now, we shall concentrate on the modal contribution on equatorial waves in light of the results of the vertical mode decomposition. We first represent the time-evolution along the equator of the baroclinic mode contribution of the sea level anomalies (Figure 3.4). Only the results for the first two baroclinic modes are presented, because the third mode shows the relatively small amplitude to the first two modes (Figure E.1). Because first meridional Rossby waves yield a weak contribution to sea level variability along the equator, Figure 3.4 provide an estimation of the Kelvin wave contribution. For mode 2 the eastward propagation appears only east of the whole basin. West of the date line, the variability of mode 2 is much weaker than for mode 1. The eastern Pacific, however, both the modes shows the contributions of the same order. To clear the relative contributions of the first two modes, we represent the difference of the two mode (Figure 3.5a). We again find that the first mode was excited at the early stage while the second mode had stronger contribution at the mature stage.

It is interesting that the equatorial wind stress favored the first baroclinic mode rather than the second one in the western-central Pacific (Figure 3.5b). Yeh et al.
Figure 3.2: Time versus longitude distribution of sea level anomalies of TOPEX data. Units is cm.
Figure 3.3: Longitude-time distribution of the equatorial Kelvin waves and of the first meridional of the equatorial Rossby wave from TOPEX data. In order to follow possible wave reflections on the western and eastern boundaries the longitude of (b) is inverted and the Kelvin wave pattern repeated twice so the eastern boundary and western boundary are common to (a) and (c), respectively.
Figure 3.4: Time-longitude plots of sea level anomalies along the equator for the contribution of (a) the first baroclinic mode and (b) the second one. Units are cm and the values greater than 2 cm are shaded.
Figure 3.5: Time-longitude plots of (a) difference between the contribution of mode 1 and mode 2 in sea level anomalies along the equator when the sea level anomalies of mode 1 are positive. The positive values are shaded. Unit is cm. (b) The distribution of zonal wind stress anomalies (SODA) above 0.4 dyne/cm$^2$. 
[2001] demonstrated that the central mode of wind stress excites the second oceanic mode which drives the observations-like interannual variability in simple coupled model, whereas the atmospheric off-equatorial mode favors the gravest baroclinic mode. In this study, however, the atmospheric central mode (shown in Figure 3.5b) strongly driven the first vertical mode to be excited. This inconsistency compared to Yeh et al. [2001] indicates that the relative contribution of vertical mode on development of warm event may be exist. The 1997 – 1998 event has been considered as a shorter one [Boo et al., 2004]. The relative contribution of baroclinic mode could determine the time scale of equatorial variability [Dewitte, 2000; Yeh et al., 2001; Moon et al., 2004]. This problem is taken up in the next section.

Turning now to the characteristics of wave sequences in terms of vertical baroclinic mode. We projected the sea level and zonal current baroclinic contributions of vertical decomposition onto the the complete basis of the long wavelength meridional modes. These are computed from the zonally varying phase speed derived from the vertical mode decomposition. The projection is performed between 9.5°S and 9.5°N.

Figure 3.6 and 3.7 represent the Kelvin and first meridional Rossby contributions to sea level anomalies along the equator for the first two baroclinic modes. In early 1997, a westerly wind anomaly near the western boundary (Figure 3.1b) forces a first and second baroclinic mode downwelling Kelvin waves. These Kelvin waves drive the positive SST anomalies in the eastern Pacific (Figure 3.1a). Note that, while the first baroclinic mode Kelvin wave generally undergoes a decrease in amplitude as it propagates eastward, the second baroclinic mode Kelvin wave amplitude increases from about central Pacific. These two downwelling Kelvin waves
Figure 3.6: Longitude-time distribution of the (left) Kelvin and (middle) first meridional Rossby mode of the first baroclinic mode contributions to sea level anomalies along the equator. Rossby wave is displayed in reverse longitude and (right) Kelvin is repeated to visualize wave reflections at boundaries.
Figure 3.7: Same as Figure 3.6 but for the second baroclinic mode contribution.
reflect into downwelling Rossby waves at the eastern boundary.

According to Picaut et al. [1997] the reflected downwelling Rossby waves near the eastern boundary attribute to the anomalous westward currents so that the eastern edge of warm pool is pushed back to westward through zonal advection. In April 1998, the second baroclinic mode contribution to downwelling Kelvin wave component of sea level anomalies and more reflected into downwelling Rossby wave at eastern boundary (Figure 3.7). There is also larger zonal current anomalies of second baroclinic mode rather than one of first mode (Figure 3.8). To clear the role of the westward current on ENSO dynamics in terms of vertical mode, first meridional Rossby mode contribution to zonal current anomalies for the first two vertical baroclinic mode are represented in Figure 3.9. Note the coherency between the Rossby wave amplitude of the zonal current anomalies and the 28°C isotherm is more prominent in second baroclinic mode component during transient period to La Niña. Picaut et al. [2002] suggested that the demise of El Niño of 1997 and its change into La Niña in sprint 1998 were due to various causes in the eastern Pacific. One of the processes, among their suggestions, is the westward current generated by equatorial Rossby wave reflection [Picaut et al., 1997]. Figure 3.9 strongly indicated that not only the role of reflected Rossby wave on reverse of El Niño but also modal contribution to Rossby wave. Although the first baroclinic contribute to the westward displacement of the 28°C–isotherm at the surface, the second vertical mode has a more dominant role in initiating the reversal.
Figure 3.8: Longitude-time distribution of the (a) first baroclinic mode contributions to zonal current anomalies along the equator and of the (b) second baroclinic mode contribution. Units are cm s$^{-1}$. 
Figure 3.9: Longitude-time distribution of the (a) first meridional Rossby mode contribution to zonal current anomalies for the first baroclinic mode and of (b) the second baroclinic mode contribution (Thin solid and dashed lines). Two sea surface temperature (28, 29°C) lines are also represented as thick lines. Note that the distribution of second baroclinic mode in zonal current anomalies has more coherency to 28°C–isotherm rather than that of the first mode.
3.4 The vertical mode and the duration of El Niño

In the previous section, we suggested that the relative roles on the development and demise of 1997–1998 El Niño. Both the first two baroclinic modes contribute to the process associated with ENSO dynamics. But the two modes contribute to El Niño at different time stage. Recently, the modal redistribution of energy in equatorial Pacific controls the the time scale of interannual variability [Dewitte, 2000; Yeh et al., 2001; Moon et al., 2004]. They studies the baroclinic mode contribution to characteristics of air-sea coupled mode by using simple coupled model [Dewitte, 2000; Yeh et al., 2001] and to the decadal modulation of ENSO [Moon et al., 2004]. These studies have stimulated us to figure it out the roles of vertical mode on all of the ENSO events. We will expand the previous section and Dewitte et al. [2003] into the generalization of baroclinic role on ENSO dynamics.

In terms of the duration of El Niño, El Niño events are classified as shorter ones and longer events [Tomita and Yasunari, 1993; Boo et al., 2004]. These two studies emphasized the role of wind stress forcing over the western north Pacific on determining the duration of El Niño. During the mature phase of ENSO, the anticyclonic circulation over this region is enhanced which tend to decay El Niño and eventually shorten the duration of El Niño. Boo et al. [2004] also demonstrated that the wind forcing (they called quasi-biennial signal) has a significant correlation with the strength of dipole mode of Indian Ocean. So, it is great worthy task to analyze the baroclinic structure in terms of duration of El Niño.

We represent the SST anomalies and zonal wind stress anomalies during 1986—
Figure 3.10: Longitude-time distribution of the (a) SST anomalies and (b) zonal wind stress anomalies during January 1985–July 1988 along the equator. Units are °C and dyne/cm², respectively.
1988 El Niño event which is classified as a longer one [Tomita and Yasunari, 1993; Boo et al., 2004] in Figure 3.10. Compared to the event of 1997 – 1998 (Figure 3.1), the 1986 – 1988 event shows the relatively long during El Niño. In Jul 1986, the positive SST anomaly developed in eastern Pacific associated with westerly wind forcing over western basin. After, positive air-sea coupled process suggested by Bjerknes [1969] causes the development of El Niño event. It is interesting to find that even the westerly wind forcing was developing in mid 1985, but no El Niño occurred in this time. Even this westerly wind stress forced the downwelling Kelvin wave propagating eastward (Figure 3.11).

To find the baroclinic mode contribution on the initiation, development and revise of 1986 – 1988 El Niño, the modal distribution of sea level anomalies for the first two modes are represented in Figure 3.12\(^1\). It seems to make a big difference between 1997 – 1998 (Figure 3.4) and 1986 – 1988 case in terms of vertical mode variability (Figure 3.12). Most of all, two figures show that the duration of the warm phase is determined by the second baroclinic mode. Note that the both modes have similarity in evolution each other during the 1997 – 1998 event except the differences at transition time as described in previous section. The westerly wind stress well forced the Kelvin waves of first baroclinic mode in mid 1985 but had no significant effect on second mode. In this time, the warm event was not developed. This results are quite well consistent with results from simple coupled model [Yeh et al., 2001] that strongly argued that the oceanic second mode is important to make a oscillatory

\(^1\)We also represented the modal distribution of sea level anomalies for first three modes in Figure E.2
Figure 3.11: Time versus longitude distribution of sea level anomalies of SODA along equator during the period of January 1985–July 1988. Units is cm.
Figure 3.12: Time-longitude plots of sea level anomalies along the equator for the contribution of (a) the first baroclinic mode and (b) the second one during 1985 – 1988. Units are cm and contour interval is 1 cm. The values above than 1 cm are shaded.
nature of ENSO (compare the results from their Exp.4 and Exp.5). The warm event was initiated in July 1986 when is corresponding to increasing amplitude of sea level anomalies of the second baroclinic mode. The first mode had stopped the growing in January 1987 but the second mode had not. After that the second mode controls the evolution of SST anomalies in eastern basin (Figure 3.10).

In previous section, we showed that the central mode favored the gravest baroclinic mode in 1997 – 1998 El Niño that is looks like inconsistent with [Yeh et al., 2001]. To discuss this problem, we show the Figure 3.13 which is similar to Figure 3.5. The zonal wind stress favored the second baroclinic mode in 1986 – 1988 El Niño, while the westerly wind forced the first baroclinic mode in 1997 – 1998 event. Therefore, it seems reasonable to suppose that the relative modal air-sea coupled processes are exist in two events. For example, the air-sea coupled process between the atmospheric central mode and the first oceanic mode is dominant in 1997 – 1998 event. Whereas the coupling between the central mode and the second mode is favored in 1986 – 1988 period. This result suggest that oceanic baroclinic modes play a different role on equatorial interannual variability through the air-sea coupled context. It indicates that if the higher-order mode (for example, second mode) is more excited, then the period of ENSO is increased, as suggested by related studies [Dewitte, 2000; Yeh et al., 2001; Moon et al., 2004].
Figure 3.13: Time-longitude plots of (a) the difference between the contribution of mode 1 and mode 2 to sea level anomalies along the equator where the values of mode 2 are positive and of (b) the zonal wind stress anomalies above 0.2 dyne/cm².
3.5 Concluding remarks

In this chapter, we found that the first two baroclinic modes could contribute to development of the 1997 – 1998 El Niño. At first stage, the anomalous westerly wind stress forced the first baroclinic mode through the Kelvin wave response and that initiated the warm SST anomalies in eastern Pacific. The second baroclinic mode is gradually increased in its impact on warm SST anomalies. The mature phase of the event is concurrent with an increased contribution of the second mode which leads to strongest El Niño. Therefore, we safely state that the inclusion of high-order mode leads to a better simulation of the interannual variability with intermediated model (see Figure E.3 and Figure E.4).

The transition to cold event are driven by the reflected Rossby waves at eastern boundary which is suggested by Picaut et al. [1997]. Particularly, the second baroclinic mode through the Rossby waves reflected from Kelvin waves, played an important role on controlling the position of eastern edge of warm pool and eventually driven the warm event to cold period. These results agree well with Dewitte et al. [2003] in which an OGCM simulation assimilated with satellite data is used to investigated the equatorial waves in the Pacific in 1994 – 1999.

From analysis of 1986 – 1988 El Niño and its comparing to 1997 – 1998 event, we first found some evidences in observation data that the dominant coupled mode between the ocean and the atmosphere determined the duration of El Niño. When the second (first) baroclinic mode are well coupled to central mode of atmosphere, the duration of warm event tend to be longer (shorter). It is also indicated that
the second baroclinic mode is essential to drive the interannual (3−4 yr) variability in the equatorial Pacific. For example, the first baroclinic mode alone could not generated the warm event, but the second mode is responsible for growing the warm event (see Figure 3.12 at mid 1985).

Our study supports the results of Yeh et al. [2001] from sensitivity experiments of simple coupled model. In their model, the central mode excites the second oceanic mode and drives the observations-like interannual variability in SST anomalies. When the gravest baroclinic mode is dominantly excited by the atmospheric off-equatorial mode, the SST anomalies show the shortened oscillation. But in our study, we could not find the distinct atmospheric mode that favors the specific baroclinic mode. In other words, we did not show what distribution of atmospheric forcing is associated with the oceanic mode. This is currently under investigation.
Chapter 4

Vertical structure variability in the equatorial Pacific before and after the Pacific climate shift of the 1970s

A part of the present chapter has already been reported in Moon et al. [2004].

4.1 Introduction

The causes of ENSO decadal variability (EDV) are currently not known. In one view, the EDV is generated at midlatitudes where the North Pacific decadal variability is subsequently communicated to the tropics by atmospheric or oceanic teleconnections.
[Zhang et al., 1998]. In another, it results from tropical processes alone [Knutson and Manabe, 1998]. Without fully excluding the influence from the extra-tropics, the latter hypothesis is seductive considering that there are increasing evidences that the vertical structure of the low frequency variability found in the tropical Pacific Ocean cannot be reduced to a single mode [Kessler and McCreary, 1993; Shu and Clarke, 2002]. The higher-order baroclinic mode can set up a slower variability through the complex coupled instabilities [Dewitte, 2000; Yeh et al., 2001]. Although the characteristics of the vertical modes vary in the tropical Pacific with time [cf. Dewitte et al., 1999], the magnitude and pattern of these variations at interannual and decadal timescales remain unknown (We already raised this issue in section 2.7).

In this study, we use the results from the Simple Ocean Data Assimilation (SODA) system [Carton et al., 2000a] to investigate the change in the baroclinic mode contribution during the period of 1950 – 1997. A special focus is on the climate shift in the tropical Pacific Ocean around the mid-1970s. In particular, the dominant ENSO oscillation period increased from 2 – 3 years during 1960 – 1975 to 4 – 6 years after the late 1970s [Wang and Wang, 1996]. The amplitude of ENSO also tends to increase. Our motivation is to investigate if such changes can be associated with the variability of the vertical baroclinic modes before and after the late 1970s. The paper is organized as follows. Data is outlined in the next section. Section 4.3 describes changes of vertical stratification associated with those of the contribution of the first three baroclinic modes. In section 4.4, we examine the impact of the observed changes in the relative contribution of the baroclinic modes based on a simple coupled model. Section 4.5 provides concluding remarks.
4.2 Data

Monthly upper ocean data, which includes upper-ocean temperature, salinity and currents, for the period of 1950–1997 were obtained from the SODA system\(^1\). SODA uses an ocean model based on Geophysical Fluid Dynamics Laboratory MOM2 physics [Pacanowski et al., 1993] with a resolution of $2.5^\circ \times 0.5^\circ$ in the Tropics, expanding to a uniform $2.5^\circ \times 1.5^\circ$ resolution from 60$^\circ$S to 60$^\circ$S. The vertical mesh has a variable resolution with a maximum of 20 levels. The vertical resolution is a $15 - m$ at the surface down to $150 m$ with increasing resolution for the deepest level reaching 3600 m.

4.3 Analysis and results

We first show the vertical profile of mean buoyancy frequency $N^2(z)$ for the central Pacific (0$^\circ$N, 160$^\circ$E – 220$^\circ$E) before and after the late 1970s (Figure 4.1a). Here, is the Brunt-Väisälä frequency. The vertical stratification substantially increases after the late 1970s at upper levels ($\sim 150 m$) indicating a stable oceanic condition and a suppressed vertical motion. In order to estimate the contribution of vertical structural changes in the temperature and salinity responsible for an increase of $N^2(z)$, we calculate $N^2(z)$ using a climatological (1960 – 1997) temperature and salinity, respectively. Figure 4.1b (c) is the same as in Figure 4.1a except for the climatological salinity (temperature). The change of vertical salinity structure before and after

\(^1\text{We already described the data in chapter 2.2. But we rewrite this section to keep the content of Moon et al. [2004]}\)
Figure 4.1: The vertical profiles of the Brunt-Väisälä frequency, $N^2(z)$ for the period of $1960 - 1975$ (dashed) and $1980 - 1997$ solid (a). (b) and (c) are the same as in (a) except for the climatological ($1960 - 1997$) salinity (b) and temperature (c). (d) is the temperature difference between the mean over $1980 - 1997$ and the mean over $1960 - 1975$ as a function of longitude and depth along the equator. Shading is for positive and contour interval is $0.2^\circ C$. 
the late 1970s does not make a difference in the profiles of $N^2(z)$ (Figure 4.1c). This result clearly shows that changes in $N^2(z)$ (Figure 4.1a) are preferentially due to the change of vertical temperature structure. Indeed, the mean longitude-temperature section along the equator between the two epochs (Figure 4.1d) shows temperature structures with sign alternating in the vertical, indicative of increased vertical stratification after the late 1970s. However, we can not exclude the possibility that these results are due to the error of the relationship between the temperature and salinity field in the SODA system \cite{Carton et al., 2000a}.

A change in stratification between the two epochs is associated with fluctuations in the baroclinic modes characteristics. The contribution of the first three baroclinic modes to the zonal current anomalies before and after the late 1970s is compared in Figure 4.2. Figure 4.2 shows the root mean square variance (hereafter, rms) of surface zonal current anomalies from the contribution of the first three modes during periods of 1960–1975 ($a$–$c$) and 1980–1997 ($d$–$f$). Note that the anomaly is defined as the deviation from the mean over the entire period (1950–1997). Figure 4.2g–4.2i are the difference of rms between the two epochs.

Larger values for zonal current variance are confined within 5°N–5°S during the two epochs. For mode 1, maximum variance is centered in the western and central equatorial Pacific, whereas the contribution of the second baroclinic mode is more important in the eastern equatorial Pacific \cite{Dewitte et al., 1999}. The third mode presents variance peaks in the far eastern equatorial Pacific reaching $8 \text{ cm s}^{-1} \sim 10 \text{ cm s}^{-1}$ (Figure 4.2c and 4.2f), i.e., $\sim 70\%$ of the mode 1 variance peak along the equator.
Figure 4.2: The rms variance of surface zonal currents for the contribution (a) of the first mode, (b) the second mode and (c) the third mode for the period of 1960 – 1975. (d) – (f) is the same as in (a) – (c) except for the period of 1980 – 1997. Units are cm s$^{-1}$ and contour intervals are every 3 cm s$^{-1}$. (g) – (i) are the difference of rms variance between the two periods ((d) – (f) minus (a) – (c)). The shading indicates regions for which the rms variance increases above the 95% confidence level. Contour intervals are 1 cm s$^{-1}$. The rms variance of zonal wind stress anomalies for the period of 1950 – 1997 is displayed (2j). Units are dyn cm$^{-2}$ and shading is above 0.15 dyn cm$^{-2}$. 

After the late 1970s, the contribution for the second and the third baroclinic mode significantly increases in the central equatorial Pacific (Figure 4.2h and 4.2i) where the zonal wind stress anomaly has maximum rms (Figure 4.2j). We found similar results in terms of meridional currents and sea level pressures with significant increase in the central equatorial Pacific (Not shown\footnote{This figure is represented in Figure D.1}). It indicates that local forcing of Kelvin and Rossby waves for the higher baroclinic modes is reinforced after the late 1970s. The higher-order modes Rossby waves with a slower gravity wave speed are associated with an increase of ENSO oscillation period according to the ocean wave dynamics \cite{Philander1984}. The frequency spectrum of NINO3 (5\degreeN – 5\degreeS, 210\degreeE – 270\degreeE) SSTA for SODA also indicates that longer periods are favored after the late 1970s (see Fig. 4b). These results suggest that an increase of ENSO period oscillation after the late 1970s is associated with the variability of the higher baroclinic modes.

In order to check this hypothesis, we first analyze the projection of the wind stress onto the baroclinic mode. Based on mixed layer dynamics \cite{Cane1984}, a body force of magnitude is

\[
\tau_n(x, y, t) = \tau(x, y, 0, t) \frac{A_n(0)}{D}
\]  

where $D$ is the depth of the ocean and $A_n(z)$ is the vertical structure function. Obviously, this indicates that the wind stress acting on each baroclinic mode depends on the surface amplitude of vertical structure function, i.e., $A_n(0)$. Figure 4.3a–4.3c show the difference of rms for $A_n(0)$($n = 1, 2, 3$) between 1960 – 1975 and 1980 –
1997 along the equator. Figures 3a-c indicate that the rms of $A_n(0)$ for the second mode significantly increases in the central equatorial Pacific after the late 1970s. For the third mode an increase of the rms of $A_n(0)$ is significant in the central and eastern equatorial Pacific. It is worth to note that the effect of an increase of surface amplitude $A_n(0)$ in the central equatorial Pacific may be significant in a coupled system because there is large wind stress forcing (Figure 4.2j). This result suggests that the atmospheric forcing for the higher baroclinic modes is increased in the central and eastern equatorial Pacific after the late 1970s, namely that higher-order baroclinic modes are favored after the late 1970s.

4.4 Simple model experiments

In this section, we test the impact of changes in the energy distribution on the baroclinic modes in a simple coupled model based on the above results. We use a simple coupled model similar to the one of [Yeh et al., 2001]. The atmospheric model is a statistical model based on the singular value decomposition (SVD) of SSTAs and wind stress anomalies from SODA for the period 1950–1997. The ocean model is an extension of the [Cane and Zebiak, 1987] model including three vertical baroclinic modes. The values for phase speed ($C_n$) and projection coefficient ($P_n$) before and after the late 1970s are derived from the vertical mode decomposition of a mean density profile along the equator (Table 4.1). As expected, the averaged values of $P_n$ in the central equatorial Pacific increase for higher baroclinic modes after the late 1970s (increase of 19% and 25% for mode 2 and 3, respectively) indicating
Figure 4.3: Difference of rms variance for $A_n(0)$ along the equator of (a) the first mode, (b) the second mode and (c) the third mode between 1960 – 1975 and 1980 – 1997. Unit is nondimensional. The horizontal dashed line indicates the 95% confidence level for rms variance increase.
Table 4.1: Parameter Values for the Ocean Model

<table>
<thead>
<tr>
<th></th>
<th>1960 − 1975 (Exp1)</th>
<th>1980 − 1997 (Exp2)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>n=1</td>
<td>n=2</td>
</tr>
<tr>
<td>Phase speed (m/s) ( C_n )</td>
<td>2.69</td>
<td>1.61</td>
</tr>
<tr>
<td>Projection coefficient ( P_n )</td>
<td>( \int_0^{-H} A_n^2(z) dz )</td>
<td>0.65</td>
</tr>
</tbody>
</table>

\( P_n = \int_0^{-H} A_n^2(z) dz \)

These values are derived along the equator (0°N, 160°E − 220°E).

that atmospheric forcing projects more onto the higher-order baroclinic modes. The change for phase speed between the two epochs is however very small [cf. Dewitte, 2000].

Two simulations (Exp1 and Exp2) are performed over a 300 year period. The simplest way to show the impact of the observed changes in baroclinic mode energy distribution in the late 1970s is to specify the oceanic parameters deduced from the above analysis in the model. Thus Exp2 is identical to Exp1 except for the value of \( C_n \) and \( P_n \) taken before (Exp1) and after (Exp2) the late 1970s. Figure 4.4a shows the power spectrum of the NINO3 SST index for Exp1 (dashed) and Exp2 (solid). The dominant period of NINO3 SST index for Exp1 and Exp2 is 4.5 yr and 6 yr, respectively. Although the difference in the dominant period between the two experiments is not as large as in the observations (Figure 4.4b), the realistic magnitude of the shift in period peaks for the model suggests that the change in the relative contribution of the baroclinic modes is associated with an increase of ENSO

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oscillation period.

Figure 4.4c is the difference of rms (Exp2 minus Exp1) for simulated SSTA. We have also displayed the difference in rms between the two epochs for the SODA SSTAs (Figure 4.4d). The rms for Exp2 significantly increases in the central and eastern tropical Pacific, which is associated with the increase of ENSO amplitude after the late 1970s. The difference in variance between the two epochs is weaker in the observations than in the model, which partly reflects difference in the time span over which the statistics is done (300 yrs for the model and 15 yrs for the observations). The patterns, however, are comparable, consistent with a Kelvin wave response: the projection coefficient being amplified for the higher order modes for the model (observations) in Exp 2 (after the late 1970s), the forced Kelvin wave for the high-order modes with a slower phase speed than the gravest mode is favored in the eastern Pacific. Note also that the meridional scale of variability for the observations is rather small ∼ 200 km at 130°W indicative of a response in SSTAs to higher-order baroclinic mode Kelvin wave forcing.

In order to test the robustness of our result, in particular with regards to the definition of the atmospheric model, we repeated experiments similar to Exp1 and Exp2 except that the atmospheric modes of the statistical model are computed over the two different periods under concerns (1960 – 1975 and 1980 – 1997) from the SODA SSTAs and wind stress anomalies. Thus Exp1 and Exp2 are repeated with the new two different atmospheric models. When the relative contribution of the forced higher baroclinic modes increases, in both cases, the model also exhibits a comparable increase of the amplitude and period of ENSO oscillation (not shown).
Figure 4.4: The normalized power spectrum of the simulated NINO2 SST index for Exp1 (dashed) and Exp2 (solid) (a). (b) is the same as in (a) except for SODA SSTA during the period of 1960−1975 (dashed) and 1980−1997 (solid). (c) is the difference of rms variance for simulated SSTA (Exp2 minus Exp1). (d) is the same as in (c) except for SODA SSTA (1980−1997 minus 1960−1975). Shading indicates regions in which the rms variance increases above 95% confidence level. Contour interval is 0.5°C.
Again, it suggests that an increase in the higher-order baroclinic modes contribution in a parameter range characteristics of the Pacific climate shift of the 1970s is associated with larger amplitude and longer period of ENSO oscillation.

It is noteworthy that we found that there are differences in the atmospheric SVD modes before and after the 1970s. According to Yeh et al. [2001], these modifications have the potential to modify the characteristics of the ENSO variability. However, we hypothesize that these modifications in the atmospheric modes are a response to a change in the ocean.

4.5 Concluding remarks

We investigated the change in the baroclinic mode contributions at low frequency with a special focus on the climate shift in the tropical Pacific Ocean around the mid-1970s. Our analysis suggests that the vertical stratification substantially increases at upper levels in the tropical Pacific after the late 1970s, which is mostly due to vertical structural changes in the temperatures within the thermocline.

The change of vertical stratification is associated with an increase of the contributions for the higher baroclinic modes to the current and pressure anomalies. After the late 1970s, the contribution for the second and the third baroclinic mode significantly increases in the central equatorial Pacific, suggesting that local forcing of Kelvin and Rossby waves for the higher baroclinic modes is reinforced [Dewitte, 2000; Shu and Clarke, 2002]. The variance analysis of surface amplitude $A_n(0)$ supports this hypothesis, of more intense atmospheric forcing for the higher baroclinic
modes after the late 1970s.

At this point, we should note some results that are not provided in Yeh et al. [2001] and Moon et al. [2004]. Appendix C and D present the SST tendency terms of their simulation. We could found that the higher-order modes tend to generate warm SST anomalies through the thermocline feedback, whereas the gravest mode is strongly associated with zonal advective feedback.

Because the SODA system is based on a data assimilation system forced by winds, there is the possibility that decadal variations of wind forcing are responsible for the EDV [Karspeck and Cane, 2002; Pierce et al., 2000; Wang and An, 2001]. If this is the case, our study indicates that such decadal wind forcing signal will project preferentially onto the equatorial higher-order baroclinic mode. There is, however, no particular reason for the wind forcing at such time scales to favor one mode over another, which suggest that the changes in stratification impacting on the wave dynamics are more likely to be due to oceanic extra-tropical connections and that decadal changes in the wind stress forcing are the result of a coupled response to the ocean. Thus our study suggests that the decadal signal associated with the climate shift of the late 1970s is linked to the change in the ENSO period and magnitude resulting from changes in the tropical coupled mode characteristics, themselves associated to modification of the energy distribution on the baroclinic modes.

Coupled climate models suggest that under climate change, the stratification of the equatorial Pacific ocean increase [Timmermann et al., 1999b; Cai and Whetton, 2000]. This takes place in associated with an increase in thermocline depth,
which tends to suppress the contribution of the high-order modes. Consequently, we can not exclude the possibility that despite of increased stratification it does not always translate into increased period and amplitude of ENSO events. For example, Timmermann et al. [1999b] argued that the response of ENSO to greenhouse warming in a coupled model generally does not show increased period or amplitude despite in presence of increased stratification. This is currently being investigated from a general circulation coupled model that includes all the different processes invoked in this study.
Chapter 5

Development of a Coupled GCM

5.1 Introduction

ENSO is the most dominant interannual variability in the equatorial Pacific. It has been widely accepted that ENSO is a manifestation of a coupled atmosphere-ocean instability [Philander, 1990]. Since there are not enough observational data to fully understand it, coupled atmosphere-ocean general circulation models (GCMs) are the best tool for understanding various aspect of the ENSO mechanisms.

Many studies have been devoted to the study of ENSO mechanisms [Wyrtki, 1975, 1985; Zebiak and Cane, 1987; Schopf and Suarez, 1988; Weisberg and Wang, 1997; Picaut et al., 1997; Jin, 1997a,b; Kang and An, 1998; An and Kang, 2001]. Because these studies have been performed based on simplified models, the hypothe-
sis drawn from those must therefore be confirmed with more comprehensive climate model.

In this study, we developed a new version of CGCM to study various mechanisms related to ENSO. It is also found that the coupled GCM is a good tool to test issued process associated with oceanic vertical baroclinic mode. The coupled model is presented in section 2. Section 3 presents model climatology. Section 5 and section 6 examine interannual variability and, in more detail, the mechanisms for ENSO, respectively. Section 7 is devoted to the characteristics of vertical mode in CGCM. A summary and concluding remarks follow in last section.

5.2 CES CGCMv2

The model used for this study is the CES Coupled General Circulation Model version 2. In this section we describe the two parts of the CGCM.

5.2.1 SNUAGCM

The atmospheric model used in CGCM is the Seoul National University AGCM (SNUAGCM), and more details can be found in Kim [1999]. The SNUAGCM has been developed in Seoul National University, which is a global spectral model with T42 resolution (approximately 2.8°). There are 20 unevenly spaced sigma-coordinate vertical levels in the model.

The SNUAGCM is based on the CCSR/NIES AGCM in Tokyo University [Numaguti et al., 1995], but has several major changes including the land surface process,
shallow convection, and PBL processes. The land surface parameterization is the same as in the land surface model [Bonan, 1996] developed in the NCAR Community Climate Model 3 (CCM3). The SNUAGCM contains the non-precipitating shallow convection in diffusion type and the non-local PBL/vertical diffusion scheme [Holslag and Boville, 1993]. The radiation process is parameterized by the two stream k-distribution method [Nakajima and Tanaka, 1986]. The cumulus parameterization is based on the Relaxed Arakawa-Schubert scheme [Moorthi and Suarez, 1992]. The orographic gravity-wave drag are parameterized following McFarlane [1987].

SNUAGCM simulated well the climate fields in precipitation and low-level winds which are associated with Asian monsoon [Yeh, 2001] and Lee [2001]. This AGCM also participated in AGCM intercomparison to simulate the anomalies associated with the 1997/98 El Niño [Kang et al., 2002].

5.2.2 MOM2 and ocean mixed layer model

The OGCM is based on the Modular Ocean Model version 2.2 (MOM2.2) of the Geophysical Fluid Dynamics Laboratory (GFDL) model [Pacanowski et al., 1993]. The model domain covered the global ocean from 30°S to 50°N. To avoid boundary problem along the northernmost and southernmost artificial walls, a Newtonian damping term is added to the tracer equations. The horizontal (longitude by latitude) grid spacing is varying from 1° × 1/3° (equator) to 1° × 3° (Northern and Southern boundary). There are 37 unevenly spaced levels in vertical direction whose thickness increase with depth from 7.5 m at the surface to 1000 m at the bottom. Realistic bottom topography was used as much as possible within the limits of grid
Horizontal eddy viscosity and diffusivity were $1 \times 10^8 \text{ cm}^2 \text{s}^{-1}$ and $1 \times 10^7 \text{ cm}^2 \text{s}^{-1}$, respectively. We used a new developed ocean mixed layer model by Noh and Kim [1999] as a vertical mixing. This model is a second-order turbulence closure one using eddy diffusivity, but it produces a well-mixed layer even under the stabilizing heat flux, consistent with the observation. This model was also shown to improve various feature of the OGCM such as the mixed layer depth and the equatorial circulation [Noh et al., 2002].

The observation shows that turbulent kinetic energy (TKE) near the sea surface exhibits $\sim 2$ order larger magnitude than that expected from the wall boundary layer [Drennan et al., 1996; Terray et al., 1996]. Noh and Kim [1999] developed a new ocean mixed layer model taking these observational evidences into consideration. The model is a second-order turbulence closure model using eddy diffusivity, but it produces a well-mixed layer even under the stabilizing net flux, consistent with the observation. The well-known Mellor-Yamada model [Mellor and Yamada, 1982], however, results in excessively high SST under heating and leads to strong stratification in upper ocean [Rosati and Miyakoda, 1988]. Noh et al. [2002] tested the new ocean mixed model in an ocean general circulation model (OGCM), and they showed that SST simulation is improved by new vertical mixing process compared to constant vertical mixing and the scheme of Pacanowski and Philander [1981]. Because the OGCM forced by prescribed wind forcing and restoring heat flux, there are no air-sea coupled processes by which the model may tend to drift, so further investigation of simulation experiment with a new vertical scheme [Noh et al., 2002]
in fully coupled GCM are required.

5.2.3 Coupling procedure

The ocean and atmosphere model exchange information (air-sea fluxes and SST) once per simulated day. Model integrated without any flux correction or restoring (except in northern and southern zonal boundary). The simulation analyzed in this study was integrated for 19 years. The initial oceanic condition is an ocean at rest with climatological January distributions of temperature and salinity corresponding to the climatology compiled by Levitus [1982]. The initial atmospheric condition is obtained by integrating the AGCM forced by climatological SST for 10 years.

5.3 Mean state

We first determine how well the climatological SST are simulated. The modeled, observed annual mean SST and their difference are shown in Figure 5.1. The simulated SST fields compare well with the observed in terms of the pattern, with warm SST in western Pacific. However, the equatorial cold tongue is too strong and extends too far western Pacific, compared with the observation. This is a common error experienced by many coupled GCM [Mechoso et al., 1995]. This abnormal cold tongue may be associated with model upwelling that is too strong in the central and western equatorial Pacific.

The model bias in the simulated SST climatology may affect the precipitation distribution, which is presented in Figure 5.2. For comparison, the observed
Figure 5.1: Annual mean sea surface temperature (°C) distributions in (a) CGCMv2, (b) NOAA SST and (c) their difference.
precipitation based on the data set of Xie and Arkin [1997]. The simulated mean precipitation agrees rather well with observations. Intense precipitation is found as expected over the narrow band of atmospheric convergence region known as the Inter Tropical Convergence Zone (ITCZ), corresponding the warm SST band crossing the Pacific to the north of the equator (Figure 5.1). There are some important difference, however, between the simulation and the observations. The simulated precipitation over ITCZ is smaller rather than observation around 120° W. This may be resulted from the unrealistic SST which controls the boundary layer wind stress, consequently modifying the moisture convergence. The strong cold tongue, as discussed above, pushing the precipitation region far westward.

The subsurface ocean temperatures from CGCMv2 and Simple Ocean Assimilation Data (SODA) are shown in Figure 5.3. The zonal gradient of thermocline (as the depth of the 20°C isotherm) is well simulated compared with assimilation data. This features may be results from implementing a new OMLM of Noh et al. [2002]. However, the thermocline depth in eastern Pacific is too shallow compared with SODA and the temperature over central Pacific is strongly cool. The causes of these two biases may be associated with the problem of cold tongue.

In this time, we do not discuss these problems any more, i.e., What causes the biases in cold tongue? or How to suppress these? The fuller study of these problems lies outside the scope of this study.

The averaged zonal current from simulation and assimilation data are presented in Figure 5.4. The zonal current tilts upward from western Pacific to eastern Pacific that is consistent with a zonal pressure gradient in the upper ocean as shown
Figure 5.2: (a) Simulated and (b) observed (CMAP) annual mean precipitation, in units of mm day$^{-1}$. Shading indicates regions in which the precipitation is above 6mm day$^{-1}$.
Figure 5.3: Subsurface temperature from (a) simulation and (b) assimilation data from SODA as shown in longitude-depth distribution. Units are °C. The values between 15 °C and 24 °C are shaded.
Figure 5.4: Subsurface averaged zonal current from (a) simulation and (b) assimilation data from SODA as shown in longitude-depth distribution. The units are cm s$^{-1}$. Positive values are shaded.
in Figure 5.3. However, because of the weak pressure gradient associated with dif-fused thermocline, the strength of the undercurrent is too weak. The observed strength is on the order of 100 cm s$^{-1}$ [McPhaden et al., 1998].

### 5.4 Simulated interannual variability

Figure 5.5 shows the time evolution of simulated sea surface temperature (SST) anomalies and zonal wind stress anomalies in the equatorial band between 5$^{\circ}$S–5$^{\circ}$N. Anomalies are defined as differences from the monthly mean. Figure 5.5a shows the large SST anomalies in the eastern-to-central Pacific. The largest amplitude of SST anomalies is close to 2$^{\circ}$C. Anomalous warm SST events appeared approximately every 3 – 4 yr (at yr 6, 9, 13, 16, 19). This recurrence frequency of the simulated warm events is close to that of observed El Niño events [Rasmusson and Carpenter, 1982]. Figure 5.5b shows that most wind stress anomalies are confined to the western part of the ocean basin, with largest amplitudes closed to 0.2 dyne cm$^{-2}$. It is clear that all major warm events are accompanied by strong westerly wind stress anomalies to the west of the SST anomalies.

Figure 5.6 shows the power spectra for the time series of the simulated and observed Niño3 index. The model has its largest peak on a timescale of about 40 months, where as the observation shows the broadened (35 – 70 months) spectral energy. This difference may be due to not only short range of simulated data but also the intrinsic features of this model. We will investigate the characteristics of baroclinic mode in following section and suggest the baroclinic modal contribution
Figure 5.5: Time-longitude distribution equatorial (5°S – 5°N) (a) sea surface temperature anomalies and (b) zonal wind stress anomalies from simulation. Units are °C and dyne cm⁻², respectively. The values whose absolute values are greater than 0.5 °C for SST anomalies and are greater than 0.1 dyne cm⁻² for zonal wind stress anomalies are shaded.
Figure 5.6: Power spectrum of the (a) model and (b) the observed (NOAA) Niño3 SST anomalies against red noise (dashed line). We used observational SST anomalies during the period of 1950 – 2000.
to interannual variation.

The model also well showed the biennial oscillation and near-annual mode [Jin et al., 2003; Kang et al., 2004] oscillation. But the model has larger energy over these fast oscillation mode. The causes for this larger energy in model than the observed frequency need to be investigated further.

Figure 5.7 shows the model spatial pattern of the rms variability for modeled and observed SST anomalies. The simulated anomalies appeared to be more westward toward western Pacific, which is results from the westward shift of the cold tongue.

5.5 ENSO mechanism in CGCMv2

The recharge oscillator model of ENSO is a simple conceptual model characterized by recharge-discharge in equatorial heat contents [Jin, 1997a,b]. The oscillatory nature of ENSO in this scenario results from the disequilibrium between equatorial zonal winds and zonal mean thermocline depth. The former is in phase with the eastern Pacific SST and also in zonal thermocline tile. The conceptual model can be easily tested by examining the equatorial mass or heat content budgets without the details of forced and free equatorial waves and their reflections.

A critical analysis of the oscillatory mechanism suggest by recharge model requires comprehensive and basinwide information of both surface and subsurface ocean states, but that is currently difficult from observation. The data provided by CGCM is a most useful alternative. In this section, we use the simulated results
Figure 5.7: The standard deviation of the SST anomalies from (a) simulation (15 years) and (b) observation (1982 – 2002). Units are °C. Contour interval is 0.2°C. The minimum value of contour line is 0.3°C.
to testing the relationships hypothesized by recharge oscillator. We first low-pass filtered the simulated data to isolate $3 - 4$ year time scale.

Figure 5.8 represents the equatorial ($5^\circ S - 5^\circ N$) thermocline depth ($Z_{20}$) anomalies, zonal eddy part of thermocline depth and zonal wind stress anomalies ($3^\circ S - 3^\circ N$) and SST anomalies. There are the buildup of heat content in western Pacific before warm SST event in eastern Pacific. Note that SST and zonal wind stress anomalies present the westward propagation. This moving nature of anomalies will be discussed in next section. The cold SST anomalies of eastern Pacific is also followed by negative thermocline depth anomalies in western Pacific. The zonal wind stress anomalies produce eastern-western tilted thermocline depth anomalies in equator.

One of the critical elements of the recharge oscillator model is the slow adjustment of the zonally averaged thermocline depth along the equator. Thus there is phase difference (theoretically, 90 degrees) between the variations of the zonally uniform thermocline depth along the equator and the SST anomalies in the eastern Pacific. Figure 5.9 shows well this relationship.

Figure 5.10 depicts the evolution of ocean heat contents in the tropical Pacific during the oscillation in Figure 5.9. There are recharge in equatorial pacific (Figure 5.10a) during pre-onset. A fast spread from the western Pacific to the eastern Pacific along the equator during the onset (warm SST anomalies in eastern Pacific). When the SST anomalies of eastern Pacific decay to normal condition, the heat contents of equatorial Pacific are discharged to off-equator (Figure 5.10c).
Figure 5.8: Time-longitude distribution equatorial (a) thermocline depth anomalies (shading) and its zonally asymmetric component (contour). (b) is the zonal wind stress anomalies (shading) and SST anomalies (contour). Note that this distribution is from the low-pass filtered data.
Figure 5.9: The time series for the zonal mean thermocline depth anomalies (m, dashed line) and SST anomalies (°C, solid) over NINO3.
Figure 5.10: The composite maps for thermocline depth anomalies. (a) represents when $[Z_{20}]^{15^\circ N}$ is at positive peak, (b) is when transition time from positive to negative value, and (c) is at negative peak. Contour intervals are 4 m. The dashed lines represent negative values.
5.6 The vertical baroclinic modes in CES CGCMv2

In this section we discuss the vertical baroclinic modes of equatorial variability in coupled GCM. Although the similar analysis are performed in OGCM [Dewitte et al., 1999], little is known about the modal variability in coupled GCM.

Figure 5.11 represents the the rms variability of surface zonal current anomalies over the 15-yr period (05yr − 19yr). As shown in Figure 2.4, the modal distributions are rather different from each mode. Mode 1 has a peak value in western pacific and mode 3 presents maxima of eastern Pacific. The contribution of the second baroclinic mode is over emphasized in whole tropical ocean relative to one of Figure 2.4b. We already discussed that the each baroclinic contribution of surface zonal current anomalies is determined by the relatively projected amount of wind stress. Thus it seems reasonable to suppose that there are remarkably increased in the contribution of second baroclinic mode to simulated surface zonal current anomalies.

To conform this, the surface values of the horizontal structure function \(A_n(0)\) are represented in Figure 5.12. Based on mixed layer dynamics [Cane, 1984], the projecting force onto each baroclinic mode is proportional to \(A_n(0)\). The \(A_n(0)\) distributions of CGCM are similar to ones of assimilation data. The second mode of CGCM, however, has much larger values than SODA data. The larger values of \(A_n(0)\) in second mode indicates that there are dominant contribution of second mode to equatorial interannual variability in CGCM. There are also increase in mode 3 of CGCM with compared to SODA. These results indicate that higher baroclinic mode
Figure 5.11: The rms of baroclinic zonal current anomalies for (a) first mode, (b) second mode and (c) third mode. Units are cm s$^{-1}$. The contour intervals are all 0.3 cm s$^{-1}$. 
Figure 5.12: \( A_n(0) \) distributions along equator for the first three baroclinic modes from (a) CGCM and (b) SODA. Red lines are for mode 1, green lines and blue lines are represents mode 2 and mode 3, respectively. \( A_n(0) \) has a non-dimensionalized unit.
are more energetic in CGCM than those in SODA.

Then, why are the higher mode favored in CGCM? The point to observe is that CGCM has a too diffused thermocline (Figure 5.3). Theoretical consideration suggested that the thicker the thermocline, the less downward energy flux penetrates through it [Gent and Luyten, 1985]. We could say that the diffused thermocline tend to trap more energy above the thermocline along the equator, which has to have some contribution on the higher-order modes (in particular second mode). We should note that this diffused thermocline’s role, tend to confine energy in upper ocean, is consistent with that of strengthened thermocline (see chapter 2). The vertical distribution of Brunt-Väisälä frequency also indicates this (Figure 5.14). We can also test the diffused thermocline in term of vertical decomposition by figuring out the variation of $A_n(0)$. Figure 5.15 presents the results that the strong stratification in upper ocean tend to increase the contribution of high-order modes, especially the second mode.

Again, Figure 5.13 shows the dominant second mode that determined the interannual variability of SST anomalies in coupled GCM by comparing with Figure 5.5. In 19 model year, the strong El Niño-like warm event is developed by contribution of the first and second baroclinic mode.

5.7 ENSO sensitivity to altered climate

In this section we shall focus on the simulated ENSO sensitivity to the tropical climatology in CGCM. In our coupled GCM, the SST anomalies of equatorial Pacific
Figure 5.13: The time-longitude plots of baroclinic mode contribution to sea level anomalies for (a) mode 1, (b) mode 2 and (c) mode 3 from simulation of CES CGCMv2. Units are cm.
**Figure 5.14:** The vertical distribution of buoyancy frequency from SODA (solid) and CGCM (dashed) at equatorial Pacific (140°, 0). The period of SODA and CGCM are 1950 – 1997 and 05 – 19 year, respectively. Note that the CGCM has strong stratification in upper ocean.
Figure 5.15: (a) Idealized $N^2$ profiles to simulate the buoyancy of Figure 5.14. Solid line represents the case of being strong stratification in upper ocean and dotted line shows the low stratification in upper layer. The solid line is stand for the CGCM. (b) Corresponding vertical structures of horizontal velocity from (a) idealized $N^2$ profiles. Left-three lines are for the first baroclinic modes and right-three lines are for the second modes. The third baroclinic modes are presented in centered-three lines.
favor westward propagation (Figure 5.8). The zonal propagation of SST anomalies is resulted from a competition between thermocline feedbacks and zonal advective feedbacks [Hirst, 1986, 1988; Hao et al., 1993; Jin and Neelin, 1993; An and Jin, 2000, 2001; An and Kang, 2001; Kang et al., 2001]. These two feedbacks, which are linked dynamically through the geostrophic relationship, tend constructively to lead to the growth and phase change of ENSO. Thermocline feedbacks are associated with the equatorial thermocline slop, which responds fairly rapidly to changes in zonal wind stress. A westerly zonal wind forcing induces a deepening of the thermocline in the east so that the SST anomalies propagate eastward associated with warming in the east side of anomalies through mean upwelling. The zonal advective feedbacks are associated with the anomalous zonal currents and zonal mean temperature gradient, which respond locally to stress anomalies. A warm SST anomaly induces westerlies to the west, which leads to the warm advection that move the SST anomaly westward. Therefore, thermocline and zonal advective feedbacks tend to favor eastward and westward propagation of SST anomalies, respectively.

We will confirm that the westward propagation of SST anomalies in our CGCM is due to zonal advection feedback, using a simple two-strip model of An and Jin [2001]. This model includes two-strip approximation to equatorial ocean dynamics and one-strip approximation to the SST equations. As shown in Jin [1997b], the ocean dynamics equations for $h_e$ in the equation strip ($y = 0$) and $h_n$ in the off-equator strip centered at $y_n$ can be approximately shown as

$$
(\partial_t + \varepsilon_m)(h_e - h_n) + \partial_x h_e = \tau_{xe}
$$

(5.1)
\[(\partial_t + \varepsilon_m)h_n - \partial_x h_n/y_n^2 = \partial_y(\tau_x/y)|_{y=y_n} \quad (5.2)\]

where \(\tau_x\) is the wind stress anomaly evaluated in the equatorial strip. The boundary condition for the two-strip model are written as

\[h_n(x_E, t) = r_E h_e(x_E, t), \quad h_e(x_W, t) = r_W h_n(x_W, t) \quad (5.3)\]

where \(r_W\) and \(r_E\) are reflection parameters depending on the boundary conditions.

The equation SST anomaly equation linearized about an upwelling climate state and the zonal gradient of climatological mean SST can be written as

\[\partial_t T_e = -c(x)T_e + \gamma(x)h_e + a(x)u_m \quad (5.4)\]

where \(\gamma(x) = \gamma_0(x)\bar{w}_1/H_{1.5}\) and \(a(x) = -\partial T(x)/\partial x\). \(T_e\) represents SST anomalies along the equator strip. We used all parameters from ideal case of An and Jin [2001], but except \(a(x)\). Table 5.1 shows the values used in this study. Note that the \(T_{eE}\) is the SST anomaly averaged over the eastern half of the equation strip and \(f(x)\) is a normalized function whose zonal integration is unity. The finite difference method is applied to \(x\)-dependence of the variables. The spatial scale of the ocean basin chosen is 16000 km and the ocean basin is divided into equally spaced 50 grids.

There is a strong cold tongue in CGCM that centered far westward (Figure 5.1). So the two basic states of \(a(x)\) in CGCM and simple ocean assimilation data (SODA) are significantly different from each other, particularly over western Pacific (Figure 5.16).

Figure 5.17 shows the SST anomalies which are calculated from the eigenvectors of the first leading mode associated with simple recharge model. In each case,
Table 5.1: Values for the two-strip model.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_{xe}$</td>
<td>$\mu b_0 T_E f(x)$</td>
</tr>
<tr>
<td>$\partial_y(\tau_{x/y})</td>
<td>_{y=y_n}$</td>
</tr>
<tr>
<td>$\mu$</td>
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</tr>
<tr>
<td>$b_0$</td>
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</tr>
<tr>
<td>$\theta$</td>
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</tr>
<tr>
<td>$y_n$</td>
<td>2</td>
</tr>
<tr>
<td>$r_W$</td>
<td>0.65</td>
</tr>
<tr>
<td>$r_E$</td>
<td>0.75</td>
</tr>
<tr>
<td>$e_m$</td>
<td>1/15</td>
</tr>
<tr>
<td>$c(x)$</td>
<td>1</td>
</tr>
<tr>
<td>$\gamma(x)$</td>
<td>$\begin{cases} 0, &amp; 0 \leq x &lt; 1/2 \ 1, &amp; 1/2 \leq x \leq 1 \end{cases}$</td>
</tr>
<tr>
<td>$f(x)$</td>
<td>$\begin{cases} 1/(x_2 - x_1), &amp; x_1 \leq x \leq x_2 \ 0, &amp; x_2 &lt; x \text{ or } x &lt; x_1 \end{cases}$</td>
</tr>
<tr>
<td>$x_1$</td>
<td>1/4</td>
</tr>
<tr>
<td>$x_2$</td>
<td>3/4</td>
</tr>
<tr>
<td>$c(x)$</td>
<td>1</td>
</tr>
</tbody>
</table>
**Figure 5.16:** Longitudinal distribution of the coefficient \( a(x) \) in the SST equation (5.4). The solid line indicates the values from CGCM and dashed line is for SODA (1980 – 1997).
there is a oscillation in eastern Pacific resembling the general feature of the observed ENSO. However, the CGCM case (Figure 5.17a) shows the strong westward propagation signal while the SODA case (Figure 5.17b) has the weak signal. This figure indicates that the west propagating nature of SST anomalies in CGCM broke out as a results of the dominant zonal advection feedback.

5.8 Conclusion

An atmospheric general circulation model (AGCM) was coupled an ocean GCM without the use of flux correction. Recently developed ocean mixed layer model is also implemented [Noh and Kim, 1999]. This coupled model (CES CGCMv2) simulates well the mean fields (temperature, currents, and precipitation) over Pacific. In comparison with other CGCMs, our model shows remarkable performance both in thermocline and equatorial undercurrent 3. But the equatorial cold tongue tends to be too strong and extend too far west. The central Pacific has a too diffused thermocline.

The CGCM is shown to produce ENSO-like interannual variability with reasonable frequency and amplitude. The analysis of ocean heat content shows that the simulated interannual cycle is associated with charge-recharge process [Jin, 1997a].

It is argued that the zonal gradient mean temperature in equatorial Pacific is responsible for the westward propagation of SST anomalies in the model. The simple

\footnote{The figures of intercomparison are not shown in thesis. Contact to prodigy latent@yahoo.co.kr for the results. Recently, Dr. Kug has performed simulation with a global version of this CGCM for 40-year and found the similar results as described in this thesis.}
Figure 5.17: Time-longitude plots of SST anomalies with different $a(x)$. (a) is derived from CGCM and (b) is derived from SODA (1980 – 1997). All other values are same as shown in Table 5.1. Units are $°C$. 
model supports this by sensitivity test. The simulated equatorial zonal temperature
gradient enhanced the zonal advective feedback so that the SST anomalies showed
the westward propagation.

Analysis for vertical baroclinic mode suggested that the second mode is strongly
favored in equatorial central Pacific. This is associated with the too diffused ther-
mocline. Because the diffused thermocline tend to trap the energy to the upper level
of ocean, the higher-order modes are well excited as shown in chapter 2.
Chapter 6

Summary and discussion

The characteristics of vertical modes are found by the decomposition of ocean assimilation data. We also studied the roles of equatorial waves on the process of El Niño in terms of vertical baroclinic modes. The new theory for ENSO decadal variability, in the light of vertical modes, during the Pacific climate change is suggested. The author developed a reliable coupled general circulation model and showed the features of the vertical structure in simulation data.

We investigated the mean characteristics of the baroclinic mode contributions to surface zonal current and sea level pressure anomalies during the period of 1950–1997 through the decomposition of ocean assimilation data. The first three modes have different distributions of variability contributed to zonal current anomalies and also to sea level pressure anomalies.

The first baroclinic mode is more energetic in western Pacific, whereas the second and third mode have larger energy in central-eastern equatorial Pacific. This
is because the thermocline slope has a strong zonal gradient with stronger stability in eastern Pacific, thereby the wind stress forcing are projected onto the each baroclinic mode in different way and eventually leads to the modal distribution. Consequently, the increased stratification leads to the more excited or forced higher-order modes. Analysis of idealized buoyancy frequency profiles supported that the both stronger stratification and thicker thermocline result in increasing contribution of higher modes. The experiments of simple ocean model, including three modes, indicated that when atmospheric forcing projected onto baroclinic mode in zonal different manner, the rms of baroclinic zonal current anomalies have distributions similar to observation.

Subsurface temperature during last a few decades indicates that the role of baroclinic modes on the interannual variability in equatorial SST anomalies may have been changed with time through the changes in vertical stratification. This stimulated the study to identify the impact of baroclinic mode on development of El Niño.

The characteristics of equatorial long waves are shown in terms of vertical baroclinic modes. We found that the first two baroclinic modes could contribute to development of the 1997 – 1998 El Niño. At first stage of the event, the anomalous westerly wind stress forced the first baroclinic mode through the Kelvin wave response and that initiated the warm SST anomalies in eastern Pacific. The second baroclinic mode is gradually increased in its impact on warm SST anomalies. The mature phase of the event is concurrent with an increased contribution of the second mode which leads to strongest El Niño. The transition to cold event are driven by
the reflected Rossby waves from the Kelvin waves at eastern boundary. Particularly, the equatorial westward currents associated with the Rossby waves for second baroclinic mode played a major role on controlling the position of eastern edge of warm pool. This indicates that the multi-baroclinic modes should be considered to improve the ability to predict the ENSO.

The second mode determined the duration of 1986–1988 El Niño. At this event, the first baroclinic mode alone did not initiate the warm event. Whereas the equatorial Kevin waves associated with the second baroclinic mode caused the warm event. With the help of Yeh et al. [2001] and this study, we carefully stated that when the second (first) baroclinic mode are well coupled to central mode of atmosphere, the duration of warm event tend to be longer (shorter). It is also indicated that the second baroclinic mode is essential to drive the interannual (3–4 yr) variability in the equatorial Pacific. We have to, however, demonstrate this relationship in other warm cases, this is under consideration.

In simulation with simple ocean model, including more vertical modes leads to better capture the equatorial interannual variability. So I have to emphasize again the one point that the predictive skill of the intermediate model is increased by including the higher baroclinic modes. This idea could be easily made by the observational results of 1997–1998 and 1986–1988 events. If one predicts well the amplitude and propagating features in second (or third) baroclinic mode, then the properties of El Niño are best forecasted. This is because the inclusion of higher-order modes produces more realistic initial conditions and better described the memory of the coupled system.
After the late 1970s, the contribution for the second and the third baroclinic modes significantly increases in the central Pacific. We found that the changes in the behavior of ENSO are associated with oceanic vertical structural changes in temperature that tend to increase the contribution of higher-baroclinic modes. Simple model experiment supported that when atmospheric forcing projects more onto the higher baroclinic modes, the amplitude and dominant period of ENSO increases similarly to what is observed after the Pacific decadal shift. We suggested this process can be another hypothesis for ENSO decadal variability (EDV).

However, it is difficult to conclude whether the vertical structure variability due to climate change results in EDV based on relatively show observational record alone. We will investigate this theory from a long-term simulation of general circulation coupled model.

The author developed a Coupled GCM (CES CGCMv2) composed of SNU-AGCM and OGCM (GFDL MOM2) with revised ocean mixed model. The model simulates well the Pacific mean field (temperature, currents and precipitation), particularly oceanic thermal structure which is responsible for interannual variability. But the equatorial cold tongue tends to be too strong and extend too far west. The central Pacific has a too diffused thermocline. The ENSO-like climate variability is associated with recharge-discharge oscillator.

With strip-down model, we found that the westward propagation signal in CGCM is resulted from the zonal advection feedbacks. These feedbacks are due to zonal mean temperature gradient in equatorial Pacific accompany by strong cold tongue. Therefore, to improve the simulation of interannual variability, we should
focus on the mean climate field in CGCM.

Analysis for vertical baroclinic modes suggested that the second mode is strongly favored in equatorial central Pacific. This is associated with the too diffused thermocline. Because the diffused thermocline could trap the energy to the upper ocean, the high-order modes were strongly activated. To remove this bias in CGCM, the more mixing in upper ocean are needed.

In summary, the equatorial interannual variability is strongly affected by oceanic vertical mode through air-sea coupled process. Particulary, the contribution of second mode has to be included to understand it. For example, the energetic second baroclinic mode with a slower wave speed increases the time scale of the ENSO and also amplifies the strength of anomalies. It may be partly because the second mode controls the transition mechanism over the contribution of the first mode. Since the contributions of baroclinic mode have been changed with time through the change of subsurface temperature, we have to consider the vertical stratification to understand or forecast interannual variability. The increased stratification leads to more contribution of higher-order modes and eventually lengthen the El Niño oscillation and amplitude. This mechanism, we suggested from the observational and model study, may be one of the possible mechanisms to understand the modulation of ENSO.
Appendix A

Continuously stratified model

We consider the equilibrium state to be perturbed is at rest, so the distribution of density and pressure is the hydrostatic balance. For a simplicity, in the absence of rotation and of friction, the set of linearized equations is

\[ u_t = -\rho_0^{-1} p_x \]  \hspace{1cm} (A.1)  
\[ v_t = -\rho_0^{-1} p_y \]  \hspace{1cm} (A.2)  
\[ p_z = -\rho g \]  \hspace{1cm} (A.3)  
\[ u_x + v_y + w_z = 0 \]  \hspace{1cm} (A.4)  
\[ \rho_t + w \frac{d\rho_0}{dz} = 0, \]  \hspace{1cm} (A.5)

where \( \rho_0(z) \) is the undisturbed density and \( g \) is the gravity. We follow the formalism of Gill [1982].

The relation between \( w \) and \( p \) can be obtained by eliminating \( \rho \) from (A.3)
and (A.5) to give

\[-\rho_0^{-1} p_{zt} = N^2 w\]  \hspace{1cm} (A.6)

where \(N(z)\) is a quantity of fundamental importance to the problem defined by \(N^2 = -g\rho_0^{-1} \frac{d\rho_0}{dz}\). \(N\) has the dimension of frequency, and is variously known as the Brunt-Väisälä frequency [Brunt, 1927; Väisälä, 1925], the buoyancy frequency, and the stability frequency. This is because the frequency of oscillation is \(N\) when the motion is at purely vertical [Holton, 1992]. The two momentum equations (A.1, A.2) and the continuity equation (A.4) give rise to

\[(\partial_{xx} + \partial_{yy})p = \rho_0 w_{zt}\]  \hspace{1cm} (A.7)

The two equations (A.6) and (A.7) clearly show separable solutions of the form

\[w(x, y, z, t) = \phi(z) \tilde{w}(x, y, t)\]  \hspace{1cm} (A.8)

\[p(x, y, z, t) = \psi(z) \tilde{p}(x, y, t)\]  \hspace{1cm} (A.9)

where \(\phi\) and \(\psi\) satisfy

\[\rho_0(z)^{-1} \psi = c^2 d\phi/dz\]  \hspace{1cm} (A.10)

\[\rho_0(z)^{-1} d\psi/dz = -N^2(z) \phi\]  \hspace{1cm} (A.11)

and \(c\) is a separation constant with the dimensions of velocity.

The equations (A.10) and (A.11) can be reduced to a single equation for \(\phi\) is

\[\rho_0^{-1} \frac{d}{dz} \left( \rho_0 \frac{d\phi}{dz} \right) + \frac{N^2}{c^2} \phi = 0\]  \hspace{1cm} (A.12)
This is of the Sturm-Liouville from describing normal modes and frequencies of oscillation of a physical system, e.g., the small transverse oscillation of a heavy flexible string under tension [Pryce, 1993]. When the Boussinesq approximation applies (i.e., $\rho_0(z)$ varies slowly compared with $\phi(z)$), (A.12) can be approximated by

\[ \frac{d^2\phi}{dz^2} + \frac{N^2}{c^2} \phi = 0 \]  

(A.13)

For a continuously stratified ocean, there is an infinite sequence of possible values (eigenvalues) of $c_\tau$ which could be arranged in descending order, and the corresponding eigenfunctions, the normal modes, are denoted by

\[ c_n, \quad n = 0, 1, 2, 3, \cdots \]  

(A.14)

\[ \psi_n(z), \quad \phi_n(z), \quad n = 0, 1, 2, 3, \cdots \]  

(A.15)

Mode zero ($n = 0$) is the barotropic mode which the solution of (A.13) is the one obtained when $n = 0$. Barotropic mode does not depend on the stratification of the ocean but is affected by the total depth of ocean.
Appendix B

WKB analysis of a

Sturm-Liouville problem

We know from our discussion of the oceanic vertical mode that the vertical structure of the vertical velocity constitute an eigenvalue problem of Sturm-Liouville form. While the traditional form of Sturm-Liouville has some kind of complexities, our boundary-value problem can be expressed as

\[
\frac{d^2 G(z)}{dz^2} + \frac{N^2(z)}{c^2} G(z) = 0 \quad (B.1)
\]

Here, \( N(z) = \left(-\frac{g}{\rho} \frac{\partial \rho}{\partial z}\right)^{\frac{1}{2}} \) is the Brunt-Väisälä frequency and \( c \) is the wave phase speed. The rigid-lid and flat-bottom boundary conditions applicable to the baroclinic solutions of interest here are that the vertical velocity vanish at the sea surface
and ocean bottom; that is,

\[ G = 0 \quad \text{at} \quad z = 0 \quad (B.2) \]

\[ G = 0 \quad \text{at} \quad z = -H, \quad (B.3) \]

where \( H \) is the ocean depth.

The ordinary differential equation (B.1) and boundary condition (B.2, B.3) for the vertical structure of the vertical velocity has an infinite number of nontrivial solutions. The eigenvalues \( c_1^{-2}, c_2^{-2}, c_3^{-2}, \ldots \) are discrete, non-degenerate (eigenvalues associated with different eigenfunctions are unequal), and are all positive real numbers; the \( n \)th eigenvalue \( c_n^{-2} \) is associated with the eigenfunction \( G_n(z) \). Although not sufficiently accurate for some applications, the so-called WKB theory [Morse and Feshbach, 1953] may be used to find approximate formulas for \( c_n \) and \( G_n(z) \). It was also (and still is) called the Liouville-Green approximation tackled by Liouville [1837] and Green [1838].

We can introduce new coordinates

\[ \Phi = \int_{-H}^{z} \tilde{N} \, dz, \quad W = \tilde{N}^{\frac{1}{2}} G \quad (B.4) \]

where \( \tilde{N}^{2} = \frac{N^{2}}{c^{2}} \). Then the partial derivatives can be transformed as follows

\[ \frac{\partial}{\partial z} = \tilde{N} \frac{\partial}{\partial \Phi} \quad (B.5) \]

\[ \frac{\partial^2}{\partial z^2} = \tilde{N}^{2} \frac{\partial^2}{\partial \Phi^2} + \tilde{N} \frac{\partial \tilde{N}}{\partial \Phi} \frac{\partial}{\partial \Phi} \quad (B.6) \]

Substituting (B.6) and (B.4) into (B.1) gives

\[ \frac{d^2}{d\Phi^2} \left( \tilde{N}^{-\frac{1}{2}} W \right) + \tilde{N}^{-1} \frac{d \tilde{N}}{d \Phi} \frac{d}{d \Phi} \left( \tilde{N}^{-\frac{1}{2}} W \right) + \tilde{N}^{-\frac{1}{2}} W = 0 \quad (B.7) \]
Using chain rule, then
\[
\frac{d^2 W}{d\Phi^2} + \frac{1}{4} \tilde{N}^{-2} \left( \frac{d\tilde{N}}{d\Phi} \right)^2 W - \frac{1}{2} \tilde{N}^{-1} \frac{d^2 \tilde{N}}{d\Phi^2} W + W = 0 \quad (B.8)
\]
Next, we can rewrite (B.8) as a simple form
\[
\frac{d^2 W}{d\Phi^2} + (1 + \varepsilon) W = 0 \quad (B.9)
\]
where
\[
\varepsilon = \tilde{N}^{-\frac{3}{2}} \frac{d^2}{dz^2} \left( \tilde{N}^{-\frac{1}{2}} \right) \quad (B.10)
\]
Thus if \( \varepsilon \ll 1 \), the approximate solution of (B.9) is
\[
W = Ce^{\pm i\Phi} \quad (B.11)
\]
which is equivalent to giving the approximate solution of (B.1) as
\[
G^{WKB}(z) = C\tilde{N}^{-\frac{3}{2}} \exp \left( \pm i \int_{-H}^{z} \tilde{N}dz' \right) \quad (B.12)
\]
where \( C \) is an arbitrary constant. The two sign possibilities in (B.12) can be superposed to construct either sine- or cosine-like behavior for the solution.
\[
G^{WKB}(z) = \tilde{N}^{-\frac{3}{2}} \left[ A \cos \left( \int_{-H}^{z} \tilde{N}dz' \right) + B \sin \left( \int_{-H}^{z} \tilde{N}dz' \right) \right] \quad (B.13)
\]
where \( A \) and \( B \) are arbitrary constants.
The bottom boundary condition (B.3) implies that \( A=0 \). Substituting for \( \tilde{N} \) from it’s definition, the WKB approximation for the \( n \)th baroclinic vertical velocity eigenfunction is
\[
G^{WKB}(z) = B \left( \frac{N(z)}{c_n} \right)^{-\frac{1}{2}} \sin \left( \frac{1}{c_n} \int_{-H}^{z} N(z')dz' \right) \quad (B.14)
\]
The boundary condition (B.2) determines the eigenvalues

\[ c_n^{WKB} = \frac{1}{n\pi} \int_{-H}^{0} N(z')dz' \quad (B.15) \]

which is given easily from \( \sin \left( \frac{1}{c_n} \int_{-H}^{0} N(z')dz' \right) = 0. \)
Appendix C

Additional results of *Yeh et al.* [2001]

In this chapter, we should discuss the results of *Yeh et al.* [2001], but did not be provided in their study. The higher-order modes tend to generate slow interannual variation associated with slow phase speed in simple coupled model. They did not address to the question that is *what controls the ENSO period in simple model?* To figure it out, we have special attention to the dominant two terms in the equatorial SST tendency terms. Although the full equation has more other tendency terms, the zonal advection by anomalous zonal current and vertical advection through mean upwelling have been considered because they play a dominant role on development of SST anomalies [*Dewitte*, 2000; *Kang et al.*, 2001].

We carry out three experiments using simple coupled model as *Yeh et al.* [2001]: the experiments differs only in the way mean upwelling of anomalous tem-
perature term is controlled by the thermocline displacement $\delta \xi$. In Exp1, $\delta \xi$ is estimated from the contribution of the first baroclinic mode. Similarly, Exp2 and Exp3 correspond to simulations where the second and third mode, respectively, determined the subsurface temperature. The reader is strongly invited to refer to Yeh et al. [2001] for detail description of model experiments.

Figure C.1 presents the time-longitude plots along the equatorial SST tendency term of $-u(\bar{T} + T)x$ overlaid upon the SST anomalies for three experiments. This term has strong contribution on the SST variability when the thermocline displacement is determined by the first baroclinic mode. Whereas, the term of $-\gamma M (\bar{w} + w) \frac{T_{\text{sub}} - T_{\text{hub}}}{H_1}$ has strong impact on SST anomalies and tend to lengthen the period of ENSO when the higher-order modes only determined the thermocline displacement (Figure C.2).

This results are quite well consistent with Dewitte [2000] in which the higher-order modes favored the equilibrium between the thermocline displacements and wind stress leading to an unstable Kelvin mode propagating slowly eastward during the El Nino. He also showed that the gravest mode excited Rossby unstable modes.
Figure C.1: Time-longitude plots of the equatorial SST tendency term overlaid upon the contoured SST anomalies. The anomalous advection of the total zonal temperature gradient from (a) Exp1, (b) Exp2 and (c) Exp3 of Yeh et al. [2001]. Contour interval is 1°C and negative values are dashed. Shading unit is °C (10 day)$^{-1}$.
Figure C.2: The same as Figure C.1, except for SST tendency term from climatological upwelling of anomalous vertical temperature gradient.
Appendix D

Additional results of Moon et al. [2004]

We also should provide the another results, not have been addressed in Moon et al. [2004]. As shown in Appendix C, we computed the SST tendency terms from model experiments. The reader is invited to refer to Moon et al. [2004] for description of model experiments.

The figure, for sea level pressures with significant increase in the central equatorial Pacific associated with the increase of stratification, is also shown (Figure D.1). It is also indicated that the increased contribution of higher-order mode in simple coupled mode gives rise to larger contribution of thermocline feedback to development of SST anomalies, so that the SST anomalies are amplified and would oscillate slowly (Figure D.2 and Figure D.3).
Figure D.1: The difference of rms variance (cm) between the two periods for (a) mode 1, (b) mode 2 and (c) mode 3. Contour intervals are 0.5 cm, 0.5 cm and 0.1 cm, respectively.
Figure D.2: Time-longitude plots of the equatorial SST tendency term overlaid upon the contoured SST anomalies. The anomalous advection of the total zonal temperature gradient from (a) Exp1 and (b) Exp2 of Moon et al. [2004]. Contour interval is 1°C and negative values are dashed. Shading unit is °C (10 day)$^{-1}$. 
Figure D.3: The same as Figure D.2, except for SST tendency term from climatological upwelling of anomalous vertical temperature gradient.
Appendix E

Supplementary figures

In this chapter, we represented the figures which are not addressed in previous chapters.

Figure E.1 and Figure E.2 show that the contributions of third baroclinic mode to sea level pressure could be neglected, since their magnitude are so small compared to the those of first two modes.

When three baroclinic mode are used to simulate the low frequency variability in equatorial Pacific, the correlation along the equator between the simulation and the observations is increased by about 20% compared to the simulation with one baroclinic mode (Figure E.3 and Figure E.4).
Figure E.1: Time-longitude plots of sea level anomalies along the equator for the contribution of (a) the first baroclinic mode, (b) the second one and (c) the third mode. Units are cm and the values greater than 2 cm are shaded.
Figure E.2: The same as Figure E.1 but for 1986 – 1988 El Niño. Units are cm and the values greater than 1 cm are shaded.
Figure E.3: Time-longitude distribution of SST anomalies along equator. (a) SODA SST anomalies, (b) and (c) are simulated anomalies when the ocean model is forced by SODA wind stress anomalies with the first mode only and three modes included, respectively.
**Figure E.4:** The correlation coefficient along the equator between the SODA SST anomalies and (dashed) EXP1, (solid) EXP2 of Figure E.3. Note that the difference between the EXP1 and EXP2 is that the first baroclinic mode only included to ocean model in EXP1, but the mode has three modes in EXP2. The correlation are increased in EXP2 by 20% rather than EXP1.
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초 록

해양자료동화 자료를 분석하여 적도 태평양에서의 연직 경압모드의 특징을 찾고, 경압모드의 변동이 태평양 해수면 온도의 경년 주기와 심년 주기 변동에 미치는 영향을 분석하였다. 특히, 1970년대 전후로 발생한 엔소의 심년변동을 설명하는 새로운 이론을 제시하였다. 그리고 신뢰할 만한 대기-해양 집합모형을 개발하고, 집합 모형이 모의하는 평균 장의 특징과 경압모드의 구조를 분석하여 집합모형의 개선 점을 제시하였다.

동서 방향의 표면 호름과 해수면 기압에 기여하는 해양의 연직 경압모드의 크기와 수평 분포를 계산하였다. 경압모드의 기여도는 부력 전동수로 이루어진 Sturm-Liouville 방정식 해를 원래 자료에 투영시켜 얻었다. 적도 태평양에서 경압모드의 변동성은 각각 모드마다 서로 다른 분포를 나타내었다. 첫번째 경압모드는 서태평양에서 가장 큰 변동을 보인 반면, 두번째와 세번째 모드는 중태평양과 동태평양에 걸쳐 크게 변동하였다.

이처럼 주된 변동 지역이 모드별로 다른 이유는 연직 방향의 성층화 정도가 각 해양에서 서로 다르기 때문이었다. 즉, 동태평양처럼 (서태평양처럼) 성층화 정도가 커지면 (작아지면) 두번째와 세번째 (첫번째) 경압모드의 변동성이 크게 나타나. 해양의 성층화 정도가 경압모드에 영향을 주는 이유는 연직 방향으로의 에너지 수송과 각 모드의 연직 구조의 특징 때문이다. 만약 해양 상층에 안정도가 크다면, 바람에 의한 에너지가 상층에 간하게 되어 연직 규모가 작은 (즉 상위 경압모드) 모드를 활성화 시킬 것이다.


새로운 대기-해양 집합 모형을 (CES CGCMv2) 개발하였다. 이 모형은 기존의 모형 (CES CGCMv1) 보다 열대 태평양의 모형에 더욱 개선된 성능을 나타냈다. 또한 다른 집합모형들에 비해 경년변동의 크기와 태평양 내부 순환응도가 잘 모의된다. 모형에서 해수면 온도 분포에서 한랭해가 중태평양으로 치우치는 문제점이 나
타났지만, 그 정도는 다른 모형들의 결과와 비슷하였다. 강수량 분포에는, 동태평양에 이중 적도 수렴대가 나타나고 있다. 모형은 관측과 유사한 크기와 주기를 갖는 강면 변동을 보였다. 또한 적도에서 동서 방향으로 흐르는 내부 환류가 기존의 모형에 비해 크게 개선되었다.

특히, 접합 모형은 중태평양에서 수온 약층의 분산된 모습을 보였다. 이런 수온 약층의 분포는 상층의 큰 안정도를 나타낸다. 모형에서 모의된 결과를 이용한 연직 경밀모드는 두번째와 세번째 모드가 관측에 비해 훨씬 큰 영향을 주고 있었다. 이런 특징은 해양자료통화 자료의 분석을 통해 얻은 결과와 잘 일치한다. 따라서 모형의 모의 성능을 더욱 개선하기 위해서는 해양 상층에 좀더 많은 혼합 과정이 필요하다는 것을 제시할 수 있다.

주요어: 연직 경밀모드, 엔소, 강면 변동, 삼면 변동, 대기-해양 접합 모형, 기후 변화

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