Implementing the Pacific Basin Extended Climate Study (PBECS)

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1. Introduction

The objective of the Pacific Basin Extended Climate Study (PBECS) is to observe and model the Pacific Ocean and overlying atmosphere well enough that the evolution of climate phenomena can be quantitatively diagnosed within the framework of data-assimilating models. This is essential to understanding the phenomena of climate variability, and the first step toward an observational and modeling system capable of improved climate forecasting. The PBECS approach is to obtain, for 15 years over the Pacific Basin, observations to constrain data-assimilating models well enough that competing theories of climate variability can be tested and models of that variability improved. The observations will include global satellites, long-term in situ observations from other planned and ongoing programs including the El Niño-Southern Oscillation (ENSO) Observing System and Argo, and new measurements put in place by PBECS to observe specific components of the climate system. This observing system must be adequate to measure and constrain within assimilating models, and on the scales important to climate phenomena, key processes like advection, air-sea fluxes, and the effects of sub-grid-scale mixing. The observations will be used directly in empirical studies to better understand phenomena, and in conventional comparisons with forward models. The novel aspect of PBECS, however, is the attention placed on using and improving data-assimilation models and procedures to elucidate the processes and dynamics of climate variation.

While the PBECS methodology may well discover new climate phenomena, the specific phenomena of immediate interest that were used to design the observing system are:

Decadal modulation of El Niño-Southern Oscillation (ENSO). The historical record over the last 100 years, and recent experience predicting ENSO, make clear that ENSO cycles have developed differently, and that some decadal periods are characterized by a prolonged tendency toward one phase of the cycle. Explanations for this vary from subtleties in atmospheric triggering, to nonlinear chaotic development of the ENSO cycle itself, to decadal changes in the "background" state of the equatorial ocean. Understanding the cause for decadal differences in ENSO evolution is key to improved climate forecasting. Examining various hypothesized mechanisms for decadal modulation of ENSO will be a primary goal of PBECS, with particular attention placed on the air-sea coupling in ENSO, and on oceanic advective pathways that bring anomalies into the equatorial zone where they can affect ENSO evolution.

Decadal variability. The instrumental record over the last 100 years also makes clear that large-scale, coupled decadal variations have occurred in both the North Pacific atmosphere and ocean. The dominant pattern of change, the Pacific Decadal Oscillation, has been connected to climate changes over North America and to fisheries cycles (Mantua *et al.*, 1997) in the North Pacific. Theoretical and model studies have proposed mechanisms by which the ocean participates in the observed decadal variability either actively or passively. PBECS should lead to a better understanding of what causes SST variability outside the equatorial zone, and support diagnosis of future decadal variability and assessment of its predictability.

The dynamics of ENSO. ENSO is the only demonstrably predictable climate phenomenon. Conceptually, relatively simple waveguide dynamics modulate thermocline depth which affects equatorial sea-surface temperature (SST). SST changes the overlying atmospheric convection and wind stress patterns and this, in turn, is atmospherically teleconnected over a significant fraction of the globe. While models successfully simulate ENSO, the relative importance of the different mechanisms involved remains uncertain. Different ENSO models depend on different mechanisms in their upper-ocean heat budgets and, consequently, evolve differently. PBECS will provide improved assimilated data sets for diagnosing future ENSOs, provide insight into ways dynamical models could better simulate ENSO evolution and, hopefully, improve ENSO forecasts through improved observations available to forecasters and improved understanding and modeling of the processes.

The idea of using data-assimilating models to link together basin-wide observations in order to diagnose and understand climate variability arose from a series of meetings aimed at developing ways for U.S. scientists to help achieve the goals of CLIVAR. As a result of those meetings the scientific basis for PBECS was developed and described in "Prospectus for a Pacific Basinwide Climate Study" (Lukas *et al.*, 1998) which can be downloaded from `http://www.usclivar.org`. The purpose of this implementation plan is not to reiterate this science plan's motivation for studying climate variability in the Pacific sector, but rather to lay out in practical terms how the important climate processes can be studied and understood. Achieving success will certainly require the cooperation of scientists from many nations around the Pacific. As this cooperation develops, the plans of U.S. scientists will certainly evolve. The PBECS implementation plan will also evolve on an approximately annual basis for the next few years.

Observing climate processes throughout an entire ocean basin is motivated by the success of the ENSO observing system in leading to skillful ENSO predictions. It is made feasible by high-quality satellite altimetry, new technologies that dramatically reduce the cost of making long-term in situ ocean observations, and the rapidly growing capabilities of dynamical models to blend observations into analyzed data sets. The general strategy is to use assimilating models and the high spatial and temporal resolution afforded by satellites to knit together cost-effective in situ observations. The new in situ technologies that make the PBECS approach feasible include:

- the TAO array that has a demonstrated ability to observe the equatorial winds and the heat redistribution that accompanies ENSO evolution,
- profiling autonomous floats for broad-scale sampling of temperature, salinity and velocity, even across the broad ocean expanses not visited by commercial shipping,
- drifters that allow the surface circulation and its variability to be observed directly,
- repeated high-resolution temperature and salinity sections, based on expendable probes deployed from Volunteer Observing Ships (VOS), that measure the lateral heat fluxes of heat and fresh water,
- a combination of VOS observations, satellite altimetry, moored profilers, and autonomous gliders to economically measure the mass and heat transports of concentrated boundary currents,
- accurate and reliable surface meteorological sensors that allow the air-sea fluxes of heat, fresh water, and momentum to be determined from moored buoys and VOS.

Sustained satellite observations are a necessary element of this strategy, including:

- altimeter mapping of the sea surface height;
- vector wind maps from a high-resolution scatterometer of the quality of QuickSCAT; and
- SST from AVHRR or microwave radiometers.

Even with frequent and relatively high-resolution satellite observations to knit them together, making in situ observations dense enough to coherently observe all the processes of climate variability is not practicable. The success of PBECS depends, therefore, on recent significant improvements to data-assimilation techniques and the models upon which they are based. These techniques will become substantially more effective as more powerful computers become available. While much work will be required to realize the promise of assimilating models, there are several groups now actively pursuing this. PBECS will provide data to constrain these analyses and will include empirical studies to test the resulting data sets with the intent of improving the models upon which they are based. In this way PBECS will apply a process-experiment approach to climate phenomena that span basins and decades. Others, outside PBECS, are expected to exploit the assimilated data sets to study climate predictability and to develop improved prediction schemes.

Because the timescales of ocean and atmospheric processes are so different, air-sea fluxes have particular acuity in diagnosing the processes involved in the coupled phenomena of climate. The ocean is driven by the air-sea fluxes of momentum, heat, and fresh water, while the fluxes of heat and latent heat (fresh water) are the mechanisms by which the ocean imposes long-time-scale variability on the otherwise rapidly evolving atmosphere. Despite their importance, available regional flux estimates are in error by much more than is required to generate significant climate anomalies. Improving large-scale surface fluxes is a primary PBECS goal. The general strategy is to use a combination of high-quality surface observations, inferences of fluxes from satellites, and constraints developed from the ocean-data assimilation process to improve the methods by which operational centers can combine all available information, including model analyses and forecasts, to prepare flux fields.

PBECS will require changes in the community. Because progress depends on rapid data exchange among observers, data analysts, and modelers, and because the data is meant to support operational activities, all PBECS data will be made public and disseminated as quickly as is technically feasible. Because of the focus on long timescales, it will be necessary to maintain PBECS for 15 years. This will place a premium on organization to insure that needed observations and analyses are maintained as the interests of individual participants change. It will also require agencies to provide continuity of funding for decadal timescales.

While PBECS specifically addresses climate variability on the basin scale over the long term, success in diagnosing and predicting this variability will depend on improved understanding of processes that can be gained only through process experiments of high intensity and limited duration. This implementation plan includes short descriptions of several process experiments that will be needed to improve the models upon which PBECS analysis is based. Although these process experiments are not strictly part of PBECS, they must be coordinated with its fieldwork and will benefit from the data fields PBECS analyses make possible.

2. The Scientific Basis and Goals for PBECS

As described in greater detail in the PBECS science plan (Lukas *et al.*, 1998) the main climate phenomena to which PBECS is addressed are decadal modulation of ENSO, decadal variability, such as the Pacific Decadal Oscillation, and the dynamics of ENSO itself. The purpose of this section is to briefly review what needs to be learned about these related phenomena, and how that might be done with a coordinated program of sustained observations, process experiments, and data-assimilating modeling.

2.1 Decadal variability of ENSO

Despite significant success in predicting the 1997–98 El Niño and many of its remote consequences, it is apparent from the historical record that ENSOs vary greatly. Indeed, models that are very successful at predicting one ENSO event can fail completely on the next. The long record of the Southern Oscillation Index (the difference in sea level pressure (SLP) between Darwin and Tahiti) indicates that both the amplitude and period of the ENSO cycle has varied considerably over the past century. The Rasmusson and Carpenter (1982) composite of the events between 1950 and 1972 suggests a sequence of anomaly patterns during El Niño events that was not observed in some subsequent events. Event by event comparison of the ten events since 1950 suggests that there is a typical sequence of large-scale anomaly evolution common to the mature phase of events even though their initial evolution and the duration of their ending phase varies considerably (Harrison and Larkin, 1998).

Over the last decade there has been vigorous investigation of possible causes for the variability of ENSO, and of its predictability. The chaotic evolution of ENSO's somewhat nonlinear dynamics has been implicated while details of the wind forcing that triggers the El Niño phase may hold the key. Models have, however, shown various mechanisms for ENSO modulation that depend on deterministic variations of the coupled ocean-atmosphere system. The common thread of these hypotheses is that a low-frequency process, in many cases one involving the subtropics, leads to changes of the "background state" upon which ENSO evolves. These models are described here to explain how PBECS can lead to an understanding of the decadal modulation of ENSO, and hopefully to an improvement of ENSO prediction skill.

2.1.1 *Atmospheric and western-boundary processes*

As discussed in the following section, several studies suggest that the tropics are capable of developing decadal variability on their own, and this variability could modulate the background upon which ENSO evolves. Other studies have explored the idea that decadal variability is generated at midlatitudes, and that this variability is then transmitted into the tropics via either atmospheric or oceanic teleconnections. For example, Barnett *et al.* (1999a) suggest that the zonal wind-stress anomaly pattern associated with North Pacific decadal variability extends into the tropics and thereby modulates El Niño by changing the east-west thermocline gradient. Lysne *et al.* (1997) and Liu *et al.* (1999) investigate the possibility that signals can be communicated to the tropics by the propagation of coastal Kelvin waves along the western boundary. Liu *et al.* (1999), however, argue that before getting to the tropics by this mechanism the thermal signals are attenuated by an order of magnitude (see also Kessler, 1991). These studies are not definitive, however, and it remains an important question whether western boundary reflection is important, or if anomalies must either originate at relatively low latitudes or be amplified by air-sea feedbacks as they proceed through the tropics.

2.1.2 *The subtropical cell*

Several hypotheses for decadal climate variability rely on changes in the heat transport of the North Pacific Subtropical Cell (STC), the shallow vertical overturning circulation that carries northern-hemisphere thermocline water into the tropics. This overturning circulation consists of subduction in the subtropics, equatorward transport of thermocline water into the tropics, equatorial upwelling, and a return flow of near-surface water to the subtropics (McCreary and Lu, 1994; Liu *et al.*, 1994; Lu and McCreary, 1995; Hirst *et al.*, 1996; Lu *et al.*, 1998); also see Section 6. Gu and Philander (1997) (hereafter shown as "GP") hypothesize that midlatitude heat flux anomalies alter the temperature of the equatorward-flowing branch of the STC, and that these anomalies are eventually advected to the equator to affect the cold tongue. Subsequent work notes that the method of propagation of these temperature anomalies depends on whether they are salinity compensated in density. Without such compensation, propagation is partly governed by planetary-wave dynamics (Liu, 1999; Huang and Pedlosky, 1999), whereas with compensation it is determined only by passive advection. The GP hypothesis is supported by Deser *et al.* (1996), who report a subsurface temperature anomaly circulating within the North Pacific STC, and by Bingham and Lukas (1994), who discuss anomalous advective transports of heat and salt by the NEC and the Mindanao Current. Schneider *et al.* (1999a,b) and Nonaka *et al.* (1999), however, argue otherwise, commenting that an interpretation of the GP idea as resulting from the propagation of planetary waves does not explain much of the observed tropical temperature variability. The role of salinity-compensated temperature anomalies (i.e., without density change, called “spiciness” anomalies) is not addressed in either study due to the paucity of observations of salinity and the freshwater flux.

Kleeman *et al.* (1999) suggest that changes in STC transport, rather than temperature, cause tropical decadal variability. In their coupled solutions, midlatitude wind-stress anomalies excite Rossby waves that reflect from the western boundary as coastal and equatorial Kelvin waves, thereby propagating midlatitude signals into the tropics. There, they alter the tropical thermal structure, leading to changes in the size and/or strength of the equatorial cold tongue. Analyzing a coupled ocean atmosphere-model, Schneider (1999) argues that changes of wind stress and ocean circulation near 10–15°N generate salinity-compensated thermal anomalies on isopycnal surfaces by anomalous advection that are subsequently advected to the equator with the mean circulation.

To summarize, Gu and Philander (1997) argue that advection of temperature anomalies by the mean flow couple mid and low latitudes, while Kleeman *et al.* (1999) conclude that advection of the mean temperature field by anomalous currents provides this coupling. Schneider (1999) combines these two approaches by invoking anomalous advection in the generation and mean advection in the equatorward spread of these anomalies. A major difference between these hypotheses lies in the dynamics of the spreading of the signals. Adjustments of the flow field are governed by planetary waves, whereas the processes invoked by Schneider (1999), and implicitly by Gu and Philander (1997), involve oceanic salinity and the passive advection of oceanic spiciness anomalies.

2.1.3 Equatorial response

No matter how thermal anomalies reach the equator, those associated with changes of the oceanic density alter the depth of the zonally averaged thermocline, rather than its east-west slope, since they do not require changes in the equatorial zonal wind for their existence. This zonally averaged change can alter SST by changing the efficiency with which upwelling cools the surface. Emerging spiciness anomalies at the equator, on the other hand, do not alter the oceanic pressure field or circulation in any way, and so directly affect SST at the existing equatorial outcrops. Because equatorial zonal winds respond rapidly and strongly to SST anomalies (Lindzen and Nigam, 1987), even small changes can set off a growing response. It is not known if the equatorial response to the arrival from the higher latitudes of waves or dynamically neutral spiciness anomalies differ.

2.2 Decadal variability of the Pacific

While ENSO is the largest interannual signal, the North Pacific also undergoes substantial variability at decadal periods. Various analyses (Zhang *et al.*, 1997; Mantua *et al.*, 1997; see Fig. 2.1) show that the dominant pattern of Pacific decadal variability has similarities to the patterns of ENSO variability, but has a broader meridional scale and involves relatively more variability in the subtropics and less in the tropics. While we are attuned to the societal value of ENSO predictions in forecasting seasonal climate, A. Leetmaa (1997, private communication) of NCEP pointed out that about half the skill in these forecasts actually comes from North Pacific decadal variability (which is now forecast mainly by persistence). Some speculate that prediction of decadal variability may be even more valuable than seasonal-to-interannual forecasts because they provide greater opportunity for societal adaptation.

Possible causes for decadal variability have been the subject of recent lively debate. The simplest mechanism is that decadal variability is essentially the low-frequency part of the weather spectrum that has been filtered by coupling to the ocean's slow physics. Alternatively, decadal variability may be the low-frequency sidebands of the nonlinear dynamics of ENSO. Increasingly, however, models are suggesting that decadal variability can result from internal feedbacks within the coupled atmosphere-ocean system. Several modeling studies suggest that Pacific decadal variability is associated with self-sustained, basin-scale modes of oscillation. Both the location and type of the generation mechanisms vary among the models, with some suggesting the tropics is the important generation region while others point to midlatitudes. This section reviews some of the theories of decadal variability as guidance to what sorts of processes PBECS must elucidate and the hypotheses it should be able to test.

2.2.1 Stochastic forcing

Even in the absence of feedback from the ocean, stochastic atmospheric forcing of the ocean leads to enhanced oceanic variance at low frequencies, and in some instances to distinct spectral peaks. Hasselmann (1976) and Frankignoul and Hasselmann (1977) first suggested observed low-frequency SST variations resulted when the upper-ocean heat budget integrates surface heat flux, converting a white spectrum of atmospherically driven surface heat flux to a red spectrum of SST. These studies were extended by Barsugli and Battisti (1998) to include feedbacks from the ocean to the atmosphere and by Saravanan and McWilliams (1998) who included the effects of oceanic advection and the preferred spatial scales of atmospheric forcing. They pointed out that, at frequencies corresponding to the advective speed divided by the forcing length scale, SST variance can exhibit a peak due to a process termed "stochastic resonance."

Similar ideas apply to the excitation of planetary waves by stochastic Ekman pumping. Frankignoul *et al.* (1997) developed a model for stochastically forced Rossby waves and show results that agree with simulations of a coupled ocean-atmosphere model, and are broadly consistent with observations at Bermuda. They suggested that stochastic wind-stress forcing may explain a substantial fraction of decadal variability in mid-ocean gyres. When the stochastic forcing has preferred spatial scales smaller than the basin size, they predicted a peak in the spectrum at time scales corresponding to the Rossby wave period associated with the length scale of the atmospheric forcing. Jin (1997a) and Neelin and Weng (1999) found such peaks in uncoupled simulations and pointed out that the underlying dynamics are very similar to those of Saravanan and McWilliams (1998), but with Rossby-wave propagation taking the role of advection.

2.2.2 Western boundary currents

The dynamics of the ocean, itself, can lead to low-frequency variations in the western-boundary currents and their eastward extensions, just in the areas where large low-frequency SST signals are observed (e.g., Zhang *et al.*, 1997). For example, Spall (1996) demonstrated that feedbacks between the Gulf Stream, its northern and southern recirculation gyres, and deep western-boundary currents can lead to self-sustained internal oscillations with decadal periods. These interactions also affect

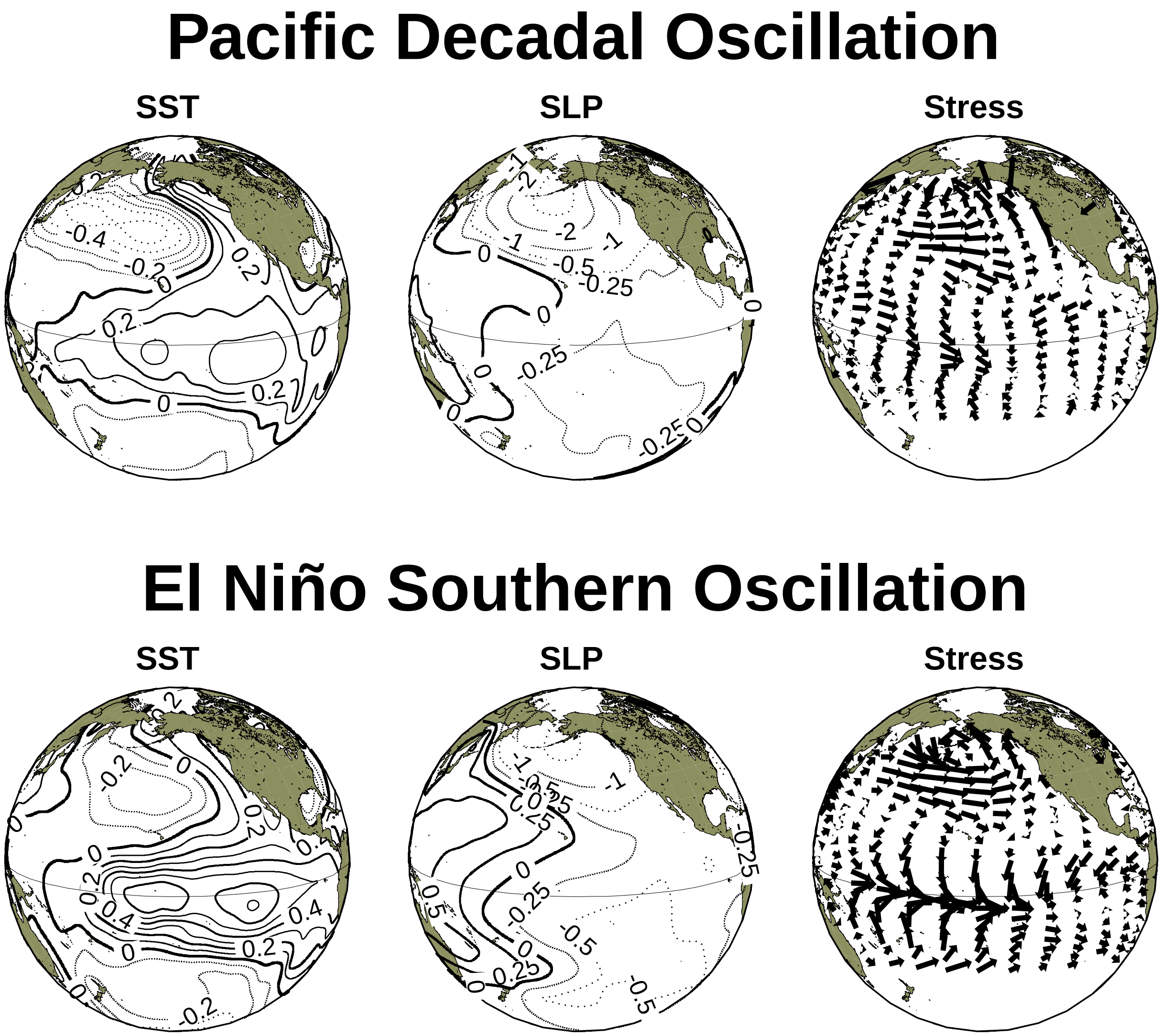

Figure 2.1: Patterns of sea-level pressure (SLP), sea-surface temperature (SST) and wind stress associated with the Pacific Decadal Oscillation (upper) and ENSO (lower).

how far the western-boundary current penetrates into the main gyre, its eddy field, its separation latitude, and the age of the waters it carries. While this oscillation depends on the existence of deep western boundary currents (as exist in the North Atlantic), double-gyre circulations in idealized single-layer models also exhibit self-sustained oscillations with periods of several years (Speich *et al.*, 1995; Jiang *et al.*, 1995; Speich and Ghil, 1994) due to the interaction of the recirculation regions and the western-boundary currents.

2.2.3 Tropical basin-scale modes

In the studies of Knutson and Manabe (1998) and Yukimoto *et al.* (1996, 1999), an internal decadal oscillation develops that involves tropical convection anomalies similar to those that occur during ENSO. Yukimoto *et al.* (1999) argue that the dynamics of this oscillation are essentially those of the delayed oscillator (Suarez and Schopf, 1988; see Section 2.3), with the lagged negative feedback being provided by the propagation of Rossby waves across the ocean. In contrast to the ENSO delayed oscillator, however, subtropical Rossby waves (near 25°N) are invoked rather than equatorial ones. Their slower propagation speeds account for the longer timescale of their decadal signals.

Similarly, Tziperman *et al.* (1994) report a decadal signal in their coupled solution that is generated by tropical ocean-atmosphere interactions. They argue, however, that its nonlinear dynamics generate variability at frequencies lower than interannual, and that it is then communicated to higher latitudes through ENSO-like atmospheric teleconnections (e.g., Alexander, 1992).

2.2.4 Decadal feedbacks in the extratropics

One of the primary candidate mechanisms hypothesized to explain interdecadal climate variability in the Pacific Ocean involves North Pacific Ocean interactions with the Aleutian Low atmospheric pressure system. This process appears to be active in fully coupled ocean-atmosphere models. Latif and Barnett (1994, 1996) used a global coupled GCM to explore a decadal oscillation associated with SST and SLP anomaly patterns similar to observed ones (Latif and Barnett, 1994; Latif *et al.*, 1997; White and Cayan, 1998). In this mode, the meridional SST gradient modulates the strength of the midlatitude westerlies, which in turn change the gyre-scale wind stress curl. A stronger SST gradient (i.e., cooler water to the north) means stronger curl and stronger gyre circulation, eventually increasing the strength of the Kuroshio that then advects warm water to the north, changing the sign of the anomalous SST gradient and leading to an oscillation with a period near 20 years. Since it existed even when tropical feedback was completely suppressed, this mode's dynamics are distinctly different from those of ENSO. Positive feedbacks involving midlatitude wind-stress and heat-flux anomalies were important for the mode's excitation, whereas subtropical gyre adjustment involving coupling of the subpolar and subtropical ocean circulations with storm-track variability were likely responsible for determining its period. Barnett *et al.* (1999b) used a hierarchy of coupled GCMs, atmospheric models coupled to a slab mixed layer, and atmospheric models forced by prescribed surface temperature anomalies, to suggest that North Pacific decadal variability can be independent of tropical dynamics and is composed of both stochastically driven and coupled-mode components. The latter has a period of 20 years that is achieved by gyre adjustment and exhibits a slow growth and eastward propagation of SST anomalies in the Kuroshio extension region.

Coupled modes similar to the Latif-Barnett mode develop in intermediate coupled models examined by Jin (1997a), Xu *et al.* (1998), Neelin and Weng (1999), Goodman and Marshall (1999), and Kleeman *et al.* (1999). For example, the model of Kleeman *et al.* (1999), which uses a statistical (empirically based) atmosphere, develops two distinct modes: an ENSO-like interannual mode and a decadal one. The decadal oscillation has a number of properties similar to the Latif-Barnett mode, including anticyclonic rotation of heat-content anomalies about the North Pacific. The authors argue, however, that the key positive feedback process is the influence of wind-speed anomalies on latent heating in the northeastern subtropics, an idea supported by the "wave-action" analysis reported in Nakamura and Yamagata (1999).

The upper-ocean heat dynamics and atmospheric responses to SST in these idealized models differ dramatically. For example, Jin (1997a) considered a longitudinally averaged heat equation that only depends on the meridional heat transport by the gyre and related the basin-wide average meridional temperature gradient to the atmospheric wind stress. Muennich *et al.* (1998) specified the anomalous wind stress from the upper-ocean depth anomalies at a key area close to the western boundary based on observed correlations. Cessi (1999) considered only the effect of horizontal, geostrophic heat transports on surface temperature, and carefully derived a parameterization of the anomalous wind stress due to changes of atmospheric eddies induced by anomalies of SST. Weng and Neelin (1998) and Neelin and Weng (1999) considered the effect of horizontal heat transport by geostrophic flow and meridional Ekman transports and determined atmospheric response from forced atmospheric-model experiments that they augmented by stochastic forcing. Talley (1999) showed that the growth rates of midlatitude coupled modes, as well as their propagation speeds and directions, critically depend on the assumptions about upper-ocean thermodynamics and the atmospheric response to SST anomalies. Given this sensitivity, observational constraints on upper-ocean physics and atmospheric response are required to decide which of these models is appropriate.

To a first approximation the subtropical decadal feedback mechanisms comprise three key processes: oceanic adjustments to changing wind-stress (e.g., thermocline response to stress curl), SST response to the oceanic adjustment (or heat flux), and atmospheric response to changing SST. We examine what is known about each component process separately.

2.2.4.1 Thermocline response to changing wind-stress curl. Observations reveal that the North Pacific thermocline adjusts to changing wind-stress curl on decadal timescales (Qiu and Joyce, 1992; Yasuda and Hanawa, 1997; Miller *et al.*, 1998; Deser *et al.*, 1999). These studies do not reveal propagating features that can be associated with Rossby waves (as are invoked in simple coupled model theories). Instead, they tend to show that the thermocline exhibits a stationary response that resembles the Sverdrup equilibrium solution and lags it by a few years. While many authors have proposed that the subtropical gyre response to the changing Aleutian Low is the dynamical link, Miller *et al.* (1998) suggest that subpolar gyre adjustment processes are dominant.

2.2.4.2 SST response to changing thermocline. Observations also show that decadal oceanic thermocline adjustment is linked to SST variations in the Kuroshio-Oyashio Extension (KOE) (e.g., Deser *et al.*, 1996; Miller *et al.*, 1998). During the transition to cool KOE SST conditions in the 1980s, the basin-scale pattern of a rising thermocline under the KOE precedes the SST response by several years. The mechanism by which the thermocline alters the KOE SST is not yet clear. Many processes can affect the SST heat budget on decadal timescales: local upwelling or mixing can directly change SST or alter the stability of the water column and change the mixed layer depth; increased horizontal heat transport can affect the SST from geostrophic subpolar or subtropical gyre western boundary currents or from directly forced Ekman currents; a changed mixed-layer depth distribution can alter the SST sensitivity to atmospherically driven surface heat flux forcing. In any case, the SST anomalies in the KOE region appear to be linked to changing large-scale baroclinic ocean dynamics and a goal of PBECS should be to learn why.

2.2.4.3 Atmospheric response to changing SST. Demonstrating atmospheric sensitivity to midlatitude SST anomalies is the most subtle aspect of subtropical feedback hypotheses. Early studies of this process concentrated on SST anomalies in the central North Pacific Ocean where principal component analysis places the maximum SST variability. However, it now appears that this central-basin anomaly may simply be a forced SST mode. Instead, the atmosphere appears to be sensitive to SST anomalies under the storm track region in the KOE. For example, Peng and Whitaker (1999) found a barotropic response to SST anomalies in the KOE as a result of transient-eddy momentum fluxes in the storm track. Ting and Yu (1999) showed that transient-eddy momentum fluxes were important to the response of a nonlinear anomaly model to anomalous heating only when the heating was coincident with the Pacific jet and storm track. Alexander and Scott (1999) found that in the storm track region of the North Pacific the climate of the coupled model and the growth rate of baroclinic waves are significantly different from the uncoupled system forced by climatological SSTs. In the Latif and Barnett model, the SST in the KOE is also the hot spot for atmospheric sensitivity on decadal timescales (Barnett *et al.*, 1999b).

Taken together, these results suggest that more research and better observations are needed to diagnose and validate the possible decadal feedback processes in the North Pacific. Long-term monitoring of the three processes is needed. How this might be done is discussed below in Section 7.

2.3 ENSO: What remains to be learned

The 1997–98 El Niño demonstrated again that we have realized significant skill in forecasting the evolution of the tropical Pacific at least 6 months ahead, especially once a warm event has begun. By the first months of 1997 (when strong westerly winds were already starting in the west) several models were suggesting that a moderate or larger El Niño was developing and would grow through the rest of the year. By mid-1997, the forecasts showed the evolution over the next 6 months quite realistically. This ability is a clear demonstration of the scientific successes made possible by a concentrated effort combining long-term observations, modeling, and assimilation techniques.

At the same time, even the best models had only limited skill in forecasting the rapid onset of the 1997 event much before the start of the initiating winds, and none predicted its extremely large amplitude. In conditions less dramatic than the growth stage of a large El Niño, our skill at predicting the smaller variations of tropical Pacific SST is relatively weak. We still have difficulty defining the conditions that presage El Niño, and do not know to what extent the oceanic state is a necessary element of its initiation, or whether the onset is perhaps initiated by essentially random wind patterns. Even now, 2 years after the peak of the 1997–98 event, with unprecedented observational documentation of the phenomena and numerous modeling efforts, we remain unable to definitively say why this event developed to be so strong.

Interpretations of the ENSO cycle have evolved from the early idea that the trade winds build up a strong zonal sea level gradient that provides the potential for an eastward flow when the winds relax (Wyrtki, 1975). Two ideas at the heart of subsequent El Niño oscillator models were first suggested by McCreary (1983): that ENSO timescales are due to a time-lagged negative feedback through reflection of Rossby waves from the western boundary, and that wave-mediated thermocline depth variations would affect east Pacific SST and then feed back to modify the strength of the trade winds. A weakness of the McCreary formulation was that a realistic timescale requires wind anomalies occurring quite far from the equator in order to generate sufficiently slow Rossby waves. Also, the eastward advance of El Niño is consistently observed to occur significantly slower than the propagation of Kelvin waves. Later models that demonstrated "delayed oscillator" physics (Zebiak and Cane, 1987; Suarez and Schopf, 1988; Battisti, 1988; Battisti and Hirst, 1989) proposed additional elements of the coupled interaction. A coupled instability develops in the central basin, in which a depressed thermocline produces anomalously warm SST due to upwelling acting on a weakened vertical temperature gradient, then the warm SST results in anomalous westerly winds over and to the west of the warm patch, which further lowers the thermocline and induces a growing anomaly. Kelvin waves carry the downwelling eastward, but the propagation of the coupled anomaly is slower than the free Kelvin speed. The delayed oscillator models showed that a slow (ENSO) timescale can be produced solely within the equatorial waveguide. One no longer expects a warm event to be triggered directly by Rossby-to-Kelvin-wave reflection; rather, in delayed oscillator models the reflected wave lowers the equatorial thermocline and sets up the coupled instability in mid-basin, which then grows with its own timescale, generates its own westerly winds, and produces additional Kelvin waves that reinforce the growing anomaly.

The delayed oscillator models had important successes in predicting El Niños, including forecasting the 1991–92 event almost a year in advance, but have shown limited ability to predict or explain the unusual series of warm events since then. In particular, there has been little evidence that Rossby wave reflection played a large part in the initiation of the mid-1990s warm events, and this absence is probably the reason for the lack of forecast success by delayed oscillator ideas (e.g., Boulanger and Menkes, 1999; McPhaden and Yu, 1999; Delcroix *et al.*, 1999).

While the termination phase of El Niños has in several cases occurred via the reflection of upwelling Rossby waves generated by equatorial westerlies during the height of the event (Wakata and Sarachik, 1991), other possibilities have emerged over the past few years. Weisberg and Wang (1997) suggested that these upwelling Rossby waves can cool west Pacific off-equatorial SST enough to produce anticyclone pairs and associated equatorial easterlies, which generate upwelling Kelvin waves without any western boundary reflection. Harrison and Vecchi (1999) pointed to the seasonal solar warming of SST south of the equator in southern summer, pulling the El Niño convection off the equator and weakening the westerly anomalies that support the flattened thermocline, and thereby ensuring that events would begin winding down in December–January as observed. Jin (1997b) suggested that the weakened easterlies during El Niño both flatten the slope of the thermocline and produce a general discharge of upper-layer water from the equatorial zone (through the Sverdrup balance $\beta v = \text{curl } \tau$ due to the wind stress curl pattern of equatorially trapped westerlies). The poleward discharge raises the thermocline (on a slower timescale than the waves that determine the thermocline slope) and thereby tends to cool the SST, terminating the event (the reverse occurs for cold events).

The 1997–98 El Niño highlighted the role played by higher frequency atmospheric fluctuations, and in particular the Madden-Julian Oscillation (MJO), during the onset and termination phases of the event (McPhaden, 1999; Takayabu *et al.*, 1999). All El Niños from the 1950s to the present have been associated with elevated levels of intraseasonal westerly surface wind forcing (Luther *et al.*, 1983; Verbickas, 1998). In each case, several episodes of westerly wind forcing lasting typically 1 to 3 weeks developed before and during the El Niño events. These winds were related to the MJO and other phenomena such as tropical cyclone formation and cold air outbreaks from higher latitudes. However, episodic wind forcing is not a sufficient condition for El Niños to occur, since such forcing is evident during non-El Niño years as well. It has been argued that episodic wind forcing is not even a necessary condition for the development of El Niños, since some coupled ocean-atmosphere models simulate ENSO-like variability without it. Nonetheless, recent model studies (e.g., Chen *et al.*, 1997; Moore and Kleeman, 1999) indicate that stochastic forcing can amplify and markedly alter the evolution of the ENSO cycle if it occurs on time and space scales to which the ocean is sensitive, and when background oceanic and atmospheric conditions are conducive to the rapid growth of random disturbances. Such studies have explicitly implicated the MJO in affecting the onset and intensity of the 1997–98 El Niño (Kessler and Kleeman, 1999). Likewise, though reflection of upwelling Rossby waves caused the thermocline to shoal in the eastern Pacific and set the stage for an end of the 1997–98 El Niño, it was a relatively sudden strengthening of the trade winds in May 1998 that produced sufficiently strong upwelling to initiate surface cooling. This sudden wind change may also have been related to the MJO (Takayabu *et al.*, 1999).

Other factors that may have influenced the evolution of the 1997–98 El Niño include the Pacific Decadal Oscillation (PDO) (Mantua *et al.*, 1997; Zhang *et al.*, 1997) which involves a basin-scale, decadally varying pattern of surface winds, air pressure, and ocean temperatures extending from the tropics to higher latitudes. The PDO has generally been in a warm phase since the mid-1970s, elevating temperatures in the tropical Pacific and affecting the background conditions on which ENSO events develop. Hypotheses for decadal timescale variations in the Pacific, and how they may interact with or affect ENSO, are just beginning to be developed (e.g., Latif and Barnett, 1996; Gu and Philander, 1997; Graham, 1994; Zhang *et al.*, 1997; White and Cayan, 1998; Kleeman *et al.*, 1999). Critical tests of these hypotheses are hampered by the inadequacy of data for defining large-scale ocean circulation, upper-ocean water mass variations, and air-sea fluxes.

These observations suggest that differences among El Niño events may be due in part to non-linear interactions between higher frequency weather variability like the MJO with lower frequency ocean-atmosphere dynamics. There are alternate explanations, however, for irregularity of ENSO events in terms of frequency, duration, and amplitude. For example, irregularity of the ENSO cycle has been interpreted in terms of chaos theory involving nonlinear interactions and subharmonic resonances of ENSO with the seasonal cycle (e.g., Tziperman *et al.*, 1994; Chang *et al.*, 1996). Others have argued that the equatorial Pacific is a stable linear system whose low-frequency modulation is little more than the realization of red noise and not susceptible to interpretation in terms of repeatable deterministic physics (Thompson and Battisti, 1999).

The two warmest years on record for global mean temperatures were 1998 and 1997, in that order. While the 1997–98 El Niño contributed to these record highs, global mean temperatures have trended up over the century. Along with this trend, there have been more El Niños than La Niñas since the mid-1970s, and the time between the "super El Niños" in 1982–83 and 1997–98 is only half the typical 30–40 year separation between super El Niños earlier in this century (Caron and O'Brien, 1998). Is El Niño related to centennial timescale warming? Is this warming likely to produce stronger and more frequent El Niños, as has been observed over the past 25 years (Trenberth and Hoar, 1996)? Only when the mechanisms of El Niño are better understood will we be able to answer these questions with reasonable certainty.

The potential interactions of ENSO with both faster intraseasonal processes and slower decadal and centennial phenomena explains the PBECS strategy. The processes of El Niño itself must be better understood, and models of it made more accurate, if the mechanisms of interaction are to be determined. At the same time, phenomena in other regions and having other timescales and other physics must be documented before their interaction with ENSO can be examined. Progress will lead to improved climate forecasts across the full range of timescales. The approach, however, cannot be as focused as the TOGA program because the range of phenomena, timescales, and regions must be so much greater. PBECS must focus on all aspects of climate variability that are likely to impact ENSO and its predictability. A priority should be enhancing the observational acuity of the ENSO Observing System to see the critical processes in the upper ocean and at the air-sea interface, expanding in situ observations to higher latitudes in order to better understand tropical/subtropical ocean interactions, and sustaining all these observations over many cycles of decadal variability. At the same time, we must learn to extract more from observations, past and present, by making better use of models and robust dynamical constraints to interpret them. Ways for doing this are discussed below.

2.4 How PBECS can improve climate prediction

ENSO represents the strongest interannual variability of the earth climate, and for this reason it has been the focus of the seasonal climate-forecasting effort. Although this effort has met with some measure of success, the recent experience of the 1997 El Niño has shown that there is considerable progress yet to be made (see Section 2.3). How might PBECS help improve seasonal forecast systems and contribute to that progress? First, PBECS can help improve the quality and quantity of data available to initialize the ocean before starting a coupled model forecast. Second, it can support the effort to improve the forcing fields that are also critical for initializing the ocean. Third, it can provide data for model validation and improvement. Lastly, PBECS can support process studies that lead to model improvements.

Different approaches have been taken in constructing seasonal forecast systems. Some simple models are intended to represent only a few dominant modes of the coupled atmosphere-ocean system. These systems run cheaply and their data requirements for initialization are modest. More complex systems, based on GCMs, demand more resources and more initialization data. However, it could be argued from recent experience that it was the added complexity of the GCMs that allowed them to capture more of the real-world variability, and to provide more reliable forecasts, of the 1997 El Niño than was possible with the simpler models. Until we understand how complex a model must be for seasonal forecasting, sufficient data must be available to initialize and to evaluate all systems. Sustained observations with resolution on the model's resolved scales are needed to detect and describe the physical processes that bring about climate change before we can know how complex a model must be to simulate those processes.

At first sight, it seems that the prediction models themselves might be the best way to determine which new observations are needed to improve initialization and validate model performance. This is, however, complicated because properly determining the skill of a system requires a large ensemble of forecasts. If the models are GCMs and the length of the forecast is hundreds of days, the cost of computing the ensemble can be daunting. Therefore, using forecast models to evaluate changes or additions to the observing system can quickly become impractical. Further, there are imperfections in the models, forcing fields, and assimilation systems so that using them to design observing systems involves some risk. A more sensible approach is to use common sense and empirical descriptions to ensure that sufficient observations are made where accurate initialization of forecasts is critical and where model performance is sensitive. A discussion of some particular types of data that might improve climate prediction follows.

2.4.1 Temperature

Subsurface temperature has traditionally been the primary data used to establish the initial conditions for operational seasonal forecasts. The present observing system provides daily observations from TAO moorings in the equatorial band, and basin-wide sampling largely by XBTs from Volunteer Observing Ships (VOS). This system generally illustrates what is needed, but with notable gaps, particularly in the Southern Hemisphere. Deployment of Argo in combination with certain VOS lines should correct those deficiencies and provide an unprecedented data set for assimilation and forecast initialization.

Accurate equatorial SSTs play a critical role in getting atmospheric convection right and therefore in making successful ENSO forecasts. SST is, thus, important both for establishing the ocean initial conditions and for evaluating the performance of the coupled model forecast. At present, climate-scale analyses of SST (weekly to monthly temporal resolution and 1° to 5° spatial resolution) differ by 0.1 to 0.2°C rms in the tropics. This kind of resolution and accuracy is probably sufficient for the present. PBECS might, however, examine the possible impact on SST analyses and model representation of upper-layer processes of increasing temporal and spatial resolution of SST. For example, increasing the rate of real-time data return from TAO moorings and drifting

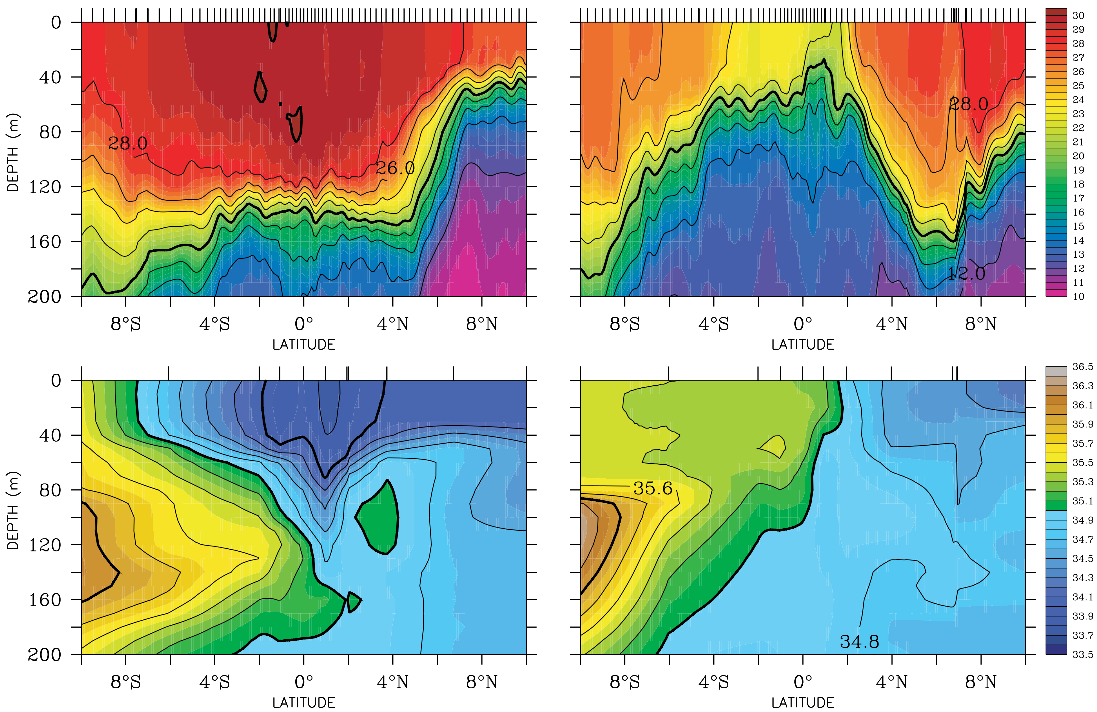

Figure 2.2: Temperature (top) and salinity (bottom) from XBT/XCTD transects crossing the equator at 140°W. The left panels are from December 1997 during the mature phase of El Niño, and the right panels are from August 1998 during La Niña conditions.

buoys to resolve the diurnal cycle may help both to understand processes and improve the quality of SST analyses.

2.4.2 Salinity

Subsurface salinity is not currently used to initialize operational forecasts, because not enough salinity observations are reported to constrain the assimilation system. However, recent work shows that variations in subsurface salinity may account for 5 to 10 cm in sea-surface height variability in the western and central equatorial Pacific (see Fig. 2.2). The evidence also suggests that the effects of salinity on sea-surface height may, in part, compensate for the effects of temperature. Salinity data from Argo and TAO moorings are needed to determine the importance of accurate salinity to the ENSO forecast problem.

2.4.3 Altimeter data

From the perspective of climate forecasting, the great value of altimeter data is their global coverage and high spatial and temporal resolution. This is particularly useful in minimizing aliasing of smaller scale features from the more widely sampled temperature data. Much work remains to be done to improve methods for assimilating altimeter data, particularly techniques for better projecting the observed variability in surface height onto the subsurface temperature and salinity

(Vossepoel *et al.*, 1999). Observations of salinity, discussed in the previous section, will provide important support for this work. Altimeter data at a resolution and accuracy matching the present standard of TOPEX/Poseidon will be required for the foreseeable future.

2.4.4 Forcing fields

In initializing a forecast model the forcing fields are as important as the data that are assimilated. Errors in the forcing cannot be entirely eliminated by assimilating data describing the state of the ocean because initialization is a complex product of the assimilated data, the forcing fields, the model and the assimilation technique. To improve initialization, better forcing fields are required. For example, different wind-stress products, when used to drive an ocean model without data assimilation, can produce differences in the depth of the 20°C isotherm on the equator that are comparable to the interannual signal. In situ data from moorings and ships will be necessary to improve the current operational flux and wind products that are used to initialize forecasts.

Improved surface forcing can also improve prediction skill through the better testing of model performance. Better forcing fields will allow detection of smaller model errors. Equally, by better partitioning buoyancy flux between heat and freshwater, improved flux fields will improve the utility of altimetric measurements of sea-level height in initialization and in model validation.

2.4.5 Data for validation

Velocity data are unlikely to be soon routinely assimilated for climate forecasting. Long time series of velocity data from moorings are rare (only five TAO moorings are equipped with current meters) while drifters and floats describe only one level, making effective assimilation difficult. Nevertheless, velocity data from moorings, floats, and drifters provide a valuable way to validate model performance. Additional observations would be useful. For example, at present the equatorial Pacific moorings are commonly used to tune ocean models, but given the uncertainties in the wind-stress and mixing parameterizations, this is not a particularly strong constraint on model behavior. In the future, current meters at other off-equatorial TAO locations, or well analyzed drifter and float fields, could provide a more complete picture of the tropical current system and more stringent tests of model performance.

2.4.6 Final remarks

The focus here has been on data for initializing seasonal forecasts. Changes in the background state that might affect the seasonal forecast must be part of the initial conditions. The initial conditions must be accurate enough to capture these changes, which may be of smaller amplitude than the interannual variability. Present forecasting experience does not allow different observations to be evaluated for utility in analyzing and/or predicting decadal variability, but the value of better description of changes in the background for ENSO cannot be overlooked.

Data that is to be useful in operational forecasting must be quality controlled and made available to the operational centers in near real time.

2.5 How PBECS and process experiments can improve coupled models

Coupled models of the climate system are used to explore, simulate, and predict climatic variability on timescales from months and longer. For prediction, the most familiar use of such models is forecasting El Niño and associated sea-surface temperature patterns in the tropical Pacific over periods of a few months to a year or so. These predicted patterns of tropical SST are then used with ensemble atmospheric-model runs to predict El Niño impact over land and at higher latitudes (Shukla *et al.*, 1999).

Beyond the prediction of El Niño, the main use of coupled circulation models has been exploration of other possible modes of ocean-atmospheric variability. Such modes of coupled variability depend on positive feedbacks, or delayed negative feedbacks, between ocean and atmosphere. Feedbacks with suitable phase lags can lead to the creation of distinct mode-like variability, although the character of the mode (oscillatory or damped) may depend on many parameters in the models. The work of Latif and Barnett (1996) gives an example of such a mode with a decadal timescale that exists in a coupled model integration. Whether or not such modes operate near these same parameter values in nature remains open to question, but the salient fact is that positive feedback between ocean and atmosphere can have a large influence on the nature of climate variability. Models that simulate these feedback processes in different ways can have widely different representations of long-term variability. The challenge for coupled modelers is to ascertain whether their models are operating in realistic parameter regimes, and how to improve them if they are not. For this work, the value of a strong and extended observational base is extremely valuable.

2.5.1 El Niño and its variability

El Niño prediction depends on coupled ocean-atmosphere models that obtain the bulk of their skill from the accurate observations of the subsurface ocean for initialization. Ji and Leetmaa (1997) show an example of such predictions and the role of subsurface observations on such predictions. We are currently trying to improve the simulation of El Niño for predictions over a year or so. It remains to be seen if the models successfully predicting ENSO are doing so for the right reasons. Coupled models for prediction of ENSO exist at many levels of detail. A common failing of coupled general circulation models is the difficulty in maintaining a proper “cold tongue” in the eastern Pacific, particularly between the coast of South America and about 110°W longitude. Here there is a strong interaction between the wind fields, the orography, the SST, and the currents that make it difficult to simulate the seasonal cycle, let alone its interannual fluctuation. A PBECS effort on the interactions in the eastern tropical Pacific offers perhaps the best strategy for improving this situation. This is discussed in Section 8.1.

While it may be argued that the existing TAO array and improving satellite observations are sufficient for the prediction of El Niño, recent efforts have begun to focus on the longer term variations in El Niño and its predictability. That is, we are coming to an understanding of whether El Niño is a stationary process, or whether significant changes can occur in the nature, strength, and predictability of El Niño. For these questions, the present observational network is weak, and coupled models are largely untested.

The variation in ENSO and its predictability has been simulated in a few coupled numerical models (Kirtman and Schopf, 1998; Rodgers *et al.*, 1999a). While these studies are intriguing, they are not definitive simulations of the climate on these timescales. More complex coupled models are being developed to simulate such effects, but their reliability is much in question. A key element of PBECS will be to maintain longer period ocean measurements to ascertain the ways the ocean actually changes and to assess the meaningfulness of coupled model results. For example, the Kirtman and Schopf study finds that ENSO variability and predictability in a hybrid coupled model are closely related, that both go through decadal-period changes, and that relatively small changes in the tilt of the thermocline and of the background winds can cause these large changes in ENSO. But the model these authors use contains a simplistic treatment of the subsurface temperature structure, and the reason for interdecadal change in ENSO variability is related largely to changes in only one of many mechanisms that can connect wind perturbations to SST changes.

More complicated coupled models may exhibit changes in the subsurface temperatures by different mechanics, and may have entirely different mechanisms for altering El Niño (Fedorov and Philander, 1999).

The proposals of Gu and Philander (1997), McCreary and Lu (1994), and others provide mechanisms whereby the El Niño dynamics may interact with higher latitudes through oceanic pathways. The Kirtman and Schopf study suggests that decadal variability can result from purely tropical sources, while the Barnett *et al.* (1999b) study finds a tropical response to mid-latitude atmospheric decadal modes. The wind changes identified by Barnett *et al.* are of the same order as those found in Kirtman and Schopf.

We are faced with many possible mechanisms for decadal scale changes in El Niño. Sorting them out will inevitably rely on a combination of coupled modeling studies, long-term ocean observations, and specific process experiments.

2.5.2 Decadal climate variations

To study long-term climate variations we may examine long coupled numerical model experiments or relatively short observational records. We have the unsatisfactory choice between a well-sampled process in which we do not have confidence or grossly undersampled nature itself. (For example, an observational view of the Latif and Barnett model result is given by Zhang and Levitus (1997), who resolved one cycle of variability in the North Pacific over 30 years of data. To conclude this is even a "canonical" event stretches credibility). While PBECS will significantly extend the observational base for climate, the program cannot advance the observational record faster than real time. The hope for PBECS is therefore that its observations will be used to significantly improve the models upon which we must rely for long-term statistics.

For this improvement, there are two strategies that should be followed: the first is to observe as much of the system as possible for as long as possible; the second is to conduct process studies to address specific unknowns in the model systems. The biggest improvements in coupled models will be focused improvements in the component models. Beyond this, the value of a long-term extensive array of measurements cannot be overstated. Experience with the TAO array has shown it to be an extremely valuable aid for improving ocean models used in El Niño prediction.

A focused process study of direct importance to coupled models on these longer timescales would be investigating the possibility for positive feedback between the SST and the surface heat fluxes in the northwestern Pacific. Positive feedback is an essential ingredient of the Latif and Barnett theory, as it provides the energy source that sustains the oscillation. These processes depend on an organized atmospheric response to SST anomalies through (a) cloud and radiative response, (b) evaporative feedback through wind speed changes, or (c) modification of storm tracks. These changes on climate timescales can be subtle and the region of interest is not well observed. Whether this positive feedback exists in nature is the first question. Many meteorologists doubt whether the atmospheric response to small changes in mid-latitude SST is "big enough" to be observable, even if it is large enough to affect climate on climatic timescales. The main feature of PBECS's approach, as discussed in Sections 5 and 7.2, is a combination of improving routine surface-flux estimates, using process experiments to provide the foundation for ocean heat budgets in the northwest Pacific, and possible increased surface meteorological measurements east of the Kuroshio-Oyashio Extension.

2.5.3 Surface fluxes

Better fluxes of momentum, heat, and water (latent heat) will assist in improving the ocean and atmosphere components of coupled models, and by making coupled-model diagnosis and verification more complete.

Of the fluxes important for improving ocean models, wind stress continues to be of primary interest. Further progress in model development is critically tied to improved surface stresses, a statement that is as true today as it has been for 20 years. Because of the ocean's sometimes nonlinear sensitivity to both wind mixing and the stress curl, resolution is as important as accuracy of large-scale averages. Near surface mixing involves relatively high-frequency turbulent exchange processes (perhaps including the diurnal cycle), so it will be helpful to characterize the temporal character of these fluxes. Fortunately, long-term improvements in surface stresses are being provided by sustained modern satellite programs. Ocean models are also ready to take advantage of significant improvements in surface heat flux components. Indeed the process of improving parameterizations of vertical and horizontal mixing and testing near surface advection are currently limited by these uncertainties. Mixed layer turbulence depends on buoyancy fluxes, not simply the heat flux, so knowledge of the surface freshwater flux is needed.

For improving atmospheric models, radiative fluxes and the air-sea exchange of water are most important because they provide a quantitative test of parameterizations of clouds and precipitation, and hence of the internal heating of the atmosphere. Indeed, many hypotheses about oceanic influences on the atmosphere involve exchanges of latent heat (see Section 5), the height in the atmosphere where that heat is released, how much heat (and hence rain) is released, and the effects of SST on atmospheric stability, cloud amount, and its impact on radiative heating in the atmosphere. Satellite observations of clouds and upwelling radiation plus surface observations of rainfall, evaporation, and radiative fluxes are all essential ingredients in developing models that correctly simulate these processes.

In improving coupled models, measured surface fluxes play an important, if circuitous, role. By the nature of coupled models, surface fluxes are internal model variables that are highly sensitive to small changes in either of the component models. It is not clear how to assimilate flux information into the coupled models to improve model physics and parameterizations. Too often, discrepancies in surface fluxes can be falsely attributed to problems in one component model, leading to counterproductive "fixes." On the other hand, since coupled models are inescapably limited by the quality of the component models, the role of surface fluxes in improving these models is extremely important.

3. Data Assimilation Modeling

A key objective of PBECS is to evaluate existing hypotheses about decadal modulation of ENSO and broad-scale decadal variability. The phenomena under investigation are dynamically complex, involving exchanges between the tropics and subtropics, atmosphere and ocean, mixed layer and thermocline, boundary currents and interior circulation, etc. The spatial and temporal scales of possible processes involved are broad: from episodic subduction under a storm to decadal communication between the tropics and subtropics. Realizable in situ observing systems will always be too sparse to directly observe all important processes, while satellites, although having broad sampling capabilities, are insufficient for testing hypotheses about climate phenomena without extra information to infer subsurface fields. Numerical models have proven to have skill in various aspects of the general circulation and are being used now on a routine basis for process-oriented studies. PBECS will extend the process-study methodology to climate by using models to bring together diverse sets of data and reliable dynamical constraints to form the best possible picture of the evolving climate system.

The use of models by themselves is limited by uncertainties in surface forcing fields, model parameters, and lateral boundary conditions (for regional models). Model errors due to coarse spatial resolution and incomplete/improper representations of physical properties provide another level of uncertainty. To achieve the PBECS science objectives, an optimal combination of ocean observations and numerical models is required and it can be achieved through rigorous data assimilation.

A primary goal of data assimilation is to use observations and numerical models to produce an estimate of the state of the ocean and atmosphere that is as close to reality as possible. To do this, the solutions of a dynamical model are brought into consistency with reality by using data and dynamical constraints. At the same time, the model estimates unobserved quantities and performs space-time interpolation/extrapolation.

A second goal of assimilation, perhaps the most important from the perspective of CLIVAR science, is to test the validity of the underlying dynamical model. The procedure attempts to minimize the misfit between model solution and observations plus the error in satisfying the model, each weighted by an a priori estimate of data and model errors. If the misfit cannot be reduced to the level of observational error, the model must be rejected.

In addition to providing validated products to elucidate climate phenomena and physical processes and to test the underlying model dynamics, assimilation supports the following: evaluation of the impact of a given data set in constraining the model state; the relative impacts of different data types; rational parameter estimation; estimating uncertainties in processes such as mixing; improving air-sea fluxes; identifying missing physics; and design of observing networks. For the latter, the adjoint model provides the unique ability to estimate the data distribution required to gain new insights into specific science questions, or to obtain a required accuracy in estimates, e.g., of heat transports. The associated sensitivity experiments are much cheaper than full ocean estimates and can be done before a full ocean estimate is available.

Typically, data assimilation is formulated as a least-squared problem in which a cost function of the following generic form is minimized subject to data and model dynamical constraints:

$$\begin{aligned} J = &\sum_t \Big[\mathbf{y}(t) - \mathbf{H}\Big(\mathbf{x}(t)\Big)\Big]^T \mathbf{R}^{-1} \Big[\mathbf{y}(t) - \mathbf{H}\Big(\mathbf{x}(t)\Big)\Big] \\ &+ \sum_t \Big[\mathbf{x}(t+1) - \mathbf{F}\Big(\mathbf{x}(t)\Big)\Big]^T \mathbf{Q}^{-1} \Big[\mathbf{x}(t+1) - \mathbf{F}\Big(\mathbf{x}(t)\Big)\Big] \end{aligned} \tag{1}$$

where $\mathbf{y}(t)$ are observations distributed in space and time, $\mathbf{x}(t)$ is the model state, $\mathbf{H}$ is an "observation matrix" that converts the model state to the equivalent of observations, $\mathbf{x}(t+1) = \mathbf{F}\big(\mathbf{x}(t)\big)$ is the model equation, and $\mathbf{R}$ and $\mathbf{Q}$ are a priori error covariance matrices for the data and model constraints, respectively. The final solution is essentially a weighted least-squared fit of the model to the data. Given the data and a model, the prescriptions in $\mathbf{R}$ and $\mathbf{Q}$ of a priori errors in the data and model constraints dictates the quality of the assimilation product.

Many methods have been devised to solve the above inverse problem. Reviews of various assimilation methods can be found in many sources (e.g., Anderson and Willebrand, 1989; Ghil and Malanotte-Rizzoli, 1991; Bennett, 1992; Wunsch, 1996; Robinson *et al.*, 1998). The more common "advanced" methods include the adjoint (i.e., four-dimensional variational, or 4D-Var), the sequential Kalman filter and related smoothers, representer, and Green's function methods. These methods differ in algorithm but are equivalent to each other so long as assumptions about data and model dynamical constraint errors (as they appear in (1)) are the same. Less sophisticated and computationally less expensive schemes include three-dimensional variational (3D-Var) and optimal interpolation where data and model constraints in the time dimension are not considered (i.e., missing the sum over t in (1)). Furthermore, there are simple methods such as nudging and data insertion methods. Instead of formally solving for a "best-fit" solution posed by (1), in these methods the model state is relaxed toward, or replaced by, data (similar to treating model errors in (1) as infinite).

3.1 Current status of data assimilation and prospects

Assimilation of in situ and/or satellite ocean data into ocean general circulation models (GCMs) using advanced methods has been demonstrated on basin to near-global scales (e.g., Yu and Malanotte-Rizzoli, 1996; Lee and Marotzke, 1997; Stammer *et al.*, 1997; Lee and Marotzke, 1998; Zhang and Marotzke, 1998; Fukumori *et al.*, 1999). Although prototypes, most of these studies explicitly account for uncertainties due to surface forcing and lateral boundary conditions (for regional studies) that account for a significant portion of the model error. Assimilation of oceanic and atmospheric data into intermediate coupled ocean-atmosphere models has also been attempted for the tropical Pacific (e.g., Bennett *et al.*, 1998; Lee *et al.*, 1999).

Recent innovations in estimation theory, combined with improvements in computational capabilities, have enabled application of optimal estimation methods to many data-assimilation applications. Various approximations have been put forth to reduce the computational requirements of the adjoint method (e.g., Courtier *et al.*, 1994) and Kalman filtering and smoothing (e.g., Fukumori and Malanotte-Rizzoli, 1995). Adjoint model compilers have been advanced (e.g., Giering and Kaminski, 1998) reducing the programming effort to generate model adjoint code. By combining enhanced observations, improved dynamical models, and data assimilation to address climate issues, PBECS will lead to (1) a much improved estimate of the state of the Pacific Ocean and (2) a quantitative evaluation of what processes have been resolved and what needs further improvement in terms of dynamical models, assimilation methods, and observing systems.

To address the data-assimilation needs of CLIVAR and the closely linked goals of the Global Ocean Data Assimilation Experiment (GODAE), a consortium has recently been formed under the U.S. National Ocean Partnership Program (NOPP) with funding from NSF, NASA, and ONR. The consortium includes scientists at Massachusetts Institute of Technology (MIT), Jet Propulsion Laboratory (JPL), and Scripps Institution of Oceanography (SIO), and is named "Estimation of the Circulation and Climate of the Ocean" (ECCO). The goal is to elevate global ocean state estimation from its current experimental status to a quasi-operational tool for climate research and prediction that will help to integrate the modeling and observational communities. The scientific goal is to describe and understand the global general circulation of the oceans and its role in climate by combining modern large-scale data sets with a state-of-the-art general circulation model (GCM). The central technical goal of ECCO is a complete global-scale ocean state estimation over the 15-year period 1985–2000 at the highest possible resolution and quality along with a complete error description.

By filling in for missing observations and using dynamics to correct sampling errors, assimilation produces complete fields of u, v, w, T, S, and p as well as mixing fluxes and parameters like eddy viscosities. These fields will support study of, for example, the connection between mid-latitude water mass formation areas and low-latitude shifts in stratification, ventilation process in specific regions, variations in transports and surface fluxes, the uptake of tracers and CO_2. A close interaction between ECCO and PBECS will help to refine those products. For specific process studies, fields can be sampled at an increased temporal sampling rate.

One factor hampering progress in ocean modeling and data assimilation is a weak link between the modeling community (including ocean state estimation) and observers, and between groups doing data assimilation and process-oriented modeling. Building the computational machinery for assimilation is the primary responsibility of such groups as ECCO. However, evaluation and improvement of the assimilation product, and the underlying dynamical model, is a challenging task that should involve assimilators, modelers, and observers. The tasks include:

- determining data and model errors and error statistics,
- preparing and quality controlling observations,
- pre-processing data for volume reduction (assimilation does not use satellite data at every pixel or current meter records at hourly intervals, but data reduction should not lose useful information),
- comparing model output with data and simple models to detect biases that may signal model deficiencies.

Given the diversity of observations for PBECS and the complexity of state-of-the-art dynamical models to be applied for PBECS, active inputs from observationalists and from modelers are indispensable to a successful data assimilation effort.

3.2 Major Issues of Data Assimilation

Although the ultimate goal of data assimilation is clear, there are many practical issues hindering its progress and success. The following are some of the major issues.

3.2.1 Determining a priori data and model errors

The most critical aspect in data assimilation is determining and understanding the error covariances of data and model constraints because they directly affect the outcome of assimilation. Errors in a model solution come from many sources. Some are associated with external factors such as uncertainties in the initial state, surface fluxes, and side (open) boundary conditions or bad parameters. Others are internal in nature such as inappropriate parameterizations (e.g., of mixing), errors resulting from coarse model resolution, or missing physics (e.g., mesoscale eddy fluxes). Assimilation can correct errors from external factors by treating the initial state, surface fluxes, open boundary conditions, or mixing coefficients as control variables, and solving for their optimal values. In principle, internal model errors must be corrected in the context of forward model improvement, although they can be masked by data assimilation. For instance, if the analysis is allowed to violate the underlying dynamical model, incorrect stratification due to an inappropriate vertical-diffusion parameterization can be repaired through assimilation of profile data.

Processes not represented by the model, and thus not correctable by data assimilation, are referred to as model representation error (after Lorenc, 1986). This type of error is included in the data-constraint error covariance $\mathbf{R}$ to avoid forcing the model to fit data it cannot represent. As such, the data-constraint errors differ from measurement error (sampling and instrumentation errors), which is usually much smaller than the representation error. Processes resolvable by the model but incorrectly appearing in a model solution due to uncertainties of external factors (e.g., forcing), sometimes called model process noise, is included in the model-constraint error $\mathbf{Q}$. The estimated state can then deviate from the prior model solution in such a way (as described by $\mathbf{Q}$) to account for uncertainties in external factors. Recognizing the sources of model errors and correctly attributing them in $\mathbf{R}$ and $\mathbf{Q}$ alleviates the risk of a biased estimation.

Estimating model errors is most challenging but even determining data-constraint uncertainties is often nontrivial. For example, the uncertainties of altimetric sea level data depend on errors associated with satellite orbit, tides, atmospheric corrections, and the inverted barometer effect. Efforts have been made (e.g., Fu *et al.*, 1993) to quantify the error covariance of altimeter data based on its reasonably good spatial-temporal sampling characteristics. For in situ data such as hydrography, the point-wise measurement error is usually smaller than that of satellite data. However, limited samples of in situ data are often used to infer the large-scale low-frequency state of the ocean through assimilation using non-eddy resolving models. In this context, transient signals such as eddies captured by the hydrographic casts are considered "noise" and a covariance of eddy structure is needed to develop the data-constraint error covariance. These eddy statistics are difficult to estimate from the limited sampling in space and time that is now available.

Knowing model errors depends critically on having sufficient data for evaluating model performance. Because much of the difficulty in estimating model errors is due to the lack of sufficient samples, the increase in observational density and in the ocean processes observed will not only improve assimilation products by increasing the data constraints, but also through better assessment of model errors.

3.2.2 Validation

An important aspect of data assimilation is complete evaluation of the quality of the assimilation product in order to identify inappropriate prescription of a priori errors or possible model weaknesses and to improve upon them. Similar to elementary linear regression, there are two conventional approaches to evaluating goodness of fit: (1) examine residual model-data misfit systematically for non-random patterns that might indicate model deficiencies or incorrect weights for data and model, and (2) cross-validation using independent data (e.g., withholding data for comparison with the assimilation product). For example, Fukumori *et al.* (1999) performed a careful and systematic analysis of the estimated state derived from the assimilation of TOPEX data into a coarse-resolution global ocean GCM. Data-assimilation techniques will not advance without such elaborate effort in evaluating the assimilation product.

3.2.3 Required computational resources

Currently, limited computational resources, both computational speed and memory requirements, are hindering progress in applying data assimilation to basin-scale ocean problems. This is particularly true when using complex models and advanced assimilation schemes. The following provides examples of typical data-assimilation computing requirements. On a 64-processor HP Exemplar-class cluster, it takes approximately 1200 CPU hours to perform a one-year global adjoint assimilation using the MIT OGCM with 1° by 1° resolution in the extratropics, telescoping to $1/3^\circ$ in the tropics, and 46 vertical levels (a total of 3.7×10^6 grid points). The ongoing global optimization at MIT/SIO over a 6-year time frame with 1° resolution requires about 300 CPU days on the NPACI/SDSC T90 (see Stammer and Wunsch, 1999). While some supercomputing centers have larger machines, most fail to meet assimilation requirements because of the competition with users not requiring as many resources, insufficient memory availability, and incompatible queuing schedules. A fundamental challenge for PBECS is to secure the computational resources its foundation of model-assimilation needs.

3.3 Model improvement

Ultimately, the PBECS goal is to develop better models to explain climate dynamics and to predict climate variability. How will the assimilation effort lead toward this? What modeling activities are needed to make the assimilation effort successful?

3.3.1 Model biases

Dynamical models are the foundation for data assimilation and, consequently, model weaknesses adversely impact the quality of model-assimilated products. One signature of this is model bias, which is non-random error in the assimilation analysis resulting from erroneous dynamical constraints in the underlying model. In meteorological forecasts, model bias often results in the rejection of good observations in order to avoid an unphysical assimilation solution resulting from inconsistency between the data and model physics. As discussed in Section 3.2, assimilation cannot correct for model errors such as those arising from missing physics, inappropriate parameterizations, or a lack of computational resolution. These errors can be compensated for with enough observations, but they can be corrected only by improving the underlying dynamical model. The cause for model bias is often difficult to determine because of the possible combined effects of incomplete physics, limited resolution, crude parameterizations, and inaccurate initial, forcing, and boundary data. The PBECS improvements in surface forcing fields and in the quality of observational fields suitable for model validation will significantly assist in detecting model errors and assimilation biases.

3.3.2 Initialization

Even assimilation models must be started from initial conditions. The NOAA World Ocean Atlas, often used for initialization, has well known shortcomings arising from the averaging used to reduce sampling noise and aliasing of interannual and interdecadal variability. In the upper ocean and in equatorial and coastal regions, oceanic adjustments rapidly bring about a quasi-balanced state between the model dynamics and forcing. But elsewhere (e.g., in and below the thermocline) oceanic adjustment times are decades to centuries, and errors in the initial conditions affect the model solutions for times longer than decadal scales of interest to PBECS. If in an assimilation study this adjustment is substantial, the model solution is unlikely to be suitable for studying climate. PBECS should improve the analyzed data for initializing models and assimilation studies.

3.3.3 Forcing

Even a “perfect” model will imperfectly reproduce the state of the ocean when its forcing is in error. In coastal and tropical regions, forcing data may be the dominant source of “model error.” Thus minimizing forcing error must be a priority for model-based studies of interannual to interdecadal phenomena. Until accurate forcing fields are available, it may well be more appropriate to use idealized parameterizations of air-sea fluxes because imposing substantially imperfect forcing fields on models can have a wide range of misleading and undesirable outcomes. PBECS will benefit from surface-flux reference sites and efforts to blend high-quality surface information and satellite surface forcing into operational meteorology products.

3.3.4 Model parameterizations

Coupled models are inherently predictive models. Their success depends on how well they work in the absence of data. The interaction between observations, assimilation products, and model development will be important for making strides in parameterizations for prediction. Experience with NCEP coupled-model forecasting underscores how important to improving parameterizations is the assimilation product. It is probably used more for comparing and validating models than for prediction. While there are inherent dangers in such reliance on assimilation products, improving and extending the analyses to high latitudes, as envisioned by ECCO, coupled with the increased observational basis of PBECS, will have a major impact on parameterization improvement for coupled prognostic models. This will be particularly true if analysis and observations of the surface mixed layer and upper thermocline can lead to more robust parameterizations. Exploration of parameterizations must proceed inside assimilation analyses and in forward model experimentation, and it is most likely to be successful if it proceeds phenomenon by phenomenon.

3.3.5 Resolution

Perhaps the biggest model choice affecting parameterizations and bias is that of spatial resolution. The coarser the model resolution the more processes that must be parameterized and, often, the more evident is the impact of the particular choice of parameterization. It would be desirable to run at the highest feasible spatial resolution were it not for the need for the solutions to reach quasi-equilibrium relative to the phenomena of interest, the need to carry out multiple simulations in order to explore sensitivity to model parameterizations, initial conditions and forcing fields, and, in the case of assimilation, the need for extensive calculations to reduce the cost function (1). These practicalities limit many types of multi-decadal model studies to intermediate resolution. There is hope that the next generation of massively parallel computers will greatly increase the computer power available, and permit more realistic resolution, but until then parameterizations will remain critical.

3.3.6 Equatorial problems

Modeling the coupled behavior of the equatorial waveguides poses special challenges. Observations show tremendous shear in the upper ocean. Very high resolution (10 m or less in a level-model sense) is the minimum to resolve the shear near the surface and near the equator. To the extent that surface-advective processes affect SST, models with less resolution will inappropriately diminish the importance of such processes. Special efforts to observe the vertical structure of near surface currents will provide a critical evaluation of tropical ocean model skill (see Section 4.4). To the extent that simulating SST change requires accurately modeling the mixed layer within the waveguide, near-surface vertical velocity field also must be correctly modeled, and penetration of solar radiation and vertical mixing across the base of the equatorial mixed layer will have to be parameterized. In addition to process experiments (see Sections 9.1 and 9.2), accurate forcing fields must be available to support this type of mixed-layer modeling.

Modeling the equatorial thermocline is also a challenge. All existing ocean models have substantial biases at least somewhere between the surface and the thermocline. Some have unrealistic stratification; others place the thermocline at unrealistic depths; many do not well reproduce the few observed profiles of currents between the thermocline and the surface. Achieving the PBECS objectives for ENSO and its decadal modulation require realistic modeling of the equatorial thermocline. A development project to make the needed improvements is needed.

Modeling near-equatorial, large-scale atmospheric dynamics in response to SLP, SST, and convection also poses special challenges. Very few data sets exist with good vertical resolution and the long duration to permit evaluation of model skill within the atmospheric waveguide (roughly within 10° of the equator). Two parts of the PBECS plan for sustained atmospheric observations described in Section 10 would be particularly useful. Special observations, perhaps including enhanced profile data from the near-Dateline islands of the equatorial Pacific, could provide critical information for improving atmospheric models of ENSO. An ITCZ atmospheric observing effort would also be very helpful.

3.3.7 Summary

The high resolution view of the interior ocean to be provided by Argo and new PBECS time series of boundary current transports will stress the simulation ability of today's dynamical models. There are now sufficiently realistic ocean models to carry out serious forward modeling exploration in support of PBECS goals. Improved surface flux fields will make it possible to evaluate the suitability of the existing atmospheric models in coupled models. We know there are undesirable features in the models of both fluids that must be remedied through model improvement that must be a PBECS activity.

4. Sustained Ocean Observations

The discussion in Sections 2 and 3 gives general guidance on what phenomena and processes PBECS must observe. This, coupled with empirical and model information on scales, defines what measurements PBECS needs. The most promising feedback processes for explaining decadal modulation of ENSO and decadal variability like the PDO involve generation of oceanic anomalies in one location, movement of this anomaly by oceanic processes of advection, wave dynamics or eddy transport, and the subsequent impact of this anomaly on air-sea fluxes at a new location. A detailed array design of the PBECS observations is beyond this plan, but general guidelines can be set out and the sections below will expand on these. The numbers here are meant only as approximate guides to the magnitude of the problem.

Anomalies can be generated in the mixed layer by buoyancy fluxes (heat or fresh water) or wind stirring. They can also be generated anywhere in the upper ocean by Ekman pumping or anomalous advection. The variables that must be observed, and the scales to be resolved, to diagnose anomaly generation are:

- air-sea fluxes of heat, water and momentum—$\delta x \sim 300$ km, $\delta t \sim 12$ hour
- profiles of temperature and salinity—$\delta x \sim 300$ km, $\delta t \sim 10$ day, $\delta z \sim 5$ m
- near surface currents—$\delta x \sim 300$ km, $\delta t \sim 10$ day

The resolutions given are estimates of the measurement density required to avoid aliasing and provide enough observations to resolve the signal of interest in the face of typical smaller scale variability. For air-sea fluxes, it is likely that, because atmospheric spatial scales and ocean timescales are large, observations on a much lower spatial density coupled with models could provide the needed surface flux density.

The strategy for understanding the propagation of subsurface oceanic anomalies will involve observing two quantities: lateral fluxes and changes in internal ocean structure. Different sampling requirements pertain in the interior ocean, in boundary currents, and in the equatorial waveguide. In the interior, observation of anomalous structure requires

- profiles of temperature and salinity—$\delta x \sim 300$ km, $\delta t \sim 10$ day, $\delta z \sim 5$ m, $z_{\max} \sim 1500$ m

Resolution of S as well as T is essential to determining the extent to which anomalies are dynamically active or density compensated. In boundary currents and frontal regions, resolution should be increased in the cross-front direction. Measuring lateral fluxes is only practical along selected transects using

- profiles of temperature and salinity—$\delta x \sim 10$–50 km, $\delta t \sim 1$ month, $\delta z \sim 5$ m, $z_{\max} \sim 1500$ m

where the sampling density varies from O(10 km) in western boundaries to O(50 km) in the interior.

In the equatorial waveguide TOGA relied on tracking anomalies from changes of ocean structure, mainly using temperature measurements from the TAO array. To diagnose the impact of large fresh water anomalies this needs to be extended to include salinity:

- profiles of temperature and salinity—$\delta x \sim 1000$ km, $\delta y \sim 100$ km, $\delta t \sim 5$ day, $\delta z \sim 5$ m, $z_{\max} \sim 500$ m

A few direct velocity observations are now made from TAO. These should be expanded to better characterize advective transport, particularly above the thermocline.

Over much of the ocean, observing the impact of oceanic anomalies on the atmosphere requires the same kinds of observations outlined above for the generation of anomalies. Along the equator, however, upwelling and the impact of very high near-surface gradients and shears additionally requires

- near-surface velocity observations—$\delta x \sim 1000$ km, $\delta y \sim 100$ km, $\delta t \sim 5$ day

Many of the ocean observations needed for PBECS are already in place, and others are being developed by other programs and nations. Central to the ocean observing system is a suite of satellite sensors that provide regular and global coverage. In situ observations cannot approach satellite coverage, but in situ observations will improve the accuracy of satellite measurements, like SST, or extend observations to fields not measured from space, like ocean temperature and salinity profiles to describe the vertical structure associated with the sea-surface height (SSH) anomalies observed by altimeters.

The backbone of the PBECS in situ measurements will be the ENSO Observing System and the array of profiling floats from Argo. An international consortium of agencies and scientists is implementing Argo to provide temperature and salinity profiles in the upper 2 km with a design resolution near 300 km and 10 days. The ENSO Observing System includes the TAO/TRITON array, which provides high-temporal-resolution observations of winds and oceanic thermal structure at about 70 locations, with ocean velocity, salinity, and additional surface meteorological measurements at a smaller number of sites. The ENSO observing system also includes an array of low-resolution XBT sections, surface drifters that observe SST and surface currents, and an array of tide gauges to supplement altimetry.

PBECS cannot be successful without the maintenance of the satellite, Argo, and ENSO observing systems for 15 years, and it is likely that the information provided by these systems will be needed indefinitely.

In addition to the satellite, Argo, and ENSO observing systems, PBECS will depend on other widespread ocean observations. This section describes the major observational elements from which PBECS will develop its ocean climate observing network. Presenting each element separately, and separate from the framework of assimilation that will integrate the data, is necessary for clarity, but it masks how PBECS must be a single integrated system simultaneously addressing all the phenomena discussed in Section 2. Its goal is to describe all the ocean changes that are important to climate on seasonal to decadal timescales, including the transports and fluxes that bring them about.

4.1 Satellite Observations of the Ocean

Space-based observations will play an important role in PBECS by providing high spatial resolution, broad spatial coverage, and frequent sampling that compliment in situ observations. The satellite-based ocean measurements that will be most useful are sea-surface height (SSH), surface wind stress, and sea-surface temperature (SST). The interannual and longer timescales of variability that are the focus of PBECS require continuous, uninterrupted observations of these variables over the full duration of the program.

Maintaining satellite programs for the 15-year PBECS observational period and beyond poses serious practical difficulties. Long, continuous, and consistent time series—exceeding the typical 3–5 year lifetimes of individual satellite missions or instruments—are extremely difficult to acquire outside of operational programs. As summarized below, radiometers for measuring SST have already been incorporated into the National Polar-orbiting Operational Environmental Satellite System (NPOESS), which is the future U.S. operational satellite program. The PBECS needs for measurements of SST thus appear to be met, with a possible gap in the data record in the middle of the PBECS program. The needs for SSH and surface wind stress will likely be met during the first half of the PBECS program with altimeters and scatterometers that are expected to be launched over the next 5 years on research satellites. After that, however, observations of SSH and surface winds available for PBECS are less certain since measurements of SSH and surface wind stress have not yet been incorporated into NPOESS.

Present and future satellites for measurements of SSH, surface wind stress and SST are summarized below.

4.1.1 Sea-surface height

Altimeter measurements of SSH include important climate signals such as steric heating of the upper ocean and large-scale, low-frequency variability of upper-ocean geostrophic currents. Satellite altimeter measurements provide unparalleled space-time coverage and accuracy of the long, baroclinic Rossby waves associated with the low-frequency variability that is of interest in PBECS. Altimetry will contribute importantly to observing the generation and propagation of oceanic anomalies.

Satellite SSH measurements are an important source of observations for the data-assimilation models to be developed as a part of PBECS. Since surface currents are generally coupled to subsurface variability through relatively simple vertical modal structures, sea-level variations contain information about fluid motion deep in the ocean interior. The combination of SSH observations with in situ measurements of profiles of density and velocity imposes essential constraints on the data assimilation models.

A timeline of altimeter missions is shown in Fig. 4.1. The TOPEX/POSEIDON and ERS-2 altimeters are now in operation. As shown in Fig. 4.2, the sampling patterns of these two satellites are very different. The repeat periods are 10 days for TOPEX/POSEIDON and 35 days for ERS-2. These two sampling patterns illustrate the tradeoff between space and time sampling. The shorter repeat period of the TOPEX/POSEIDON orbit reduces temporal aliasing of mesoscale variability, but results in a coarse ground track pattern. The longer repeat period of ERS-2 results in more closely spaced ground tracks but with larger temporal aliasing.

Future altimeter missions are shown as dotted lines in Fig. 4.1. The design lifetimes of altimeters are typically 3 years but, as the present 7-year TOPEX/POSEIDON data record shows, altimeters can last much longer than 3 years. The joint NASA/CNES Jason-1 mission and European Space Agency Envisat mission have been formally approved for launch during 2000. Jason-1 will sample the TOPEX/POSEIDON ground track and Envisat will sample the ERS-1 and ERS-2 ground

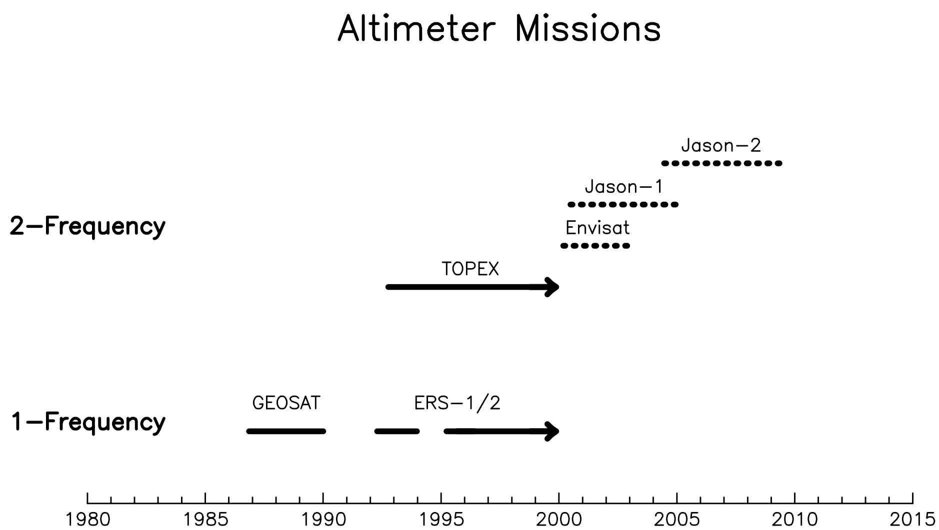

Figure 4.1: Timeline of altimeter missions. Arrowheads indicate altimeters that are in operation as of November 1999. The exact launch dates of future altimeter missions (dotted lines) are subject to change. The Jason-2 mission has not yet been formally approved by NASA.

tracks. Follow-ons to these two altimeter missions have been proposed but have not yet been approved. Of particular interest is the Jason-2 altimeter that would sample the TOPEX/POSEIDON and Jason-1 ground tracks. If there is overlap between each of these three successive altimeter missions, an uninterrupted record of highly accurate SSH along the present TOPEX/POSEIDON ground track will be available from September 1992 through at least the first half of the PBECS observational program.

While supportive of a collaboration with CNES to launch Jason-2, NASA has not yet formally committed funding for Jason-2. NASA has, however, clearly indicated its intention to transition altimetry to the U.S. NPOESS operational satellite program before 2010. The presently planned NPOESS configuration cannot support highly precise radar altimetry. The future of the very accurate satellite altimetry required for PBECS is thus uncertain beyond Jason-1 and is very uncertain beyond Jason-2.

4.1.2 Surface wind stress

Surface wind analyses from the ECMWF operational weather forecast model are thought to be the best source of surface wind forcing presently available for ocean circulation modeling. The modeling implications of the limited spatial resolution of the ECMWF surface wind analyses have not yet been fully explored. Although the grid spacing of the ECMWF model is 1°, the scales that are actually resolved are much longer. As shown in Fig. 4.3, the spectral energy in the ECMWF surface wind fields drops steeply at wavelengths shorter than about 700 km. Wavenumber spectra from aircraft observations suggest that the approximate log-linear spectral rolloff at the longer wavelengths resolved in ECMWF surface wind analyses also extends down to wavelengths at least as short as 100 km (see dashed line in Fig. 4.3). The ~700 km resolution of operational weather analyses is especially a problem in the tropics and near eastern boundaries where winds vary

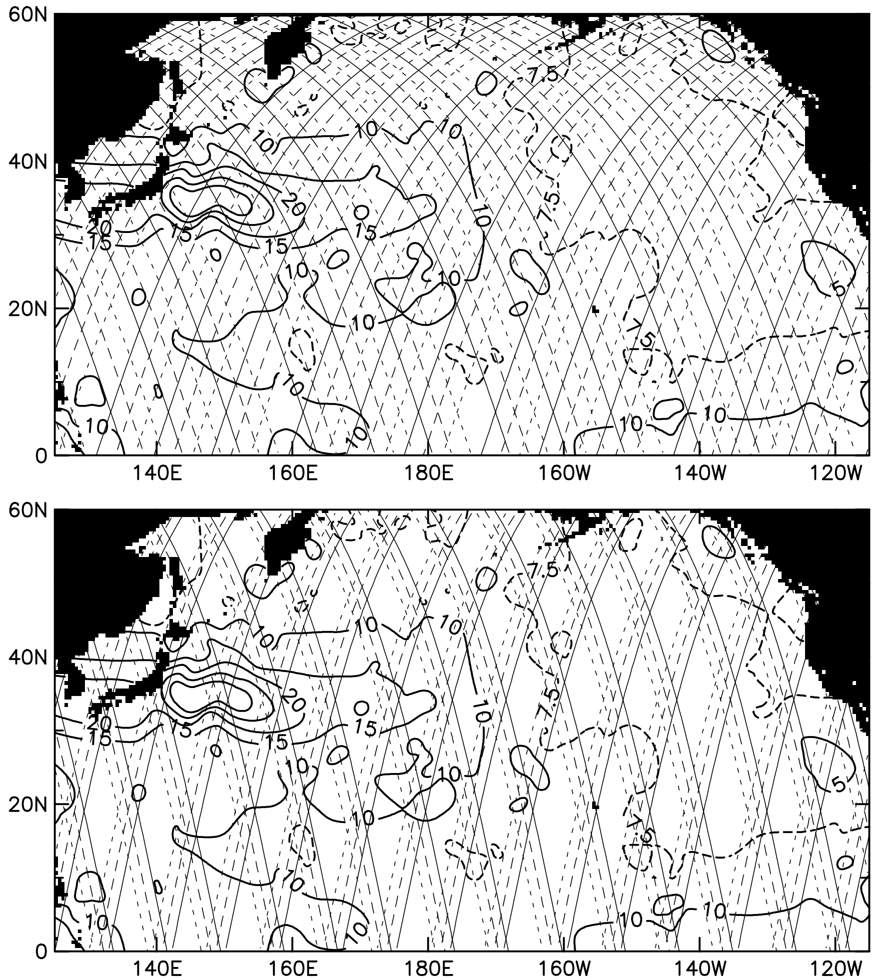

Figure 4.2: Ground tracks for the first 9 days of the TOPEX/POSEIDON 10-day repeat period (upper), and the first 9 days of the 35-day repeat period for ERS-2 (lower). In both panels, the ground tracks are overlaid on the standard deviation of SSH (in cm) computed from 6 years of TOPEX/POSEIDON data. The solid, dashed, and dotted lines correspond, respectively, to days 1–3, days 4–6, and days 7–9 of the orbital repeat periods. Note the subcycle in both orbit configurations that shifts 2.8° eastward every 3 days for TOPEX/POSEIDON, and 1.4° westward every 3 days for ERS-2. The ground track for the first half of the 35-day ERS cycle has 1.4° spacing as shown. During the second half of the repeat cycle, the ground track is shifted to interleave the grid from the first 17.5 days, leading to a 0.7° spacing for the full 35-day repeat cycle.

laterally over scales much shorter than 700 km, generating narrow bands of strong wind-stress curl that are poorly resolved by operational analyses.

At present, the only satellite measurements of surface wind stress are from a microwave radar scatterometer. The potential importance of scatterometry can be inferred from Fig. 4.4, which shows the vector squared correlation between NSCAT measurements of 10-m winds and ECMWF analyzed 10-m wind fields. Globally, the lowest correlations are found in the tropical band 20°S to 20°N. The correlations are especially poor in the eastern tropical Pacific and Atlantic. The regions of low correlation extend from the tropics poleward along the eastern boundaries in all three oceans.

With the development of realistic boundary-layer parameterizations in numerical atmospheric general-circulation models, the operational meteorological community has found that assimilation of scatterometer surface vector-wind measurements can yield improved operational weather forecasts. This has generated considerable interest in real-time assimilation of scatterometer data into the numerical weather prediction models. At present, the assimilation schemes are relatively simple. The development of more sophisticated assimilation schemes should improve the impact of scatterometer data on the forecast skills of the models. This, in turn, will improve the accuracies of the analyzed surface heat-flux fields derived from operational weather forecast models. These heat-flux fields are likely to play an important role in PBECS.

A timeline of scatterometer missions is shown in Fig. 4.5. The QuikSCAT and ERS-2 scatterometers are now in operation. Their sampling patterns are very different (Fig. 4.6); the spatial coverage of the wide-swath QuikSCAT scatterometer is more than three times that of the ERS-2 scatterometer. This greatly improves the accuracies of wind fields constructed from QuikSCAT data (Fig. 4.7). Future scatterometer missions are shown by dotted lines in Fig. 4.5. The design lifetimes of radar scatterometers are typically 3 years. The NASA SeaWinds scatterometer to be launched in 2001 on ADEOS-II and the European Space Agency ASCAT scatterometer to be launched in 2003 are both formally approved missions.

The SeaWinds scatterometer is identical to QuikSCAT. As shown in Fig. 4.6, the sampling coverage of the dual-swath ASCAT scatterometer is about half that of QuikSCAT and SeaWinds. The errors in wind fields constructed from ASCAT data are thus proportionately larger (Fig. 4.7). The launch of SeaWinds will provide coverage simultaneous with QuikSCAT, dramatically improving the space-time resolution and accuracy of wind fields constructed from the tandem scatterometers (Fig. 4.7). The AlphaSCAT follow-on to SeaWinds has been proposed to NASA. As with the Jason-2 altimeter, NASA is supportive of collaboration with the Japanese space agency to launch AlphaSCAT but has not yet formally committed funding for the proposed mission. As in the case of altimetry, NASA has clearly indicated its intention to transition satellite measurements of surface vector winds to NPOESS operational satellites before 2010. The presently planned NPOESS configuration cannot accommodate radar scatterometry.

As a substitute for scatterometry, NPOESS is pursuing a low-cost, low-power passive polarimetric microwave radiometer technique for estimating vector winds. While this technique shows great promise, it is not yet proven. If development continues on schedule, the U.S. Navy will launch a polarimetric radiometer in 2003 on the Coriolis/WINDSAT satellite. Until the spaceborne capabilities of this technology have been demonstrated, the future of highly accurate and densely sampled satellite observations of surface vector winds will remain uncertain beyond SeaWinds on ADEOS-II and highly uncertain beyond AlphaSCAT.

4.1.3 Sea-surface temperature

Sea-surface temperature (SST) is a key variable for the studies of ocean-atmosphere interaction that are a primary focus of PBECS. It is the surface boundary condition for the overlying atmosphere.

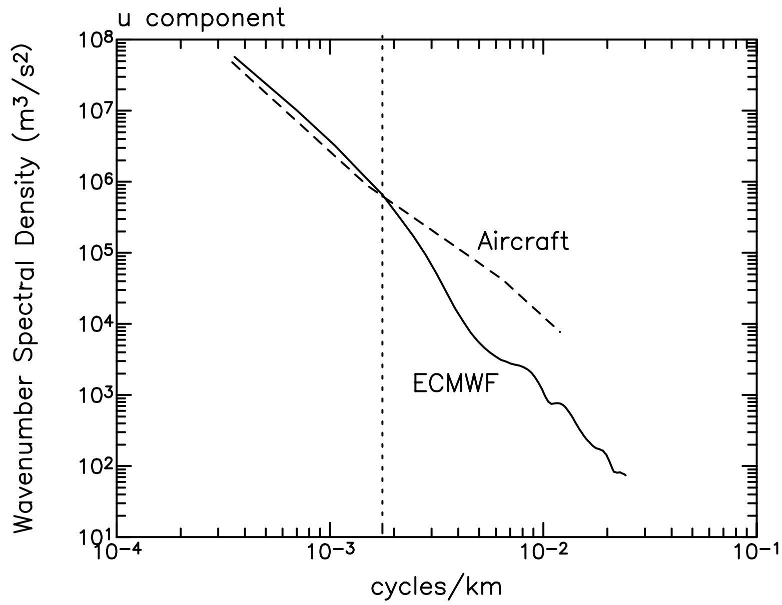

Figure 4.3: Wavenumber spectra of the zonal wind component from ECMWF analyses of 10-m winds (solid line) and of aircraft observations (dashed lines, from Nastrom and Gage, 1985). The vertical dotted line corresponds to a wavelength of 700 km.

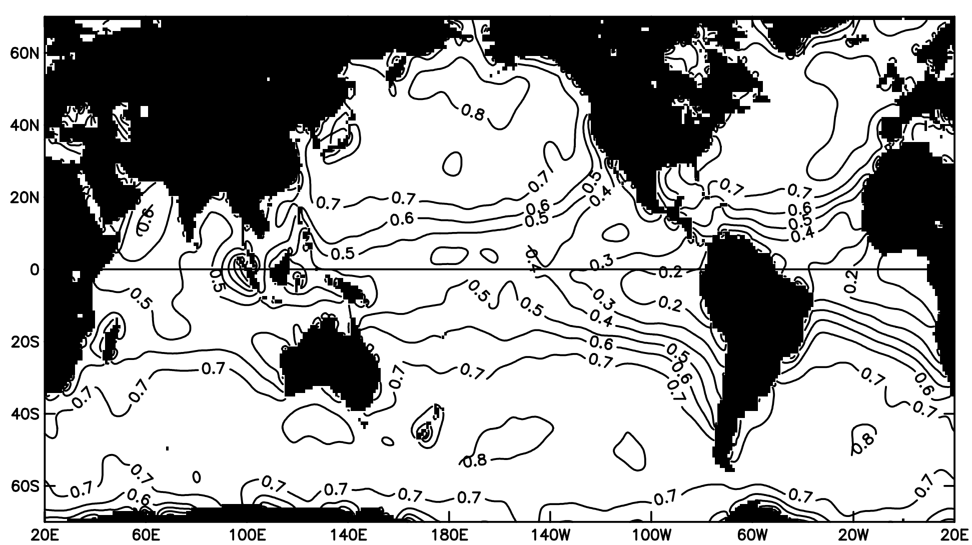

Figure 4.4: The vector squared-correlation (normalized to have a value between 0 and 1) between NSCAT vector winds and ECMWF analyzed 10-m vector wind fields interpolated to the times and locations of NSCAT observations (Crosby *et al.*, 1993).

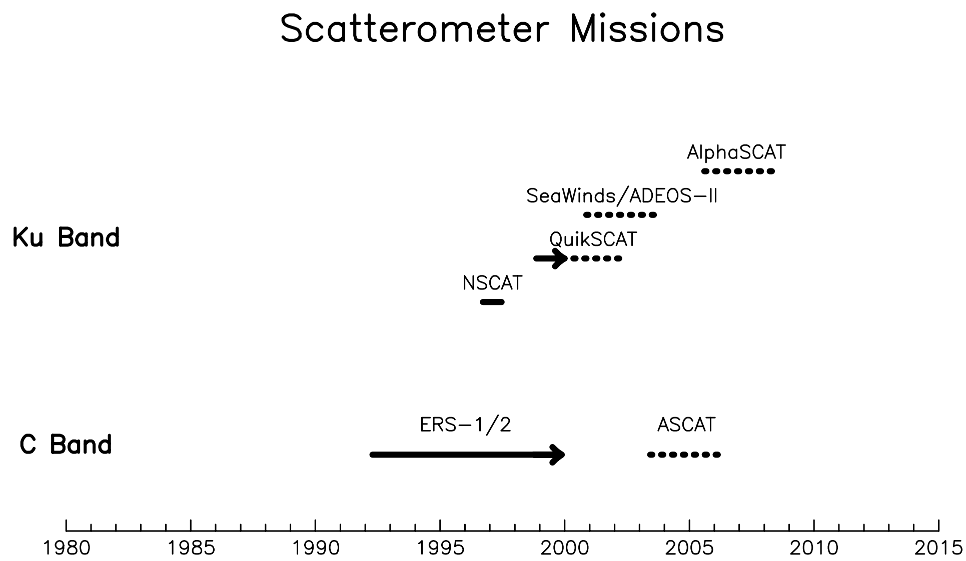

Figure 4.5: Timeline of recent past, present, and future scatterometer missions. Arrowheads indicate scatterometers that are in operation as of November 1999. The exact launch dates of future scatterometer missions (dotted lines) are subject to change and the AlphaSCAT mission has not yet been formally approved by NASA.

It is also the basis for indices of climatic variability of the ocean-atmosphere system. Satellite observations of SST offer much more complete space-time coverage than ships and buoys, but to achieve the accuracy required for climate studies, remote data must be blended with in situ measurements to deal with calibration problems resulting from environmental factors like clouds and dust.

Satellite measurements of SST have been available since 1973 from the infrared radiometers on the NOAA operational satellites, with high-quality SST estimates from the Advanced Very High Resolution Radiometer (AVHRR) available since 1979. An AVHRR is included on each of the two NOAA Polar Orbiters that are generally operational at any given time. There are also infrared radiometers on the NOAA Geostationary Operational Environmental Satellites (GOES) and on several European and Japanese satellites. Infrared radiometers are able to measure SST only in cloud-free conditions, which is a significant limitation since the sea surface is obscured by clouds about 60% of the time in the tropics and more than 75% of the time at middle and higher latitudes (Hahn *et al.*, 1995). Undetected clouds are one of the largest sources of error in infrared estimates of SST. In addition, stratospheric aerosols from major volcanic events, for example, have a significant impact on the accuracy of SST retrievals from infrared radiometers (Robock, 1989; Reynolds, 1993) as do a number of other biases (Reynolds *et al.*, 1989). Infrared measurements of SST will be available operationally throughout the PBECS observational program.

Many of the problems inherent in infrared measurements of SST can be overcome with passive microwave remote sensing at low frequencies (6–11 GHz) where, in rain-free conditions, the atmosphere is nearly transparent. Rain-contaminated observations are easily identified (Wentz and Spencer, 1998). A time line of microwave SST missions is shown in Fig. 4.8. The TRMM Microwave Imager (TMI) that is presently operating on the Tropical Rainfall Measuring Mission (TRMM) has

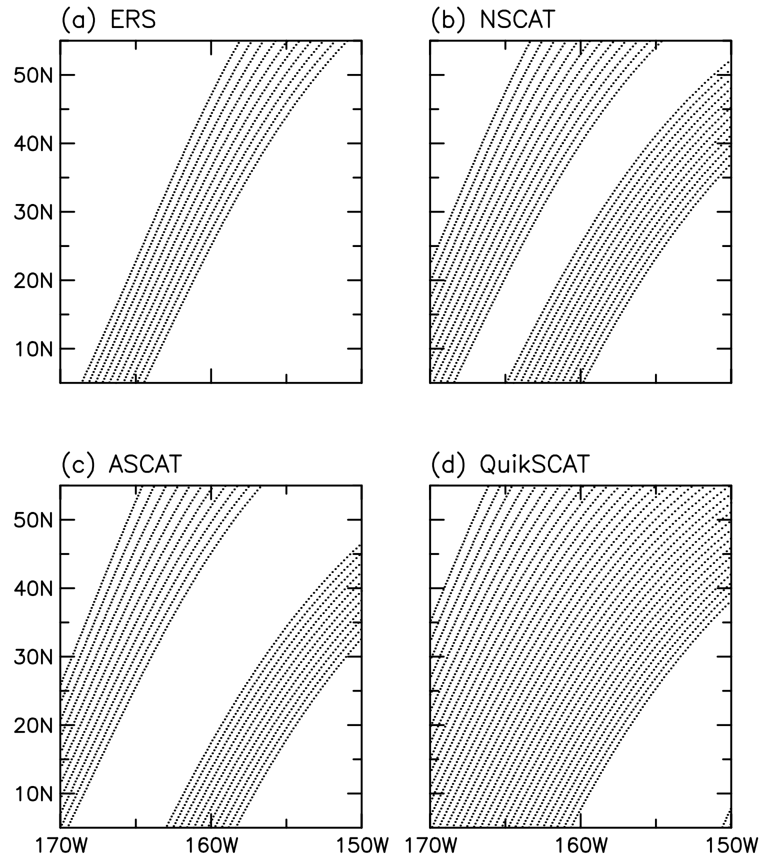

Figure 4.6: The sampling swaths of recent past, present, and future scatterometers.

been measuring SST since December 1997 with a spatial resolution of 46 km and an rms accuracy of 0.5°C (Wentz, 1998).

Future microwave SST missions are shown by dotted lines in Fig. 4.8. The Advanced Microwave Scanning Radiometer (AMSR) that will be launched in the near future on the EOS-PM platform and the ADEOS-II satellite includes a new low-frequency channel that will provide SST retrievals with even greater accuracy than are being obtained from the TMI. Thereafter, microwave measurements of SST will be available operationally from the Conical Microwave Imager/Sounder (CMIS) that will be included in the U.S. NPOESS operational satellite program. Depending on the exact launch dates of the various satellites and the duration of the AMSR data records from EOS-PM and ADEOS-II, there may be a gap in the microwave SST data record part way through the PBECS observational program (see Fig. 4.8).

4.1.4 Future U.S. operational satellites

The lead time to incorporate a new satellite sensor into the U.S. operational satellite system is of the order of a decade or longer. A coordinated long-range plan must therefore be developed at the earliest possible opportunity to assure continuity of the oceanographic satellite sensors for climate observations. In response to this need and to budgetary constraints, a new program called the National Polar-orbiting Operational Environmental Satellite System (NPOESS) is under de-

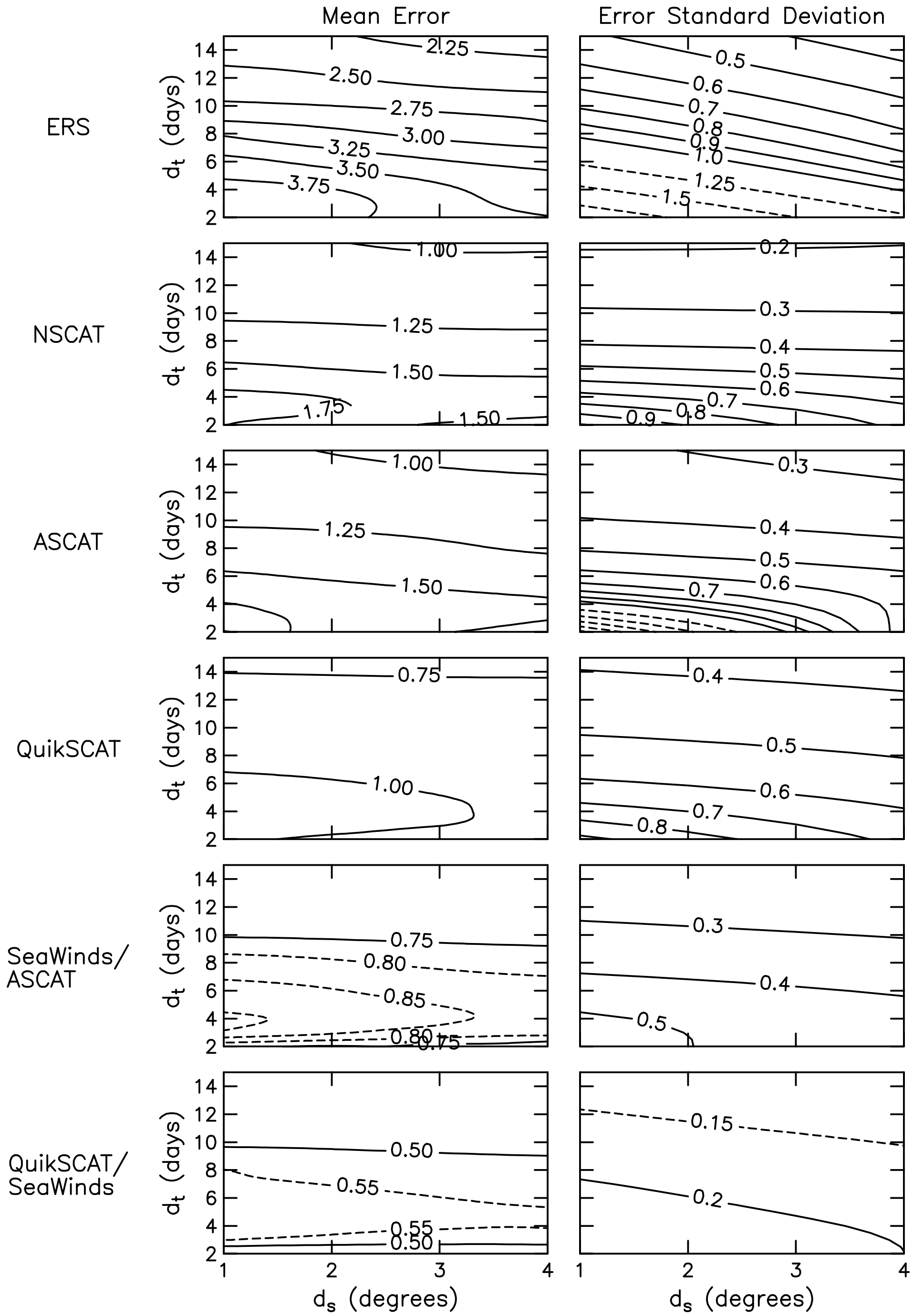

Figure 4.7: The mean (left panels) and standard deviation (right panels) of sampling errors in wind component fields constructed from various single and tandem scatterometer datasets, shown as a function of the half-power points *ds* and *dt* of the (isotropic) spatial and temporal filter transfer functions, respectively. For reference, the half-power points of a 2° by 7-day block average, for example, are about 3° and 12 days. The means and standard deviations in this figure were computed from 1° grids along 10°N, 25°N, 40°N, and 50°N at daily intervals over a 20-day period based on spatial and temporal autocorrelation functions estimated from NSCAT data.

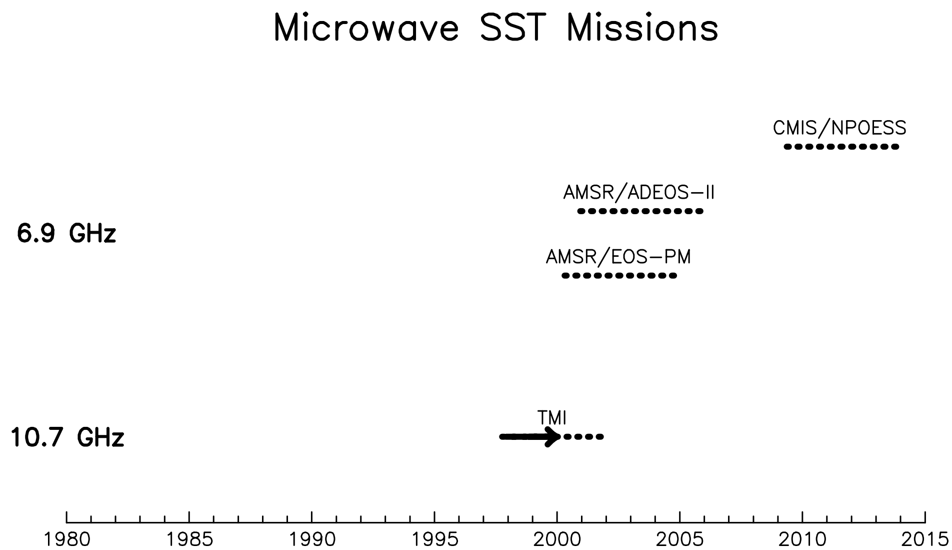

Figure 4.8: Timeline of recent past, present, and future microwave radiometer missions with low frequencies of 10.7 GHz (horizontal and vertical polarization required) and 6.9 GHz that are capable of measuring SST. Arrowheads indicate that the TRMM Microwave Imager (TMI) is in operation as of November 1999. The exact launch dates of future radiometers (dotted lines) are subject to change.

velopment. The mission of NPOESS is to provide a convergence of the NOAA, NASA and Defense Meteorological Satellite Program (DMSP) satellites into a single program to acquire, receive, and disseminate global and regional environmental satellite data. Where appropriate, one of the goals of NPOESS is to transition technology from the NASA research satellite program to an operational status. NPOESS-1 is expected to launch in 2009, followed by NPOESS-2 in 2010. The plan thereafter is to maintain two polar-orbiting NPOESS satellites in orbit at all times. Much of the instrument compliment on NPOESS will consist of the present instruments useful for operational weather forecasting. As noted above, the CMIS for microwave measurements of SST is included on NPOESS-1. SSH and surface vector winds have not yet been given high priority in the NPOESS planning. Unless instruments for measuring these variables with the accuracies and sampling needed are adopted by NPOESS, the satellite records of SSH and surface wind are in jeopardy of falling short of the full duration of PBECS.

The NPOESS program and its relationship with NASA Earth Sciences missions are presently under review by the National Research Council Committee on Earth Studies. The outcome of the NPOESS development process will play a crucial role in the success or failure of satellite-data acquisition and analysis efforts during the second half of the PBECS 15-year observational period.

4.2 Argo

The Argo profiling float array will be the central element of the in situ ocean observations for climate. PBECS cannot be successful without sustained basin-wide temperature and salinity profile data. Argo provides an interior array of temperature, salinity, and velocity measurements spanning the entire PBECS domain (Fig. 4.9). From the Aleutians to 40°S and across the full width of the

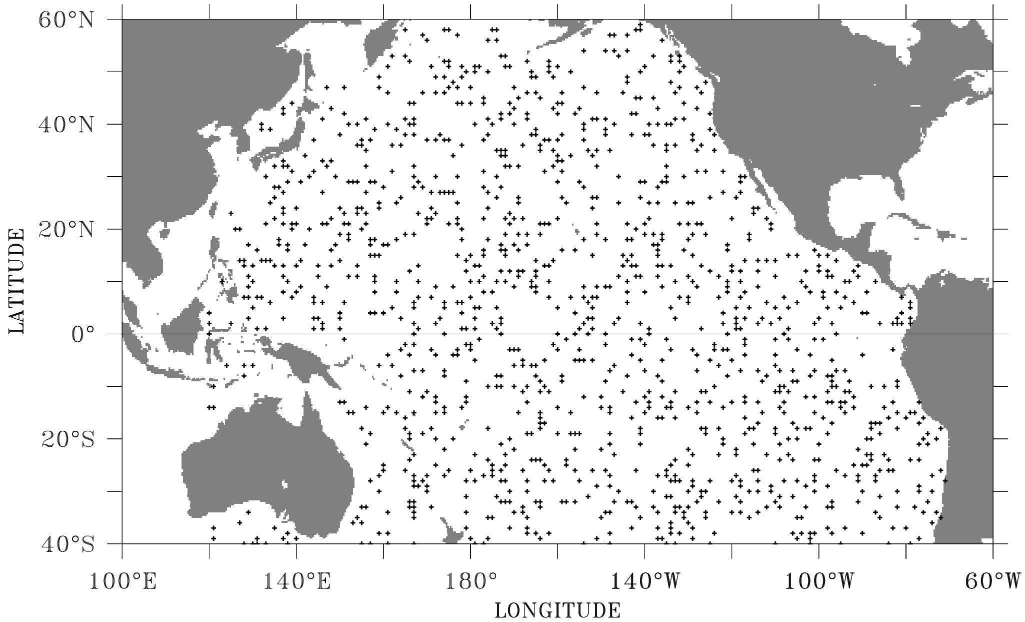

Figure 4.9: A random array of 1270 locations in the Pacific Ocean north of 40°S simulating the Argo array in the PBECS domain.

Pacific, over 1200 floats will profile every 10 days from about 2000 m to the sea surface. They will provide information in near real time for data analysis, model initialization, and assimilation. The float array will tie together the Pacific observing system, providing the large-scale context for regional measurements.

A key aspect of the float array is its complementary relationship to satellite altimetry. Profiling floats and altimetry together provide a dynamically complete description of sea-surface height and its subsurface expression. Anomalies in sea-surface height may be written as:

$$h' = \frac{\alpha}{g} p_{REF} + \frac{1}{g} \int_{p_{REF}}^{0} \alpha'(p) dp + \text{Errors} \tag{2}$$

where h is sea surface height, α is specific volume (a function of temperature, salinity, and pressure), p is pressure and primes denote anomalies from the temporal mean. The second term on the right (dynamic height) is calculated directly from float profile data. The first term on the right is obtained from the float's drift velocity (e.g., Davis, 1998) assuming a geostrophic balance. The left-hand side is measured by the altimeter. Hence, on the large spatial scales common to the float and altimetric measurements, the combined system accounts for both the density-related and reference-pressure contributions to sea-level variability. Models assimilating altimetric height alone cannot yet accurately describe this decomposition and the depth-distribution of the density signal. The ocean's dynamics and evolution depend critically on vertical structure, so the subsurface array is a necessary part of the total observing system.

Previous studies (e.g., Gilson *et al.*, 1998) show high coherence between altimetric data and dynamic height in the tropics, indicating dominance of baroclinic variability at low latitudes. This high coherence can be exploited to estimate subsurface density structure from altimetric height.

Thus, the high along-track spatial resolution of the altimeter can be used to interpolate between more widely spaced subsurface density profiles. However, even in the tropics, there are significant residuals in sea-surface height due to reference-pressure variability. These residuals increase with latitude (e.g., McCarthy *et al.*, 1999), and the reference pressure contributions are large at high latitudes. The need to measure all the terms in equation (2) is strongly emphasized in an ocean-scale observing system. Moreover, temperature alone is not sufficient to adequately estimate dynamic height. Salinity variability is a significant contributor to sea-surface height (e.g., Gilson *et al.*, 1998) in addition to providing primary diagnostic information for the freshwater budget.

Aside from its complementary relationship to altimetry, the profiling float array will also provide fundamental information for PBECS on a stand-alone basis:

Pathways, circulation, and transport. Floats that drift at thermocline levels will provide direct observations of water parcel trajectories and associated water mass properties. Knowledge of thermocline pathways and of the contribution of interior circulation to heat and freshwater advection are central to PBECS. The floats provide both direct observations of velocity at a reference level and geostrophic shear estimates.

Heat and freshwater storage. Closure of the oceanic heat and freshwater budgets in PBECS will require measurements of advection, storage, and air-sea exchange. Storage of heat and freshwater in the coupled climate system is dominated by the oceanic component. The float array is the only element of the PBECS observation strategy for measuring large-scale heat and freshwater storage. A primary design criterion for Argo (below) is to obtain sufficient independent estimates of storage on climate-relevant large spatial scales to average over noise due to mesoscale eddies and small-scale features.

The design of the Argo sampling array is an ongoing exercise in balancing requirements against the practical limitations imposed by technology and resources. A complicating factor is that the statistics of ocean variability are poorly known in many regions, making array design a necessarily iterative process. The design of an interior ocean array has been considered from a variety of different angles (Argo Science Team, 1998, 1999). The findings most appropriate to PBECS are:

1. Previous and ongoing float studies: A 5-year deployment of about 300 floats in the tropical and South Pacific Ocean (Davis, 1998) was found to be adequate for mapping the mean geostrophic pressure field at mid-depth, but not its time variability. The domain included nearly half the global ocean. These results showed that a large increase in float density was needed for mapping of time varying fields. Recent experience with profiling floats in the North Atlantic at much higher spatial density emphasizes this finding.

2. The existing upper-ocean thermal network: Numerous network design studies have been carried out, using the XBT data sets to provide necessary statistics, as summarized by White (1995). In approximate terms, an array with spacing of a few hundred kilometers is sufficient to determine heat storage in the surface layer with an accuracy of 10 W/m^2 on seasonal timescales and over areas 1000 km on a side. This improves to about 3 W/m^2 for interannual fluctuations. Combination of profile and altimetric data can improve this even further.

3. The altimetric data set: Spectral analysis of altimetric data shows that over the globe half of the variance in sea level is at wavelengths shorter than 1000 km (Wunsch and Stammer, 1995). If the climate signal of interest is defined to include all wavelengths longer than 1000 km, then an array with 500-km spacing provides a 1:1 signal-to-noise ratio. Spacing of 250 km would improve this ratio by more than a factor of 3. The unresolved variability—fronts, mesoscale eddies, etc.—has a short decorrelation time, typically 10–20 days, compared with the seasonal and longer climate signals of interest. Therefore, temporal averaging can increase

the signal-to-noise ratio. As a function of latitude, the half-power point varies from 1300-km wavelength in the tropics to 700 km at 50°N (Stammer, 1997).

4. Requirements for assimilating models: Initially, these are not distinctly different from the requirements for pure data analysis. The models require sufficient data to determine the statistics linking point measurements to the smoothed fields of the simulations. They also require sufficient data to estimate comparison fields for rigorous testing of model results. While there is still much to be done to determine the optimal data sets for assimilation, it is clear that the requirements remain substantial.

The initial plan for Argo calls for floats at approximately 3° spacing in latitude and longitude (Fig. 4.9). It is appropriate that PBECS adopt the same target since the requirements discussed above were established with PBECS objectives in mind. The 3° specification implies an increase in float density with latitude—but one somewhat less than called for by altimetric wavenumber spectra. The resulting greater signal-to-noise ratio in tropical latitudes is justified by the known importance of the tropics in ENSO predictability. The southern boundary of the PBECS float array should be 40°S, corresponding to the poleward edge of the subtropical gyre. The phenomenology of PBECS requires similar high density sampling of tropical and subtropical circulation, heat, and salt on both sides of the equator.

All Argo data will be publicly available as quickly as is technically practical. The Argo data system will provide near-real-time data via the GTS, with automated quality control, for operational users and forecast centers. A data set with scientific quality control, including expert examination of individual profiles and sequences, comparison to ancillary data sets and climatologies, and best recalibration of salinity data, will be available 3 months from collection.

Argo has strong international support and participation. Partners in implementation of the Pacific array include Japan, Canada, and Australia. Investment by our partners is contingent on a strong U.S. float program. Commitment of U.S. resources will provide powerful leverage for a Pacific-wide observing system.

4.3 High resolution temperature-salinity transects

The High Resolution XBT (HRX) program provides boundary-to-boundary profiling along selected lines, with closely spaced XBTs to resolve the spatial structure of mesoscale eddies, fronts, and boundary currents. The present set of regularly sampled Pacific HRX lines is shown in Fig. 4.10. Probe spacing is typically 10 km in boundary regions and 50 km in the ocean interior. Most profiles go to 800 m and XCTDs measure salinity at approximately 5% of the stations. Time series of HRX lines are as long as 13 years in the case of PX6 (Auckland-Suva), with most lines spanning the TOPEX era. The repetition frequency is about four times per year.

Within PBECS, HRX lines provide an eddy-resolving subsurface complement to the profiling float array and altimetric data sets. The float array is designed for broad-scale sampling-observation of heat and freshwater storage averaged over large ocean areas, plus large-scale components of the circulation. However, accurate determination of oceanic transport of heat, freshwater, and thermocline water masses requires that boundary current transport and interior eddy transport also be observed. The HRX transects provide the needed eddy-resolving measurements along a few lines, providing geostrophic and Ekman-transport variability on seasonal and longer timescales. With respect to altimetry, the HRX lines can reveal the subsurface structure of mesoscale and boundary-scale features not observed by the float array.

Examples of how the HRX transects address PBECS objectives are provided by recent analyses of the Taiwan to San Francisco line (Gilson *et al.*, 1998; Roemmich and Gilson, 1999; Roemmich

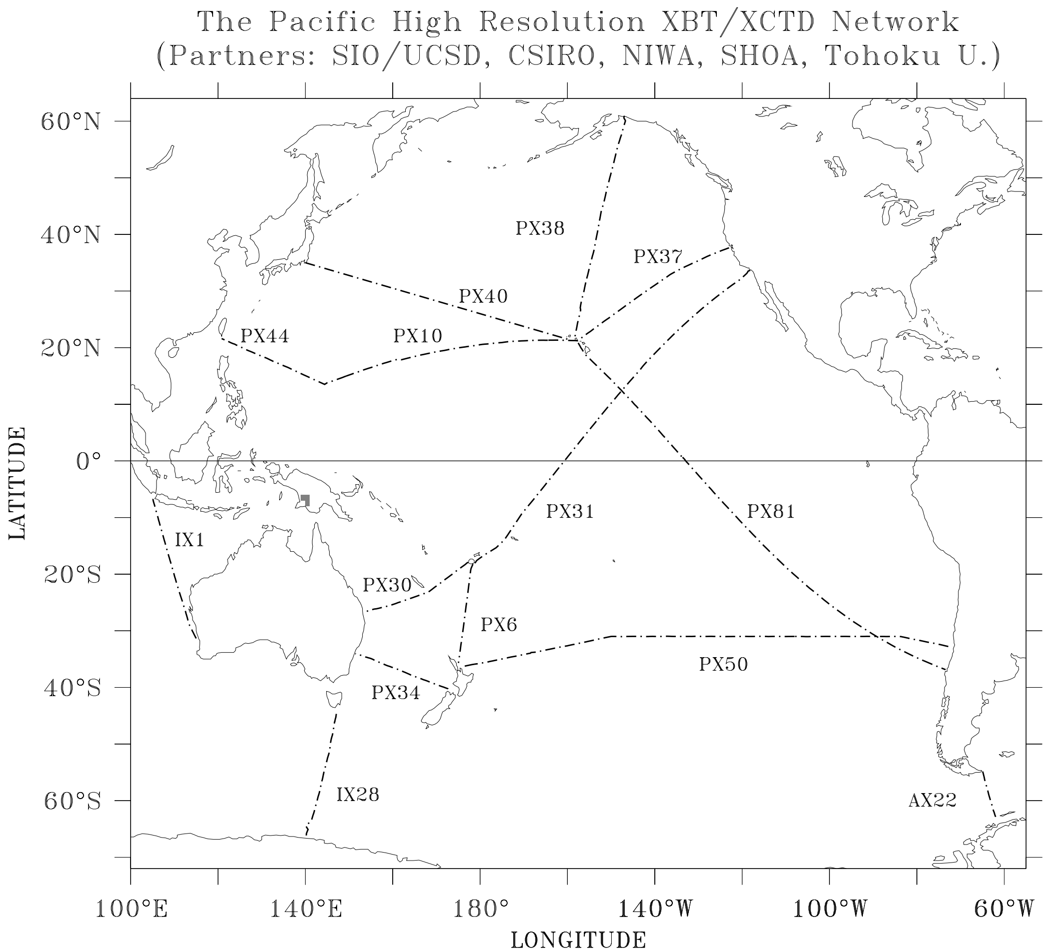

Figure 4.10: Presently sampled routes in the Pacific High Resolution XBT/XCTD network.

et al., 1999). A typical HRX temperature section between Taiwan and San Francisco is shown in Fig. 4.11. The northward Kuroshio is seen in the isotherms sweeping upward at the western boundary. A sequence of large mesoscale eddies is seen near the dateline. In averages over several hundred eddies between 1992 and 1998, Roemmich and Gilson (1999) found that the net effect of the eddies is to significantly enhance the shallow meridional overturning cell. The eddies transport surface water northward and thermocline water southward as a result of a systematic tilt of the eddies toward the west with decreasing depth. Eddy heat transport ($v'T'$) amounts to 0.1 pW in the mean. Total ocean heat transport is 0.76 pW, but varies interannually from about 0.5 to 1.0 pW (Roemmich *et al.*, 1999). The variability is due to changes in the northward Ekman transport that are approximately balanced (interannually) by changes in southward net geostrophic transport, plus changes in the number and strength of eddies. Basin-wide effects of eddy-scale phenomena must be taken into account in PBECS and are not addressed by other elements of the observing system.

The existing HRX transects, carried out on an international basis, are shown in Fig. 4.10. They include:

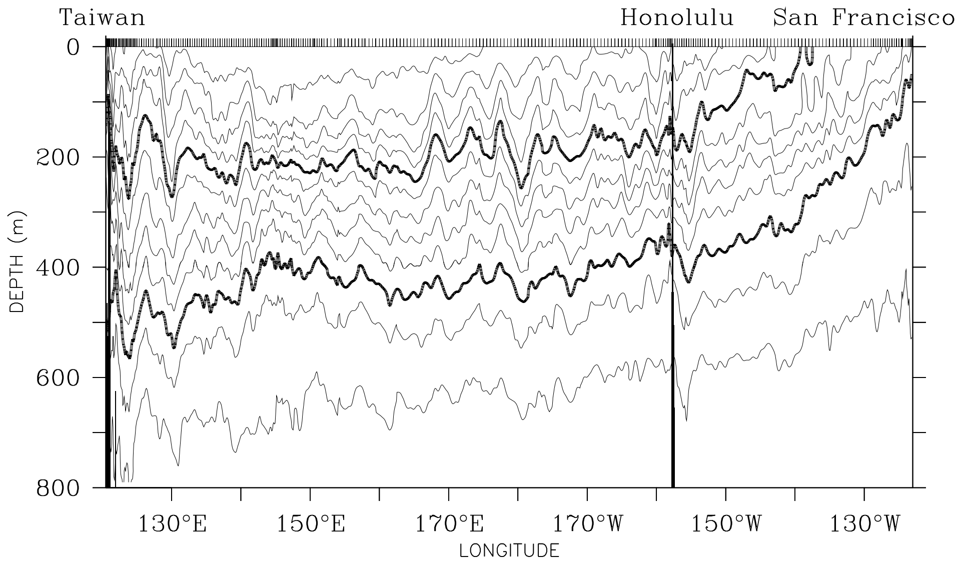

Figure 4.11: Temperature section during May 1996 along PX37/10/44, San Francisco to Taiwan. Contour interval is 2°C, with bold lines at 10° and 20°C.

1. zonal crossings of the North Pacific (PX37/10/44) and South Pacific (PX50/34), with meridional transport objectives as discussed above,
2. meridional lines in the central and eastern Pacific (PX6/31, PX81/38) to observe the structure and fluctuations of the zonal tropical current systems,
3. choke point transects across all entrances and exits to the Pacific—including the Indonesian throughflow (IX1), the Southern Ocean south of Australia (IX28), and Drake Passage (AX22),
4. additional western boundary current crossings (PX30, PX40).

Other transects identified as high priority for PBECS include meridional lines in the far eastern Pacific (e.g., Peru–California) and western Pacific (e.g., Noumea–Japan), and a zonal line at lower latitude in the South Pacific. A review of the present XBT networks was recently carried out as part of the international design of an ocean-observing system for climate (Smith *et al.*, 1999). It was concluded that, as the Argo float array is implemented, the XBT sampling should transition from area-based to line-based modes, specifically to accomplish the line-based objectives noted above. Such a transition will align the network with PBECS needs by providing the supplemental eddy- and boundary-scale information at a number of locations. The redesign and redeployment of the XBT networks should be influenced and optimized for PBECS objectives.

In addition to routine XBT and XCTD profiling, the HRX ships are also used for ancillary measurement programs and to test new instrumentation. For example, a full suite of IMET meteorological sensors is now installed on one ship with a second set scheduled for next year. The IMET program identifies systematic errors in traditional meteorological observations and in operational products including air-sea fluxes and wind stress. Other ancillary programs include sea surface salinity, atmospheric and sea surface CO_2, and deployment of profiling floats and surface drifters. New instruments now under development or testing include an improved XCTD, recoverable 200-m CTD, and electronic XBTs (to 2000 m).

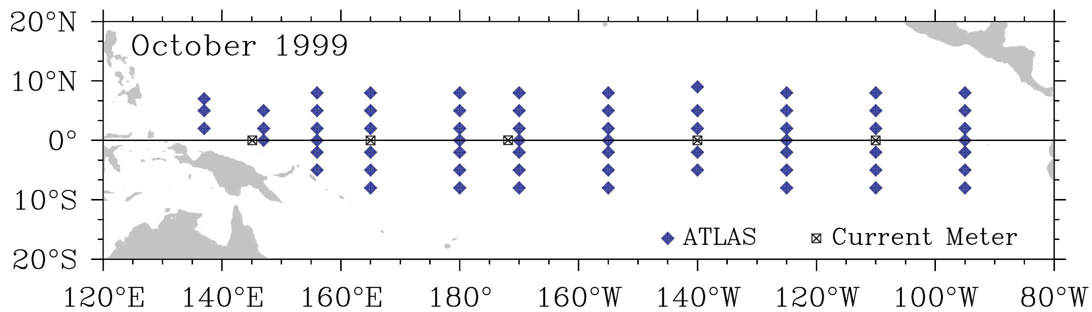

Figure 4.12: The TAO array.

4.4 The TAO array

The Tropical Atmosphere Ocean (TAO) array consists of approximately 70 ATLAS and current meter moorings in the equatorial Pacific between 8°N and 8°S, 95°W and 165°E (Fig. 4.12; McPhaden *et al.*, 1998). Surface winds, air temperature, relative humidity, SST, and ocean temperatures in the upper 500 m are measured at all locations. Five sites along the equator are instrumented for upper-ocean velocity profiles. At several sites, surface salinity, rainfall, and shortwave radiation are also measured. Data are transmitted to shore in real time via Service Argos. They are distributed via the GTS to operational meteorological and oceanographic centers around the world. Beginning in 2000, the western portion of the TAO array (137–156°E) will be instrumented with Japanese TRITON moorings. Quality controlled TAO/TRITON data will be available to the research community as a seamless data stream without restriction via the World Wide Web, as have TAO data in the past (`http://www.pmel.noaa.gov/toga-tao/`).

The TAO/TRITON array will be the central element in the PBECS observational network in the equatorial Pacific. This array measures the key variables necessary for monitoring, detecting, understanding, and forecasting year-to-year climate swings associated with the ENSO cycle. Some TAO moored time series are approaching 20 years in length, which makes them suitable for studies of decadal variability as well. Sampling is typically at 10-min intervals, so that high frequency variability can be averaged rather than aliased. Hour averages of surface meteorology and daily averages of subsurface temperatures are telemetered via satellite and relayed over the GTS. High temporal resolution is important in the equatorial wave guide where energetic ocean dynamic and thermodynamic evolution occur on timescales of days to weeks.

The following enhancements to the TAO array could be implemented for PBECS. The technologies are proven, pilot demonstrations have been completed or are underway, and the costs are incremental compared to the overall costs of operating the array. What is required in each case is the commitment to such new measurements as part of a sustained observational program.

4.4.1 Salinity

The western and central equatorial Pacific are characterized by large interannual variations in surface and subsurface salinity (Delcroix *et al.*, 1998; Kessler, 1999). Analyses of CTDs show that lack of salinity observations can sometimes lead to errors in dynamic height that are comparable in size to the ENSO signal in the western Pacific (see also Fig. 2.2). Errors in dynamic height affect the pressure field and heat storage inferred from altimetry, and if uncorrected lead to errors in assimilation analyses and initial conditions for coupled ocean-atmosphere models forecasting ENSO (Ji *et al.*, 1999). Salinity data are also useful for understanding how the hydrologic cycle over the ocean couples to the heat balance of the surface layer via formation of salt stratified barrier layers (cf. Lukas and Lindstrom, 1991; Ando and McPhaden, 1997). In addition, salinity is a valuable tracer of the meridional overturning circulation in the tropical and subtropical oceans.

Moored salinity measurements have been made as part of the TAO array for over 10 years (e.g., Sprintall and McPhaden, 1994) but only from a limited number of sites. For TRITON, salinity is a standard measurement. Promising new methods to derive upper-ocean salinity profiles using sea-surface salinity data, mean T-S curves (e.g., Vossepoel *et al.*, 1999) and/or EOF analysis (Maes, 1999) are under development but direct observations are needed. Salinity will be measured by Argo, XCTDs and perhaps drifters. If, however, there is a need for higher temporal resolution in studying ENSO and its decadal modulation, this could be obtained by enhancing in situ salinity measurements on TAO.

4.4.2 Velocity measurements

At present, the TAO array includes current measurements at only five sites along the equator (Fig. 4.12). At these locations, ADCPs and mechanical current meters are deployed to provide profiles of velocity in the upper 250 m. The failure of geostrophy near the equator makes direct velocity measurements useful at all depths. These data have proven very valuable for model validation and development, and for diagnosing oceanic processes involved in generating ENSO timescale variations. Because of the near equipartition of kinetic and potential energy in low-frequency motions near the equator, direct velocity observations are particularly sensitive diagnostics of ocean-circulation models. However, direct velocity measurements are sparse, particularly near the surface (upper 50 m) and at off-equatorial sites, and this has limited the ability to define the time-varying velocity field near the equator.

Surface drifters diverge from the equator because of poleward Ekman flows. Shipboard ADCP data lack sufficient temporal and near-surface coverage. As part of PACS/EPIC, some near-surface real-time current meters are already planned along 95°W (at 10 and 40 m). Also, TRITON moorings along 156–137°E will be instrumented to measure velocity at a single 10 m depth. However, large expanses of ocean (110°W to 165°E, inclusive) will be uninstrumented for direct moored velocity measurements. Additional current measurements could be a valuable and cost-effective addition to the TAO array.

4.4.3 Surface fluxes

Knowledge of the fluxes of heat (turbulent and radiative), fresh water (evaporation and precipitation), and momentum (wind stress) at the air-sea interface is central to better understanding ocean-atmosphere coupling. As discussed in Section 5 on air-sea fluxes, present flux estimates based on VOS measurements, numerical weather prediction models, and satellite data are not accurate enough for climate studies. For example, they exhibit errors in the tropical Pacific on seasonal-to-interannual timescales that are sometimes as large as the climate signals of interest. The PBECS strategy for fluxes includes maintaining a number of moorings and VOSs, making high-quality surface observations that will allow the bias in flux analyses to be tested, corrected, and/or validated. The PACS/EPIC program will contribute to this effort in the eastern Pacific by equipping several TAO buoys along 95°W to measure additional surface meteorology during 1999–2003. Similar efforts to add upgraded surface meteorology measurements to some TAO buoys, and to maintain them during PBECS, should be planned.

4.5 Autonomous gliders for boundary currents

Monitoring boundary currents is challenging because they are swift, narrow, and highly time dependent. The WOCE strategy of using dense mooring arrays to monitor transport is well proven, but the expense of maintaining such arrays for climate timescales is considerable. Alternatively, it is also often possible to rely on geostrophy by measuring the horizontal pressure field through a combination of moorings, autonomous vehicles, satellite altimeters, and, where available, expendable profiles from Voluntary Observing Ships. Since pressure is the spatial integral of current, horizontal resolution can be coarser than for velocity measurements. In principle, pressure measurements straddling a boundary current are enough to measure mass transport (but not heat and freshwater transport). In practice, the absolute pressure field is difficult to diagnose and the horizontal extent of boundary currents is difficult to determine because of adjacent recirculation features. To meet PBECS objectives, boundary current measurements must determine mass, heat, and, in most cases, freshwater transports, and extend from the boundary into any recirculation regions.

Three relatively new technologies offer promise for monitoring boundary currents geostrophically: gliders, hydrographic moorings, and satellite altimeters. Gliders are capable of repeatably measuring temperature and salinity structure from the sea surface to 2 km depth either at one nominal location or along programmable routes. Hydrographic moorings are able to measure temperature and salinity profiles from near the surface to near the ocean bottom on sub-daily intervals. Satellite altimeters measure the sea-surface height relative to an unknown spatially varying mean. We here discuss the possibility that autonomous gliders can make it more feasible to monitor boundary currents.

Gliders can be guided autonomously to cover horizontal ranges of several thousand km, but only by traveling relatively slowly, about 0.5 kt. This is slower than even the depth average over the upper ocean of many boundary currents, particularly those on the western ocean boundary. Gliders can hold station (or be "virtually moored") only inshore or offshore of these strong flows. They can survey boundary currents repeatedly by crossing them while being swept downstream and then gliding upstream offshore where flow is weak or recirculating. At the speed they travel, a survey across a 200 km wide boundary current might be repeated approximately monthly for a year or more from a single glider deployment.

In addition to collecting temperature and salinity profiles, the difference between glider displacements over the ground and their dead-reckoning displacement gives a measure of depth-averaged current. Gliders determine their position by GPS fixes taken at the sea surface. Speed through the water can be measured or estimated from vertical velocity, attitude, and known hydrodynamic performance. In principle, geostrophic shear from glider profiles can be combined with depth-averaged speed to estimate absolute geostrophic current.

Moored hydrographic profilers (Fougere and Toole, 1998) measure temperature, salinity, and velocity from the ocean bottom to near the surface. They employ an instrument package that climbs up and down the mooring line using a traction motor. In strong currents, hydrodynamic drag on the vehicle and mooring line overcomes the available thrust so that moored profilers, like gliders, cannot maintain a fixed-point time series in swift currents. However, pairs of moored profilers spanning a current could monitor the shear of geostrophic transport over nearly the entire water column.

Satellite altimetry gives estimates of the surface pressure field across boundary currents with respect to the unknown geoid shape. By repeating tracks, useful estimates of the time dependent

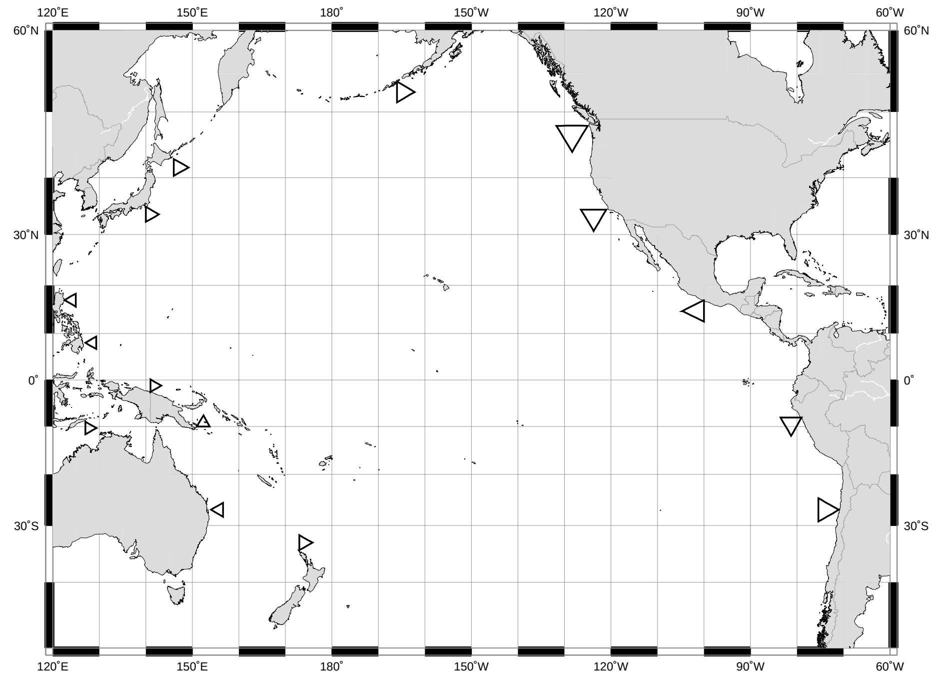

Figure 4.13: A hypothetical array of autonomous glider boundary current measurements for PBECS. Each triangle schematically represents the path a glider might follow on a 30- to 40-day period.

part of the surface flow of boundary currents can be made. Unfortunately, errors associated with altimeter transition from land to ocean along these tracks introduces additional error and it is not now clear how close to shore altimeters can be accurate.

Later in this plan a number of boundary flows around the Pacific are identified as figuring in climate variability phenomena. A possible set of repeat survey transects across boundary currents is given in Fig. 4.13. The triangular patterns, representing schematic glider tracks, are 200 km on a side for western boundary currents and 500 km on a side for eastern boundaries; surveys would be carried out so that currents are crossed in a downstream sense. The locations cross the major boundary currents of the Pacific basin, are suitable for monitoring cross-basin pressure differences, and coincide with established VOS XBT lines. Ideally, tracks would coincide with altimeter tracks but this was not done in the figure. As a guide to potential cost, gliders continuously surveying all 15 triangular patterns might cost as little as \$500,000/yr to maintain for 5 years.

The combination of glider repeated transects across a boundary current, time series of hydrographic structure on opposite sides of it, and altimeter tracks across it provides valuable comparisons. In order to evaluate the efficacy of each method and decide what density of measurements is necessary for long-term monitoring of all the boundary currents in the Pacific basin, a well-designed field trial is needed.

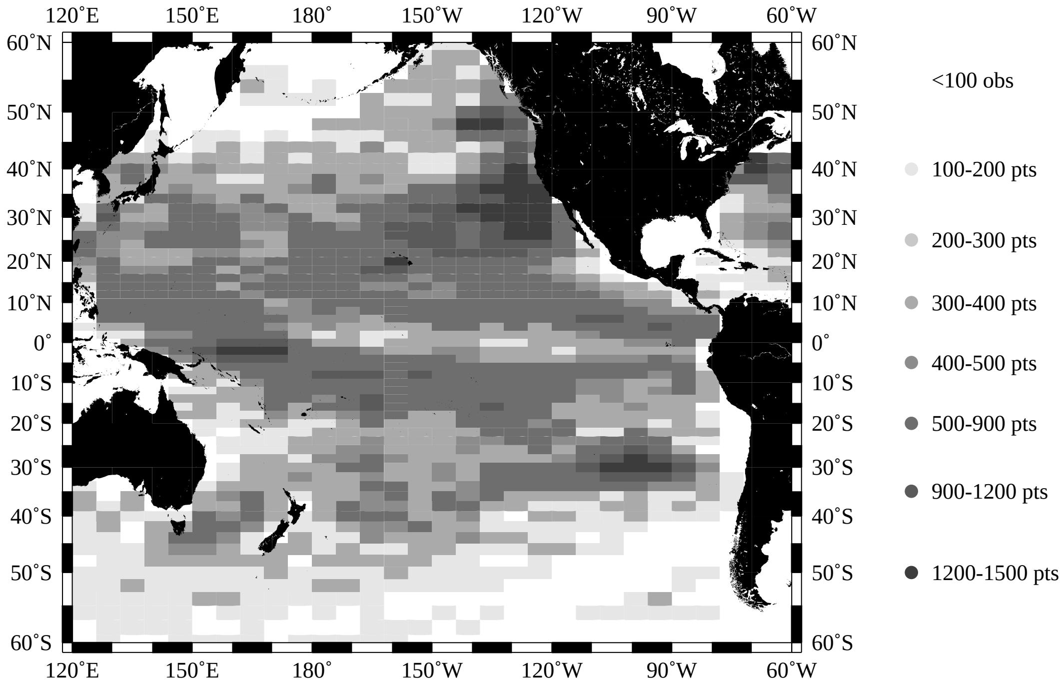

Figure 4.14: The density of drifter observations in the Pacific during 1999. Gray code represents the number of 2-day-average surface velocities measured at 15-m depth per 2° latitude by 6° longitude box.

4.6 Surface drifters

On decadal timescales and longer, advection dominates storage in the conservation equations determining the temperature and freshwater distribution of the upper ocean. The advection process is complex because it can be three-dimensional, and both the time-mean and temporal changes of the temperature and salinity gradients, as well as of the velocity, can combine to produce the spatial and temporal evolution of the fields. In contrast to seasonal-to-interannual timescale phenomena, where the storage rate of heat is important and can be determined principally from the local rate of change of temperature (e.g., Moisan and Niiler, 1998), in decadal processes knowledge of the velocity field is paramount. For assimilation analyses to be useful in describing decadal-scale changes, the analyzed upper-ocean circulation must be both constrained and tested with direct velocity observations. For example, in the shallow limb of the subtropical overturning cell the Ekman component is stronger than the geostrophic component. Because the strength of Ekman circulation depends upon the vertical and horizontal distributions of small-scale mixing that are highly parameterized in models, it is essential to directly observe this circulation component. Direct observations of the circulation of the upper ocean provide a powerful data set for testing model performance, whether they are Eulerian (Saunders *et al.*, 1999) or Lagrangian statistics (Baturin and Niiler, 1997).

Since 1988, drifter observations made with drogues at 15 m (see Fig. 4.14) have provided a basin-wide description of the tropical near-surface velocity and SST. In 1999, the drifter array within 20 degrees of the equator provided over 32,000 daily-average velocity and SST observations. These tropical observations will be continued as part of the ENSO observing system and enhanced by the Consortium on the Ocean's Role in Climate. Northward of 30°N, a small number of drifter deployments will occur for a few years off the Oregon coast and within the Japan-East Sea. While the near-surface velocity data set in the tropical basin is, and will continue to be, robust, observations in the subtropics and subpolar basins will no longer be available unless PBECS mounts them. Data in these regions was collected between 1992 and 1997 as part of WOCE-SVP. The WOCE data set in the subpolar North Pacific is not robust enough to define either the mean circulation or its variance. In the subpolar basins of the North Atlantic, near-surface circulation observations have revealed strong topographic constraints on the surface current systems (Poulain *et al.*, 1996), indicating a significant barotropic component of the circulation. Strong barotropic circulation features in the Kuroshio Extension have been documented with moored current-meter observations (Niiler *et al.*, 1985). The combined data set of circulation observations by Argo floats and drifters can be used to further explore the vertical structure of the subpolar gyres in the Pacific.

The circulation of subpolar North Pacific could, for example, be systematically observed with 206 drifters per year deployed from VOS; 110 deployed in the central gyre north of 40°N would achieve the array density used in WOCE and 48 could be deployed in the Alaskan Stream and 48 in the Oyashio. These boundary-current deployments could occur quarterly from fishing and fisheries survey vessels along four transects with four drifters deployed on each. The U.S. might be responsible for the eastern basin half of the array while Japanese colleagues might take responsibility for the western-basin deployments. Some cooperation might be given by operational meteorological agencies of U.S., Canada, and Japan who can use the drifters to enhance surface observations of atmospheric pressure and winds for marine forecasts.

4.7 Repeat hydrography

While Argo will provide temperature and salinity profiles in unprecedented density, there are at least three reasons this will not completely replace traditional hydrographic observations for climate purposes:

1. There are many variables (e.g., time-tagged chemical tracers like CFCs, CO_2, nutrients, and oxygen) that cannot be measured from long-lived autonomous instruments. These fields are undergoing climate timescale changes of measurable magnitudes. Tracers, particularly time-tagged tracers, give a very useful view of subduction, mixing, and advection processes that adds to what can be learned from temperature and salinity. The most important information may be the changes of property distributions on neutral (isopycnal) surfaces. These changes are slower than those of the density field that Argo will be adept at measuring, so occasional repeat sections will be useful.

2. For the foreseeable future, extracting the highest accuracy from Argo salinity profiles will depend on comparison with other well-calibrated observations. To some extent recently deployed floats, with recent reliable calibrations, can provide the needed comparisons. This will, however, never provide the level of accuracy in determining property relations (i.e., temperature-salinity relations) that can be obtained from high quality hydrographic observations. Hydrographers typically "calibrate" their CTD profiles with more accurate bottle data and Argo sensors are simply CTDs that have not been calibrated for some time.

3. Argo sampling is confined to the upper 2000 m. This encompasses all the water ventilated within the Pacific Basin, but gives no measure of how the deeper waters, ventilated in the North Atlantic and the Antarctic, are changing. Changes in these deep waters probably do not feedback on the atmosphere on decadal timescales, but CLIVAR, to which PBECS

contributes, is also concerned with longer timescales associated with anthropogenic climate change. Some observations of the deep ocean are needed so that future analysts can test hypotheses about the role that the thermohaline circulation plays in climate variability on these timescales.

PBECS has not yet determined what, where, or how often repeat hydrographic observations should be made. Something like five to seven zonal and meridional lines throughout the basin, each occupied on a 5- to 10-year period, would be reasonable. A fairly comprehensive suite of observations would be advised until we can understand how the various tracer fields can be related to each other. PBECS will work with other nations around the Pacific to develop a sampling and appropriate sharing of opportunities.

4.8 Ocean time series

With the exception of the TAO array and moorings associated with improving surface fluxes, the PBECS observations now identified will not provide time series at fixed points. While drifting and VOS sampling provide cost-effective areal coverage, the reasons to augment these with fixed-point ocean time series were spelled out in the recommendations in relation to climate modeling made at the OOSDP/GOOS/CLIVAR Workshop on Ocean Time series (Baltimore, MD, 1997):

1. Sustain long, consistent, in situ time series of ocean climate variables for the following purposes: direct comparison with ocean-model hindcasts; determination of statistics of signals and model errors; assessment of the climate state relative to the historical record, and to rates of change of different components of climate system; specification of initial conditions for climate predictions and for calculation of skill scores.

2. Develop new in situ time series of ocean climate variables for the purposes above, and to help resolve the spatial structure of important signals.

3. Maintain existing ocean weather buoys for their contribution to improving ocean surface forcing fields and their error characteristics.

4. Deploy a limited number of state-of-the-art air-sea flux buoys to improve and calibrate numerical weather prediction (NWP) flux fields in distinct regions of the climate system.

5. Deploy high-quality meteorological buoys in sufficient numbers and appropriate places to detect discontinuities of analyses.

This same meeting noted that ocean climate observations differ in value: integrating variables (such as sea surface height, heat content, and tracers) are particularly valuable; observable climate indices should be developed and maintained for key oceanic regimes relevant to the climate system; and collocated, multivariate time series are particularly valuable for determining climate-relevant feedbacks.

The only existing Pacific time-series station (other than approximately seasonal CalCOFI hydrographic stations along the California Coast) is from the Hawaii Ocean Time series (HOT) project, which has measured physical, chemical, and biological variables at a deep-water site 60 miles north of Oahu, Hawaii. A comprehensive set of observations has been made at approximately monthly intervals since 1988, including the following physical variables: ocean and surface temperature, salinity, upper-ocean velocity, dissolved oxygen, nitrate and nitrite, phosphate, silicate, pH; alkalinity, various forms of organic and inorganic carbon, atmospheric pressure, air temperature, humidity, wind velocity, precipitation, and cloud cover.

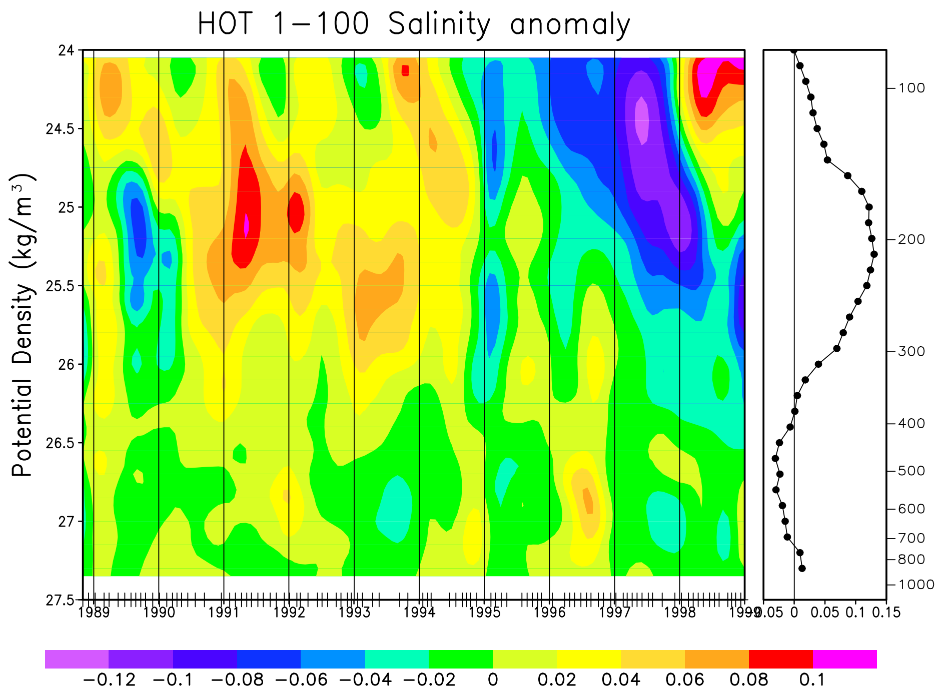

Figure 4.15: Salinity on potential density surfaces from the Hawaii Ocean Time series (HOT) project. Note substantial freshening which Lukas (1999) attributes to changes in subtropical precipitation.

Among other climate-related signals that have been observed by HOT is a pronounced freshening of the upper and main pycnocline, shown in Fig. 4.15, with peak-to-peak variation of more than 0.15 psu. This freshening has been related (Lukas, 1999) to decadal variation of rainfall patterns over the North Pacific that affect the interior ocean through subduction processes. While this signal should be evident in assimilated Argo salinity measurements, having at least one Eulerian time series of salinity would be important to check the skill of the analysis-plus-observations system. If the HOT observations cannot be reproduced sufficiently well, then we would be alerted to the need to spend extra effort to determine the source of the errors, which might be due to insufficient data, inadequate model physics, surface forcing fields, assimilation technique limitations, failure of various instruments, or combinations of these.

5. Surface Fluxes

As discussed in Sections 2 and 3, air-sea fluxes of momentum, heat, and water play a central role in testing and improving the component of coupled ocean-atmosphere models. They are also particularly useful for diagnosing the mechanisms of ocean-atmosphere coupling in models and nature. The importance of the fluxes in understanding ocean and atmosphere budgets and their variability has long been recognized (e.g., Bunker, 1976; Talley, 1984; Carissimo *et al.*, 1985; Trenberth and Solomon, 1994), but flux estimates have been plagued with errors due to the lack of direct and indirect measurements, and uncertainties in how to parameterize them from routine observations (Gleckler and Weare, 1997; Blanc, 1985; Josey *et al.*, 1999).

Over a typical year in the extratropics, the amount of heat that is stored in the ocean (spring and summer) and lost (fall and winter) is mostly balanced locally by air-sea fluxes (Gill and Niiler, 1973; Moisan and Niiler, 1998). Although there is a strong consensus that in the Pacific the oceanic heat transport is northward and divergent (supplying heat to the atmosphere) north of 5–10°N (Moisan and Niiler, 1998; Talley, 1984), the magnitude of the inferred heat flux divergence is uncertain because the net ocean-to-atmosphere flux has an uncertainty of 10's of W/m^2. This may be compared to the mean net ocean shortwave flux into the sea surface of about 175 W/m^2, or the mean net longwave flux emitted by the ocean surface of about 65 W/m^2 (Gill, 1982). A recent study by Josey *et al.* (1999) illustrates the difficulty of obtaining an accurate large-scale heat budget using bulk formulae and routine marine weather data: their global ocean average net heat-flux was a 30 W/m^2 ocean gain as derived from the 1980–1993 COADS (Comprehensive Ocean Atmosphere Data Set, a climatology based largely on VOS reports of surface meteorology (da Silva *et al.*, 1994)).

A key component of the hydrological cycle is evaporation from the ocean surface which, as a global average, is estimated to exceed the 1 m/year of ocean precipitation by approximately 10% (Baumgartner and Reichel, 1975). The distribution of evaporation and precipitation has great geographical structure, however, and regions of greatest evaporation (subtropical highs, western boundaries of the ocean basins) do not coincide with regions of greatest precipitation (tropical and subtropical convergence zones and mid-latitude storm tracks). Evaporation cools and concentrates salt in the ocean, supplies the atmosphere with heat when this vapor is condensed, and supplies the planet with precipitation. Water vapor is the atmosphere's most important greenhouse gas, and it is widely held that evaporation will increase as global temperatures rise (IPCC, 1996). However, important questions remain, such as the actual rate of increased evaporation per unit rise of sea-surface temperature, how much the atmospheric water vapor content will change, to what extent will there be negative feedbacks by cloud cover and atmospheric humidity, and how these properties will be distributed over the global ocean (Yang and Tung, 1998; Raval *et al.*, 1994).

All of the coupled phenomena of interest in PBECS (see Section 2) depend on surface fluxes for coupling, and the questions about decadal variability involve subtle connections between SST and surface fluxes. Unfortunately, "the present generation of global coupled ocean-atmosphere GCMs contains considerable systematic errors both in terms of net surface heat flux and simulated SSTs" (Meehl, 1997). An example of the importance of fluxes to climate is the sensitivity of coupled ocean/atmosphere GCMs to shortwave fluxes in the heavy low-level stratus region off the Peruvian coast. There is a marked improvement of the global atmosphere and ocean simulation when this stratus influence is fixed in the model (Ma *et al.*, 1996). The Pacific climate shift in 1976–1977 (Trenberth, 1990; Zhang *et al.*, 1997) was followed by a string of deepened winter Aleutian Lows over the North Pacific. With this change in circulation regime, multiple causes drove the North Pacific SST regime into cool anomalies along the western and central ocean with warm anomalies along the eastern boundary (Miller *et al.*, 1994). Anomalous turbulent fluxes were important causal agents in both cooling the central North Pacific and warming the eastern boundary region, but wind mixing and advection were also identified, indicating that an observational program must be broad enough to identify several mechanisms.

While large (50–100 W/m^2) monthly/seasonal regional latent and sensible flux anomalies are relatively easily seen, it is much more difficult to determine interannual and interdecadal net-flux deviations. Similarly, area-mean flux anomalies that emerge over large areas, such as the entire mid-latitude North Pacific, are dubious because of systematic errors contained in each of the components.

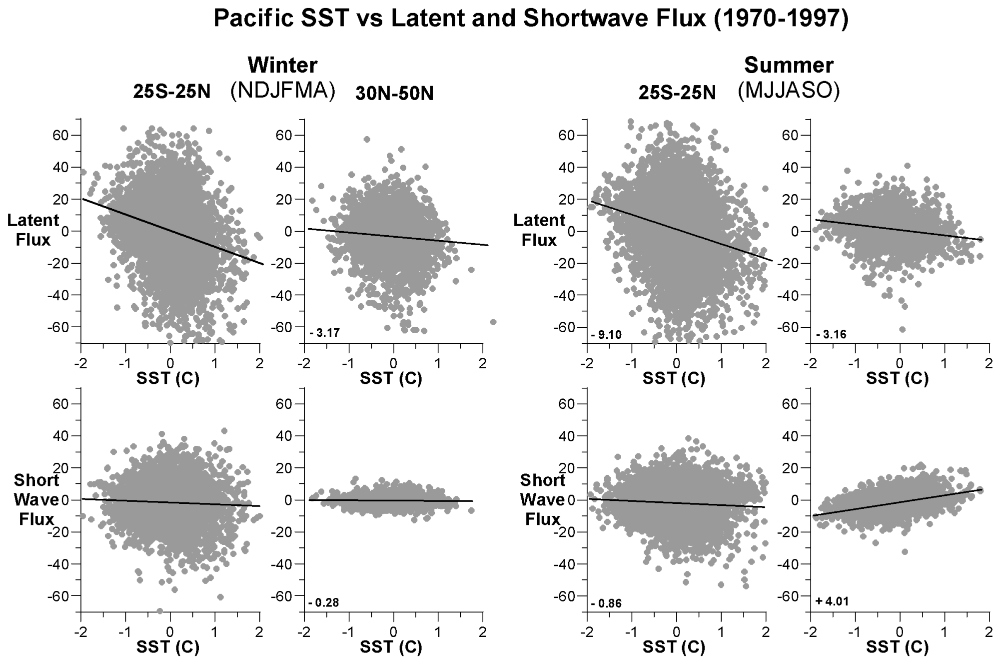

Figure 5.1: Latent and shortwave flux and SST monthly anomalies (1970–1997) in the tropical Pacific (25°S–25°N) and North Pacific (30–50°N) for November–April (left) and for May–October (right). Fluxes and SST are from COADS data. Fluxes are from bulk formulae as in Cayan (1992a). Anomalies are from 1950–1997 climatology.

5.1 The Ocean's Role in Climate

Climatologically, air-sea fluxes provide vital sources and sinks of heat and moisture to the global atmosphere (Trenberth and Solomon, 1994). The geographic distribution of oceans, continents, and the attendant fluxes has an important influence upon the disposition of storm tracks and quasi-stationary planetary waves. Mean ocean-surface conditions modulate the atmospheric phenomena. For example, it is well established that deep tropical convection is limited to oceanic regions with surface temperatures in excess of 28°C (Graham and Barnett, 1987; Inamadar and Ramanathan, 1994), and these same regions harbor strong intraseasonal wave activity such as the Madden-Julian Oscillation (Madden and Julian, 1994).

In the extratropics, it is clear that the atmospheric circulation, via anomalous winds and fluxes, is the dominant driver of upper ocean thermal anomalies (e.g., Davis, 1976), but a long-standing issue has been the extent to which anomalous sea surface temperature (SST) feeds back to affect the atmosphere, which has its own strong internal variability (Robinson, 2000). An important question, that lies at the heart of many modeling and diagnostic studies, is whether, and under what conditions, significant heat flux is associated with SST anomalies. Indeed, the sign of the relation between anomalous heat flux and SST is a direct indication of whether the SST modulates the flux or vice versa. Latent and shortwave fluxes appear to be the two most important flux anomaly components (Cayan, 1992a). These components are very different in their origin and connections to SST.

Figure 5.1 illustrates how the latent and shortwave fluxes in the tropical and North Pacific are related to SST anomalies, as determined from a collection of monthly-anomaly pairs of SST and fluxes from 1970–1998. In the tropics, regardless of season, latent anomalous flux loss (from the ocean) increases with increased SST anomaly at a rate of about -10 W/m^2 per °C, supporting the idea that the ocean can drive the tropical atmosphere. In the extratropics, particularly in warm seasons, there is also a tendency for latent flux losses to increase with increased SST, but the response is smaller, only about -3 W/m^2 per °C. The extratropical North Pacific exhibits a similar connection in the cool season, but the relationship has more scatter, presumably an indication that the simplistic "ocean driving atmosphere" (or SST damping) relationship can be overridden by other factors. Because similar results are found for the North Atlantic (not shown), these results appear to be at least qualitatively consistent. Interestingly, in the warm season there is the opposite tendency for warm SST anomalies to occur when the shortwave flux is into the ocean; the sensitivity is about $+4$ W/m^2 per °C. This link appears to be the positive feedback between warmer SST, reduced marine layer static stability, and diminished stratus cloud cover noted in previous studies (Weare, 1994; Klein and Hartman, 1993; Norris *et al.*, 1998).

There is some evidence that, as timescales lengthen, there is a shift in the balance between atmosphere-forcing-ocean and ocean-forcing-atmosphere. At weekly to seasonal timescales (Cayan, 1992b; Deser and Timlin, 1997), the overwhelming link between anomalous heat flux and SST is that of the atmosphere driving the ocean, particularly in the extratropics. In the tropics there is also a discernible ocean-forcing-atmosphere (SST damping) signal, where anomalous latent flux losses from the ocean increase as the SST increases. At interannual timescales, however, other mechanisms such as oceanic advection of thermal anomalies play a more significant role in driving the ocean temperature, and there is increasing evidence, even in the extratropics, that fluxes become more determined by the SST, as shown in Fig. 5.2 (middle). At decadal timescales (Fig. 5.2, bottom), the tendency for the ocean to heat the atmosphere in the extratropics is even more evident, although comprehensive marine records are not long enough to provide high statistical confidence. Determining this low-frequency relation is particularly challenging because several variables are involved, the SST "signal" is only a few tenths of a degree (White *et al.*, 1997), and is small compared to observation and flux parameterization errors. Some errors have time variability of their own, which can masquerade as climate variability. To confirm and elucidate these results will require stable long-term observations of the fundamental weather variables over the Pacific basin.

5.2 The status of presently available flux fields

At present, surface flux data are mainly obtained from three sources: in situ measurements from ships and buoys, remote sensing from satellites, and output from numerical weather prediction (NWP) models such as those at NCEP and ECMWF. Each source has advantages and limitations. In situ observations of moderate quality are available along ship routes from Volunteer Observing Ships (VOS). High quality surface meteorology and surface fluxes can be obtained from a limited number of moored buoys. All surface observations depend on bulk parameterizations to produce fluxes and these need improvement, particularly under high winds and stable conditions. Satellites now provide global measurements of SST, surface radiation, and vector wind, and studies are underway to examine the extent to which sensible and latent heat fluxes can be derived from remotely sensed data. However, satellite measurements need in situ calibration and validation, and satellite orbits limit the time resolution possible in the global fields. NWP models have the advantage that they can provide global surface meteorological and flux fields on global grids at time intervals (6 hours) that resolve diurnal and synoptic weather variability. These fields, though,

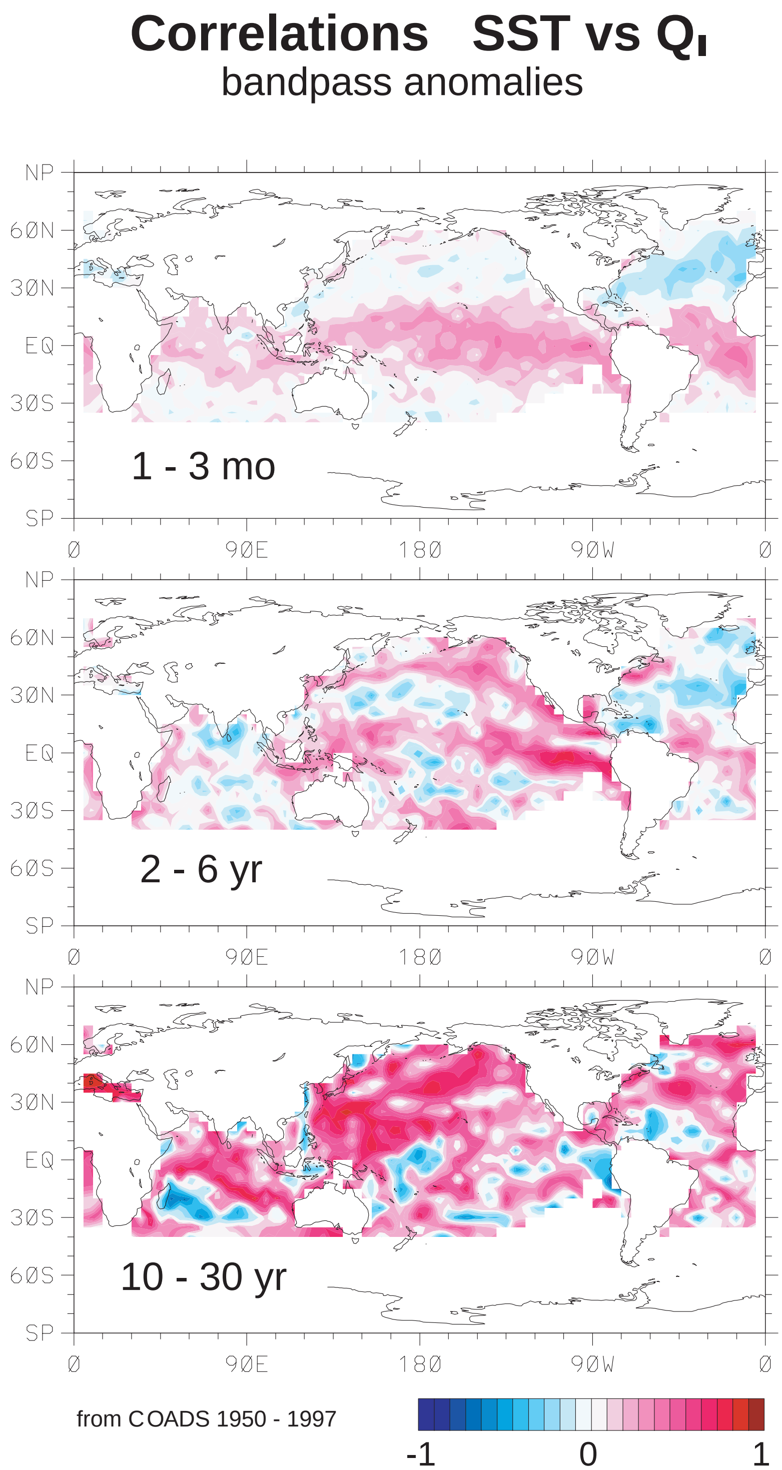

Figure 5.2: Correlations for band pass filtered latent flux vs. SST anomalies for monthly (1–3 mo, top panel), interannual (2–6 yr, middle panel), and decadal (10–30 yr, lower panel). Fluxes and SST from COADS data. Latent flux based on bulk formulae as in Cayan (1992a). Anomalies from 1950–1997 climatology.

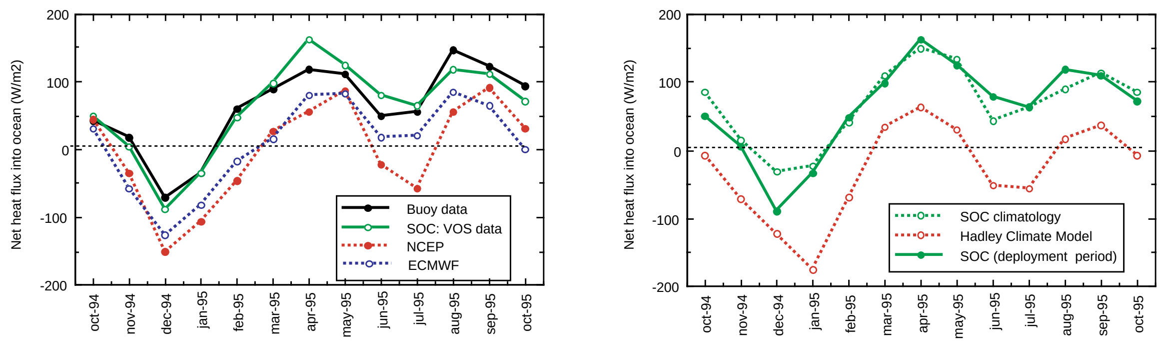

Figure 5.3: Time series of monthly net heat flux at 15.5°N, 61.5°E in the northern Arabian Sea (adapted from Weller *et al.*, 1998). On the left is a comparison of the net heat flux based on data collected by an IMET-equipped buoy from October 1994 to October 1995; from VOS-based fluxes for the same time and from the vicinity of the buoy in the Southampton Oceanography Centre (SOC) surface-flux data set; and from surface fluxes computed using the COARE bulk formulae and the surface meteorology from the nearest grid points in the ECMWF and NCEP model fields together with the model net longwave and net shortwave fluxes. On the right is a comparison of the buoy net heat flux from 1994–1995, the SOC climatology (1983–1995) at the site, and the net heat flux from the Hadley Centre Climate model (Hall *et al.*, 1995).

may contain large uncertainties due to deficiencies in the models and their inability to assimilate satellite and in situ surface meteorological data.

An illustration of the magnitude of error in the NWP fields is provided by Fig. 5.3, taken from Taylor *et al.* (1999a) on strategies for obtaining surface fluxes globally. At this Arabian Sea site both the NCEP and ECMWF models had incoming shortwave radiation fluxes that were smaller than observed, and latent heat flux larger than observed. Throughout the year the ECMWF and NCEP net heat fluxes were biased low. The annual mean net heat flux from NCEP was -5 W/m^2 and from ECMWF was 10 W/m^2 compared to the 60 W/m^2 observed at the buoy. There was, however, good agreement between the buoy observations and the Southampton Oceanography Centre (SOC) VOS-based net heat flux. Comparison of the SOC VOS climatology of net heat flux with the climatology from the Hadley Centre climate model suggests that the climate model also fails to provide an accurate measure of the air-sea heat exchange.

5.3 Strategy: In situ observations

The goal for the PBECS flux program should be obtaining air-sea fluxes over the Pacific basin with net heat flux accurate to better than 10 W/m^2. A net heat flux of 10 W/m^2 would heat the upper 500 m of the ocean $\sim$0.15°C in 1 year, a signal the size of decadal changes reported in the Atlantic (Parilla *et al.*, 1994). To achieve this goal, the components of the net heat flux, which vary on a wide range of time and space scales over magnitudes approaching 1000 W/m^2, must each be measured to an accuracy of a few W/m^2.

Deployments of well-instrumented buoys coincident with ship measurements of surface fluxes (including direct eddy-covariance measurements) have shown that mean fluxes ($\sim$3-week average) from such buoys meet the heat flux goals (Weller and Anderson, 1996). Buoys are now equipped to measure rainfall and vector winds to determine the freshwater and momentum fluxes. On a limited basis, such buoys are also equipped to measure the turbulent flux directly, providing both accurate measurements of the mean observables, and supporting studies of the transfer coefficients for bulk formulae. However, it is unlikely that buoys could be deployed in numbers sufficient to map the fluxes over the Pacific.

Historically, fluxes have been mapped using surface meteorology from the VOS together with empirical formulae. Early VOS data had data-quality problems, especially due to different methods of obtaining SST, to disturbance of the airflow by the ship itself, and to errors in calculating the absolute wind velocity (Folland and Parker, 1995; Kent *et al.*, 1993a; Taylor *et al.*, 1999b). More recently, as in the preparation of the SOC version of VOS-derived flux fields, corrections (Kent and Taylor, 1995, 1996, 1997; Kent *et al.*, 1993b; Taylor *et al.*, 1995, 1999b; Taylor and Kent, 1999) have significantly improved the quality of VOS-based fluxes. The success of these improvements is evident in the good agreement between the SOC VOS-based net heat flux and the buoy-observed net heat flux shown in Fig. 5.3. Direct flux measurements can be made from ships (Fairall *et al.*, 1997) and would provide the means to continue improvements to the bulk formulae as well as validation for fluxes derived by other methods.

Comparisons of VOS fluxes with well-equipped buoys, and logistical considerations of VOS observations, show where additional improvements can be made. Traditional VOS instrumentation has been basic and limited, including barometer, anemometer, engine injection temperature sensor, and wet- and dry-bulb thermometers. Direct observations of shortwave and longwave radiation have not been made. While the gain from improved instrumentation for VOS has been evident, a specific ship remains on a given route for only a few years, so the strategy for improving the VOS must include ease of installation and removal. An example of the accurate and portable hardware that has been developed to install on VOS is the Improved Meteorological System, IMET (Hosom *et al.*, 1995) that has been installed on a number of U.S. research vessels and, in a pilot program, on two U.S. VOS in the Pacific. IMET uses sensors chosen for accuracy, reliability, and their ability to stay in calibration during unattended operation. Sensors and electronics are combined into a digitally addressable module that retains its calibration information. A standard PC can be used for data acquisition and display. The need for extensive hardwiring during the installation on the VOS is relieved by using both RF and through-the-hull acoustic data transmission from sensors. The present set of IMET modules includes wind speed/direction, air temperature, sea surface temperature, relative humidity, precipitation, incoming shortwave radiation, incoming longwave radiation, and barometric pressure.

The in situ Pacific flux array will be a combination of well-equipped surface moorings labeled, "surface flux reference sites," other surface moorings including those of the National Data Buoy Center and the TAO array, improved VOS, routine VOS, and drifting buoys. Surface flux reference sites measure all the observables needed to determine the air-sea fluxes of heat, freshwater, and momentum with the intent of achieving the accuracy obtained in TOGA COARE (Weller and Anderson, 1996). These buoys provide high temporal resolution (e.g., 1-minute samples for 1 year with satellite telemetry of hourly values).

Improved VOS will have the equivalent of IMET sensors for the observables needed for air-sea fluxes, including shortwave and longwave radiation and precipitation, and will be processed using the methods developed at SOC. Equipping and operating the improved VOS should be coordinated with the present Ship Of Opportunity Program (SOOP). VOS data taken close to the surface flux reference sites could be compared to buoy data to quantify the quality of the VOS data and to guide choices of the regionally appropriate empirical flux formulae (especially for the regular VOS without radiation sensors). Surface-flux reference sites would be selected to sample climatologically different regimes in the Pacific, to locate some sites close to ship routes, and others where data are sparse.

The VOS data, especially those from the improved VOS, could be used at various time and space lags from the surface-flux reference sites to quantify the spatial and temporal variability of the surface meteorological and air-sea flux fields. Such information is needed for mapping and to develop error statistics for data assimilation. During mapping, surface drifters can provide additional SST data and, as technology develops, other surface meteorology. The coastal NDBC and other surface moorings would provide additional time series of some observables.

The obvious limitations of such an observing system are spotty spatial coverage and limited space/time resolution. In data-sparse regions the uncertainties in even seasonal fields would be large. As a result, for some studies in PBECS, such as an examination of ocean influences on storm tracks and the role of synoptic atmospheric forcing in the evolution of the ocean in the northwestern Pacific, insufficient detail would be available and a second strategy could be employed. For a period of several years, an array of surface moorings would be deployed to explicitly resolve the space/time variability of the surface forcing fields, perhaps as a component of a regional process study.

5.4 Strategy: Blending NWP and remotely sensed fields

In situ data alone cannot provide the spatial and temporal resolution or the uniform and consistent coverage needed for PBECS. The coverage limitations of flux fields derived from in situ data alone are evident when such fields are compared with those from an NWP reanalysis (White and da Silva, 1999). Taylor *et al.* (1999a) note the decrease in correlation between the fields as one moves away from the heavily traveled ship routes. The 6-hourly time resolution, global coverage, and dynamic framework for assimilation of in situ data provided by the NWP models are attractive attributes, as is the global coverage provided by satellites.

Development of NWP models has been driven by the needs of weather forecasting without adequate attention to the quality of the fluxes they produce. Indeed, NWP flux estimates may actually degrade when improvements in forecast skill are achieved (Siefridt *et al.*, 1999). Further, assimilation of accurate surface meteorological data from VOS or buoys may, because of scale mismatches between in situ observations and model fields or because of model deficiencies, also degrade forecast skill. To prevent this, NCEP, for example, now assigns large uncertainties to in situ data: 2.5 m/s for surface winds from ships and most buoys, 2.2 m/s for TAO buoys, 1.6 mb for observed sea level pressure, 2.5°C for observed air temperature, and 20% for observed surface humidity. NWP models do not simulate clouds or radiative fluxes well, their boundary layer structure may not be well resolved and their flux parameterizations may introduce error. In several locations sensible and latent heat fluxes computed with the NWP surface meteorology and TOGA COARE bulk formulae (Fairall *et al.*, 1996) have agreed with buoy fluxes better than the fluxes provided by the model. The model surface meteorology can also be in error; Moyer and Weller (1997) noted a 1°C bias in surface air temperature over the northeast Atlantic in the NCEP reanalysis.

In addition to the in situ array discussed above, a multi-thrust strategy should be followed to bring together all available information to produce the best possible flux fields. The three main thrusts are:

1. **Use in situ data to correct NWP fields.** In the short term, accurate gridded flux fields can be created using buoy data to identify biases in NWP fields and correcting them. Weller *et al.* (1999) used data from an array of five surface moorings together with NCEP and ECMWF fields to produce hourly fluxes on a 1° grid between 10–40°N and 10–40°E. Such a procedure assumes that the errors in the NWP fields have large space and timescales (this needs to be verified) and is an empirical—and thus probably only a regional—"patch" of model deficiencies. This approach still does not make use of all available data. Observations

from improved VOS would not, for example, be assimilated by the NWP models due to the large uncertainties assigned to them.

2. **Improve NWP surface meteorology and fluxes.** In partnership with NCEP, ECMWF, and other modeling groups, work to improve NWP surface meteorology and fluxes would itself involve several elements. First, there should be an ongoing comparison of model surface meteorology and fluxes with in situ data. Data from surface-flux reference sites would be compared with model data to define biases. The WCRP Working Group on Numerical Experimentation has agreed to work together on this comparison in parallel with their ongoing Atmospheric Model Intercomparison Project (AMIP). Second, ways to routinely assimilate more of the accurate in situ surface data should be investigated. Third, the flux parameterizations used in models must be examined. While recent process studies such as TOGA COARE have improved the bulk formulae for point observations, adaptations may be necessary for the formulae to work well at model grid points. These efforts will need computer resources beyond those used in running the operational models and should be undertaken by teams having expertise with in situ hardware and data, and with model codes for assimilation and representation of atmospheric processes. This could be accomplished, for example, in partnership with modelers at NCAR working on the next generation model for NCEP. The goal would be to improve that model's abilities to assimilate surface data and produce flux fields before it becomes operational, and to follow this by ongoing study of the model's performance once it becomes operational.

3. **Improve the use of remote sensing data.** Satellites provide observations of SST, wind, surface shortwave radiation, precipitation, and less directly of surface longwave radiation, sensible and latent heat fluxes. Of particular interest are precipitation and surface radiation, as the NWP model fields of these quantities have large errors. A partnership between remotely sensed data, with their excellent coverage, and in situ data for calibration and validation should be explored. The MCSST data set prepared by optimal interpolation of satellite AVHRR, moored buoy, drifting buoy, and VOS SSTs provides a model of how to proceed. A workshop on determining turbulent fluxes via remote sensing in July 1999 brought together the in situ, remote sensing, and NWP communities to lay the groundwork for an intercomparison project. PBECS should contribute data from its long time-series stations to this comparison and work toward improved combined remotely sensed and in situ data sets.

4. **Assimilate remotely sensed and in situ data into NWP models**. The accuracy of satellite retrievals is affected by myriad factors in the atmosphere represented in NWP models. Satellite coverage does not resolve the diurnal cycle, or in some cases, synoptic weather events captured in NWP analyses. The greatest value of remote sensing will, therefore, be realized when these data are assimilated with in situ observations and robust kinematic and dynamic constraints to produce surface flux maps. NWP models now assimilate remotely sensed atmospheric-profile and surface data, and efforts should be made to improve their ability to assimilate these data and a wider range of remotely sensed fields. One approach is to use NWP fields, satellite data, and in situ data all as input to an assimilation process that accounts for the error characteristics of the model and data. An example of such a methodology is the variational objective analysis of Hoffman (1982, 1984), which is similar to that used by Atlas *et al.* (1993) for producing SSMI wind analyses and by Legler *et al.* (1989) for the FSU pseudostress analyses. This analysis involves minimizing a cost function consisting of a number of terms expressing misfit to data and kinetic constraints. Each term in the cost function has a weight determined by cross-validations with high-quality flux observations from

surface-flux reference sites and improved VOS. As PBECS progresses, more ocean data will be available and pilot ocean data assimilation will begin. These may permit ocean heat and water budgets to be added as further constraints in producing flux fields. They may also offer an independent way to validate the surface flux fields.

In summary, the PBECS strategy to obtaining accurate surface flux fields over the Pacific includes:

- adding high quality surface observations from buoys and VOS,
- using these data in cooperative diagnostic studies with both model centers and remote sensing investigators to validate flux products and to understand how fluxes are involved in climate variability,
- supporting development of the next generation AGCM to be used by NCEP so that it can routinely and successfully assimilate all kinds of routine and high-quality data, both in situ and remotely sensed.

6. Subtropical Overturning Circulation

As discussed in Section 2, a number of theories for decadal modulation of ENSO and for other processes of decadal variability involve oceanic connections between the subtropics, where atmospheric forcing of the ocean is strong, and the tropics, where the ocean efficiently affects the atmosphere. These connections are usefully discussed in the context of the subtropical overturning circulation in which water upwells at the equator where it modulates sea surface temperature, is carried poleward by Ekman transport and subtropical western boundary currents, subducts in the subtropics, and finds its way back to the equator through a combination of interior "Sverdrup" advection and concentrated currents. A main focus of the PBECS observing and assimilation system is understanding this Subtropical Cell (STC) and its role in the feedback processes that affect decadal variability and the ENSO cycle.

6.1 Interior processes

The link between the atmosphere and subsurface ocean in the subtropical Pacific is dependent on processes that ventilate the thermocline. When the wintertime mixed layer restratifies under springtime heating the newly formed thermocline has properties (i.e., temperature, salinity, potential vorticity, chemical tracers) set by interaction with the atmosphere. Once subducted, this water is insulated from further interaction with the atmosphere, thus setting water properties and circulation patterns in the permanent thermocline. The rate at which water leaves the zone of seasonal mixing into the permanent pycnocline has been termed the subduction rate (Marshall *et al.*, 1993; Qiu and Huang, 1995). This rate is closely tied to wind-curl forced downwelling and the component of upper-layer velocity along the gradient of mixed-layer depth.

An important method of transmitting the effects of atmospheric forcing to the subsurface ocean is the process of mode-water formation. Called "mode waters" because they are a common class in a census of water properties, these weakly stratified waters are typically formed in regions of strong wintertime air-sea fluxes and deep mixed layers, and then subduct mainly as a result of flow along the gradient of mixed-layer depth. Mode waters are important in the heat budget of the subtropical oceans because they are so common and because they are formed by strong atmospheric forcing. This was first recognized by Worthington (1959) in his study of 18-Degree Water in the subtropical North Atlantic. Mode water has since been identified in the western

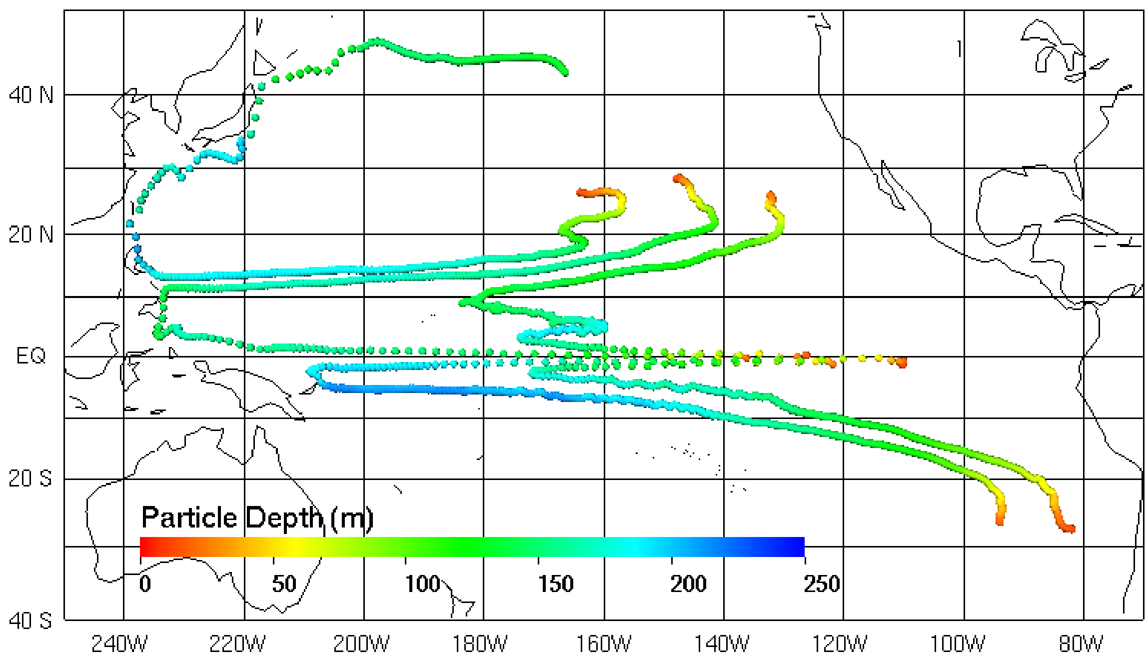

Figure 6.1: Particle trajectories from the ocean general circulation model of Gu and Philander (1997). Colors indicate trajectory depth.

North Pacific (Masuzawa, 1969), South Pacific (Roemmich and Cornuelle, 1992), central North Pacific (Nakamura, 1996; Suga *et al.*, 1997), and eastern subtropical North Pacific (Hautala and Roemmich, 1998). Atmospheric forcing can cause changes in mode water properties including the volume of a density class, the mean density and temperature of a mode water, and its location.

There are two main ways that upper ocean anomalies can be generated and subsequently impact the climate system. In addition to changes in mode water formation, property anomalies in the upper ocean can be formed by subduction throughout the polar side of the subtropical gyres. Gu and Philander (1997) suggest that subducted anomalies from the eastern Pacific can be advected along mean-current pathways (e.g., Fig. 6.1) to the tropics and then be upwelled to influence the tropical ocean-atmosphere system. In this view, subducted temperature anomalies, generated by anomalous air-sea fluxes or mixing, reach the equator by mean advection. Alternatively, as pointed out by Kleeman *et al.* (1999) and Schneider (1999), thermocline anomalies can also be generated by velocity perturbations in the (wind-driven) gyres through anomalous advection of mean properties. Mechanisms are Rossby and Kelvin waves or anomalous advection along gradients or across fronts. A combination of the two mechanisms is likely. PBECS observations and analyses must determine the important processes for generating anomalies. What is the relative importance of generation by air-sea fluxes, wind mixing, local wind forcing, anomalous advection, or other internal ocean processes? Where are subducted anomalies primarily generated?

Where subtropical anomalies are generated critically affects the path that they take to the equator (Luyten *et al.*, 1983; McCreary and Lu, 1994), because Rossby wave-ray characteristics limit the streamlines of the mean gyre flows. Anomalies generated in the eastern North Pacific and at relatively low latitudes (off Southern California) can travel to the equator along a mid-oceanic pathway, while those generated in the central North Pacific arrive at the equator via the western boundary currents. What is the relative importance of anomalies arriving at the equator via mid-ocean vs. the western boundary?

The motion of thermocline anomalies is governed by a combination of planetary wave dynamics, advection, and eddy transport, depending on the anomaly structure and the eddy field. Spiciness anomalies are passively advected along isopycnals. Theoretical studies (Liu, 1999; Huang and Pedlosky, 1999) suggest that anomalies that are first-mode baroclinic planetary waves travel westward with little effect from the mean circulation, while higher vertical modes follow the mean flow, albeit at different speeds than do passive tracers. Thermocline anomalies can also be propagated by an effective-advection velocity (Gent and McWilliams, 1990) that differs from the mean velocity by an eddy "bolus transport" velocity. This bolus velocity is a manifestation of eddy fluxes that are not of the conventional "diffusive" kind.

Interpretation of observed thermal anomalies as isopycnal depth anomalies shows them to propagate along isolines of potential vorticity (Schneider *et al.*, 1999a), but the degree to which these anomalies are density compensated is not known owing to a lack of salinity data. Open questions PBECS must resolve include: Are anomalies propagated as planetary waves, advected as T-S anomalies, or moved by anomalous eddy fluxes? Do they propagate in isolation, or continuously interact with the atmosphere? Are anomalies dissipated by fluctuating thermal and velocity fields driven by Ekman pumping, or by mesoscale variability? Higher sampling density, particularly for salinity, is needed.

PBECS interior ocean observations and assimilation analyses must describe the properties of ocean anomalies, their generation, and their equatorward transport through the subtropics and tropics. (Passage through boundary currents and emergence on the equator are considered in Section 6.2). Observing anomaly generation will depend mainly on the improved surface-forcing fields, as described in Section 5, and Argo. Anomaly transport in the interior may occur as changes of the gyre-scale circulation, as planetary waves or by changes in the eddy transport of properties. For gyre-scale variability, including detection of anomalies as they are generated, we will rely on the Argo array to measure temperature, salinity, and velocity over the entire tropical and subtropical Pacific. It is not practical to deploy an eddy-resolving array over the complete domain, so the PBECS strategy is different for large-scale and mesoscale variability. Each is discussed briefly.

6.1.1 Large-scale variability

The PBECS array must provide sufficient data to map, on seasonal timescales and with good signal-to-noise, variability in large-scale equatorward geostrophic circulation, to initialize models, and to constrain assimilation analyses. The Argo array of profiling floats will obtain temperature and salinity profiles approximately every 3° of latitude and longitude every 10 days.

As a proxy for illustrating the resolving power of Argo, we use altimetric data in the southeastern Pacific, subsampled at Argo resolution. The solid line in Fig. 6.2 shows a time series of the northward geostrophic surface current using all TOPEX data in the region 140–80°W, 18–9°S. In this region, the equatorward interior flow in the thermocline feeds the Equatorial Undercurrent (Johnson and McPhaden, 1999). The first Argo float deployments in the Pacific (funded by NOPP) are in this region. The time series in Fig. 6.2 indicates that the equatorward surface transport changes by more than 3 Sv per 100 m of depth on seasonal to interannual timescales. In terms of surface height variability, the 11-cm fluctuations are about 20% of the 55-cm slope of mean sea-surface height in the same region (as indicated by historical hydrographic data). If the signal seen at the sea surface decays linearly to zero at 400 m depth, then the magnitude of transport variability is about 6 Sv. How well is this variability resolved by the Argo array? The dashed lines in Fig. 6.2 show the standard deviation of surface-transport estimates based on 100 realizations of

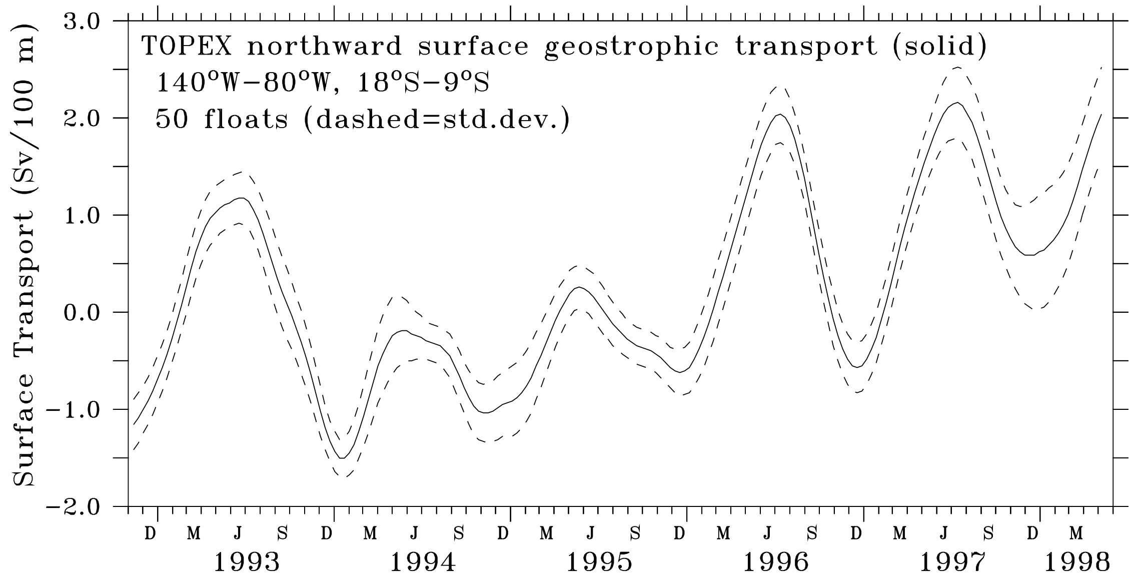

Figure 6.2: Time series of meridional surface geostrophic transport from TOPEX in the region 9–18°S, 140–80°W, smoothed with a 50-day Gaussian filter. The dashed line shows the standard deviation of the estimates, based on 100 realizations of 50 randomly subsampled points.

a randomly subsampled selection of 50 TOPEX points. The subsamples at Argo resolution provide good estimates of the large-scale geostrophic transport variability, with high signal-to-noise ratio.

An additional objective of the float array is to directly determine the Lagrangian pathways of equatorward flow in the thermocline. Some floats will park at thermocline depths in order to satisfy this objective. The tradeoffs inherent in different parking levels are not yet well understood, and more experience is needed with floats at both thermocline and deep levels. Floats parking at 2000 m will likely remain closer to their deployment location and have less bio-fouling of the salinity sensors than floats parking in the shallow tropical thermocline. In the initial Argo deployments in the southeastern Pacific, approximately half of the floats will park in the thermocline and half at a deep level. This ratio will evolve over time to optimize the array, as more is learned about the behavior of floats at each level and of the signals observed by the array.

6.1.2 Mesoscale variability

Two components of the broad-scale observing system have some eddy resolving capability. The TOPEX satellite altimeter has resolution of about 8 km along track, and the High Resolution XBT transects return temperature profiles at spacing of 10 to 50 km, depending on location. The PBECS strategy is to use the combination of these to observe the space-time characteristics (altimeter) and subsurface expression (HRX) of mesoscale variability.

The existence of Rossby wavelike westward propagating features is ubiquitous in the altimetric dataset (e.g., Chelton and Schlax, 1999). HRX data between San Francisco and Taiwan along an average latitude of 22°N reveal the subsurface structure of these features (Roemmich and Gilson, 1999; also see Section 4.3 above) and their significant contribution to meridional transport variability (Fig. 6.3). While the PBECS observing system cannot sample the entire Pacific at eddy resolving spacing, the altimeter sees the sea surface over the entire domain and the HRX transects

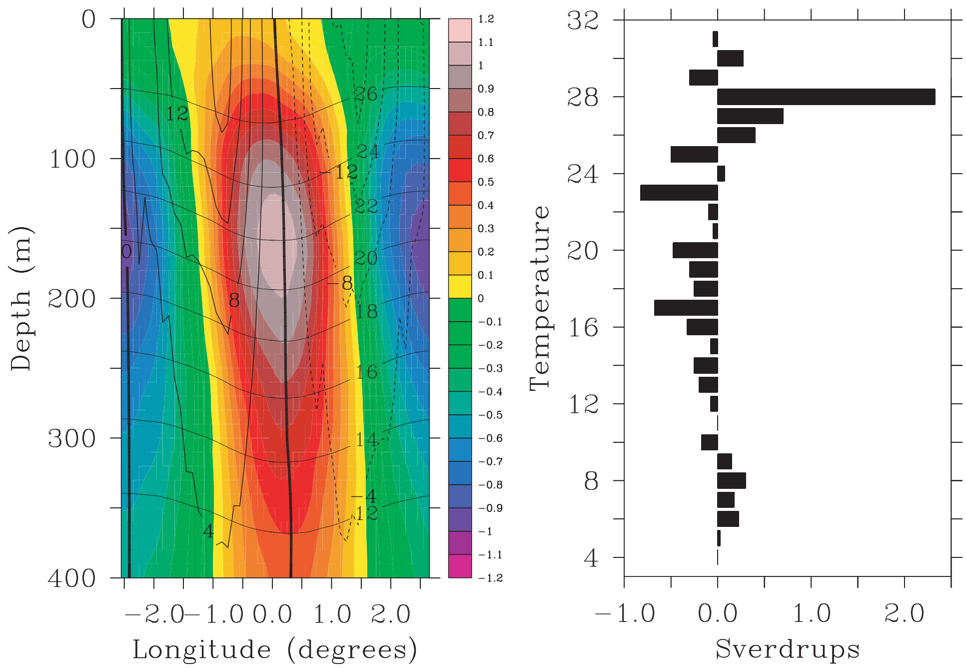

Figure 6.3: The left panel shows the mean structure of eddies (averaged over 350 eddies) along the San Francisco to Taiwan transect. Temperature and geostrophic velocity are contoured. Colors indicate temperature anomaly. The right panel shows the mean transport in temperature classes induced by the tilt of the eddies (i.e., due to $< v'T' >$). The eddies contribute over 4 Sv to the mean equatorward transport of the thermocline (Roemmich and Gilson, 1999).

reveal the subsurface structure along selected lines. Zonal HRX lines measure geostrophic transport variability in the equatorward limb of the tropical/subtropical overturning cell. Meridional HRX lines reveal the detailed structure of the strong zonal current system and sharp T and S gradients of the tropics (see Fig. 6.4). Because the zonal velocities often dominate the weaker meridional velocities of the overturning circulation, the linkage of the interior circulation to the boundary currents is emphasized. The overturning cells are the integral combination of boundary current and interior transports, with the proportions and even directions of those transports changing with latitude as the strong zonal currents add or remove water from the boundary currents. The observing system must span boundary and interior flows as well as resolving both large-scale and mesoscale variability in the interior circulation.

6.2 Low-latitude western boundary currents

One hypothesis for decadal modulation of ENSO, and for decadal variability itself, is that subtropical oceanic anomalies are advected to the equator where they can effectively influence the tropical atmosphere. While it is possible for water subducted in the subtropics to reach the equator through the interior "Sverdrup" circulation—or possibly through strong horizontal mixing due to tropical instability waves—it can also reach the equator through the western boundary currents of the tropical gyres. In the North Pacific this is the main path. PBECS will rely on the broad-scale observing system (Sections 4.1 and 4.2) to observe the interior pathways, but special consideration

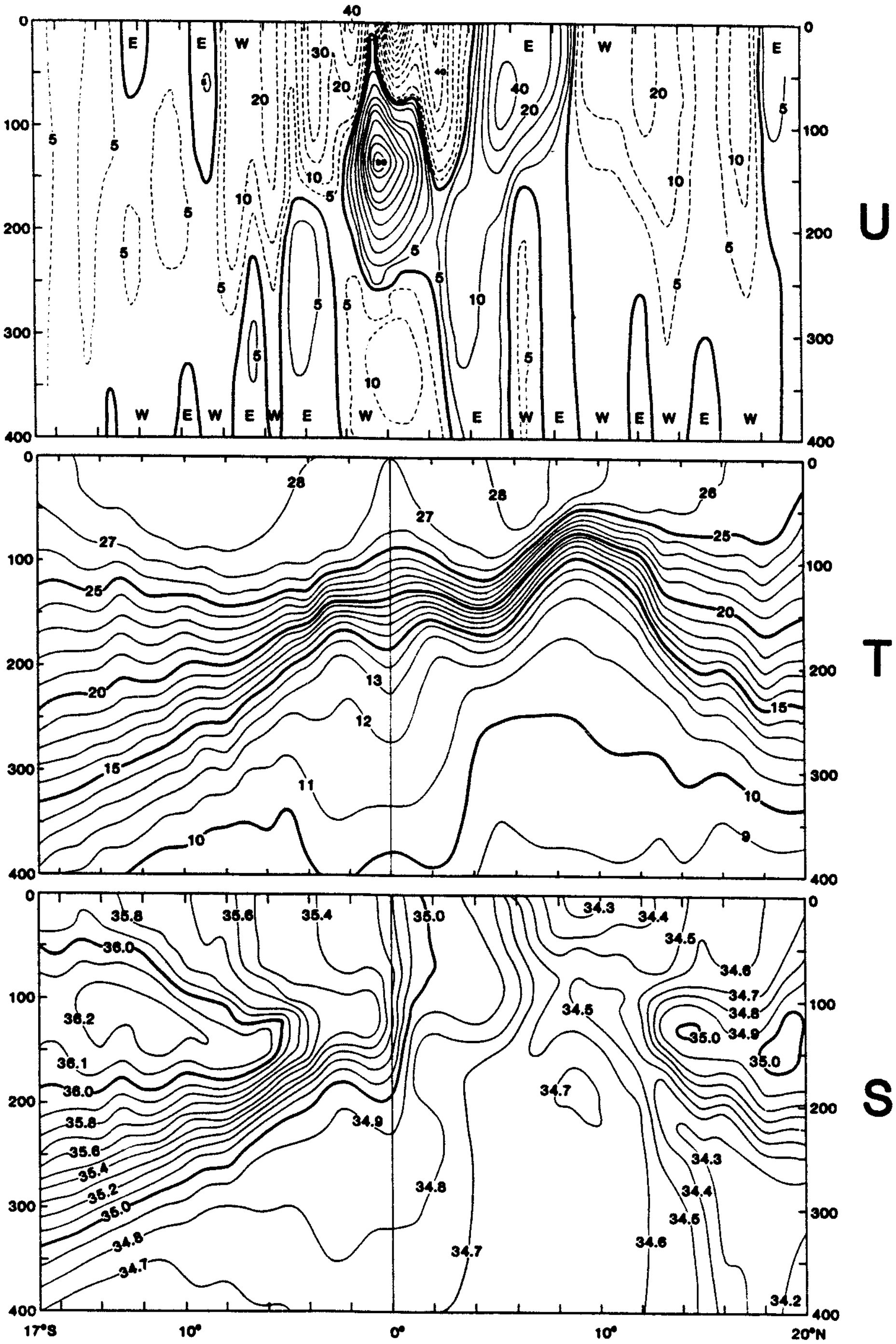

Figure 6.4: Meridional section of U, T, and S from the Hawaii-Tahiti Shuttle along 150–158°W (after Wyrtki and Kilonsky, 1984).

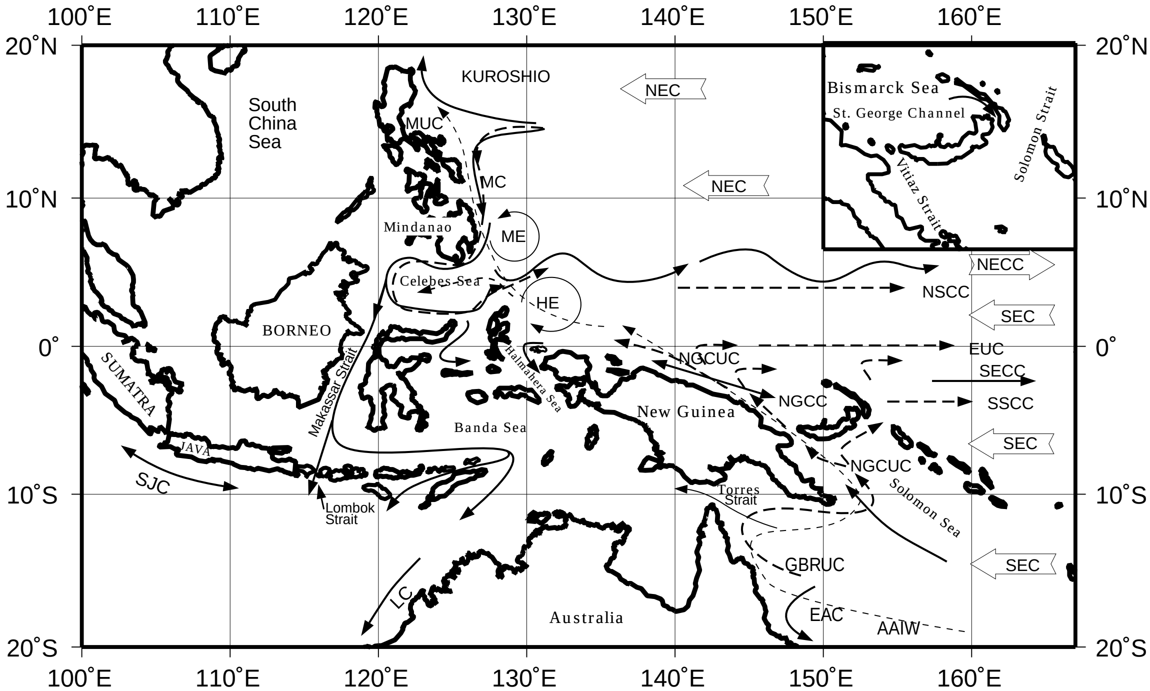

Figure 6.5: Schematic of the low-latitude west Pacific currents (after Fine *et al.*, 1994). Of interest here are the North Equatorial Current (NEC), Mindanao Current (MC), North Equatorial Countercurrent (NECC), New Guinea Coastal Current (NGCC), and New Guinea Undercurrent (NGUC), and the complex flow from south of Mindanao through the Indonesian Seas.

must be given to the low-latitude boundary currents. Figure 6.5 shows the main surface and shallow subsurface currents in the western tropical Pacific.

6.2.1 The Northern Hemisphere

Although the zonal-average picture of the subtropical cells, and the general downward slope of the thermocline to the west, suggests a large-scale geostrophic inflow to the equator, a mean upwelling favorable windstress curl in the ITCZ produces a thermocline ridge near 8–10°N (Fig. 6.4). This forces most of the equatorward flow originating in mid-latitudes to the western boundary in the North Equatorial Current (NEC) (see Lu and McCreary, 1995; Johnson and McPhaden, 1999; and Fig. 6.6). The NEC is both the southern limb of the northern subtropical gyre and the northern limb of a cyclonic tropical gyre. It is a Sverdrup response to ITCZ upwelling curl whose eastward flow occurs in the North Equatorial Countercurrent (NECC). Since the mean wind-stress curl north of the equator has mainly the same sign across the whole basin, both the NEC and the NECC increase transport towards the west, with geostrophic transports at 137°E of about 60 and 40 Sv, respectively (Qiu and Joyce, 1992). The NEC bifurcates at the Philippine coast near 14°N (approximately at the latitude of the zonally averaged zero-curl line), with part of its water flowing north as the Kuroshio into the northern gyre (see Section 8.2) and part flowing south as the Mindanao current into the complex circulation near the equator (Figs. 6.5 and 6.1). This bifurcation can be considered the western edge of the dynamical boundary between the two gyres.

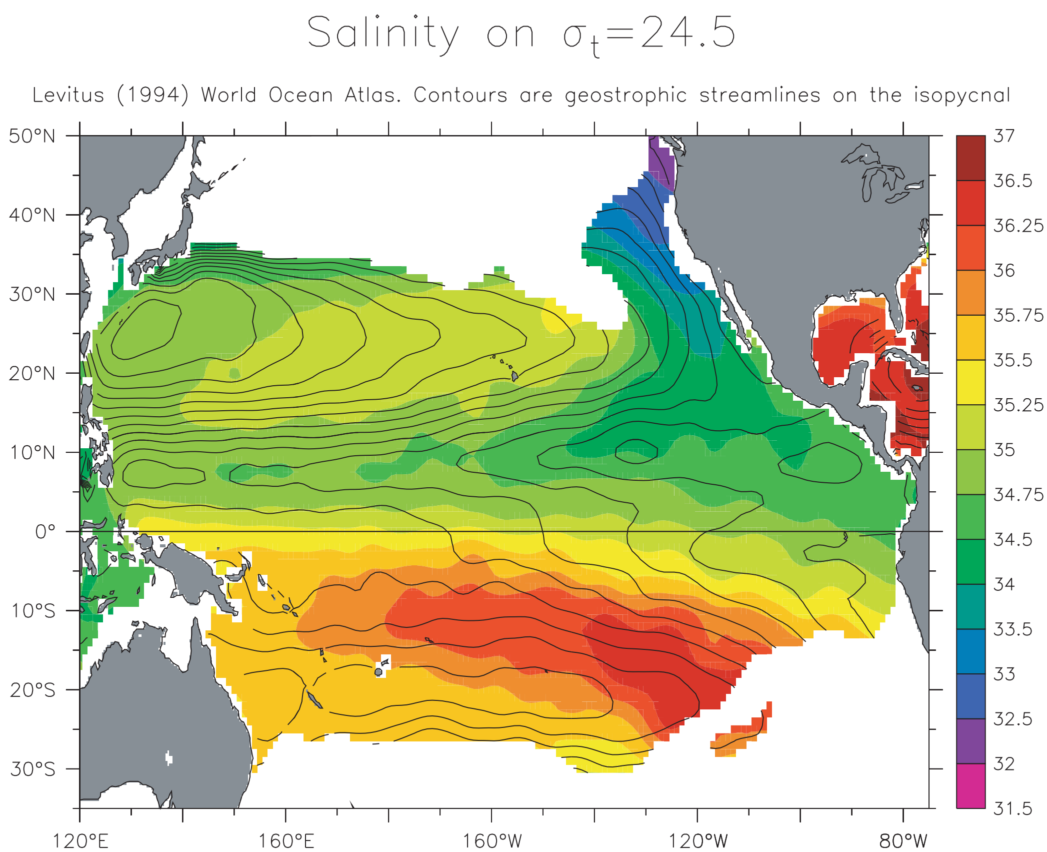

Figure 6.6: Salinity and geostrophic streamlines on the isopycnal $\sigma_t = 24.5$ (after Kessler, 1999) centered in the tropical thermocline. The contrast of salty water from the South Pacific and fresher water from the North Pacific leads to a sharp contrast in the equatorial zone. Off the equator, geostrophic streamlines depict flow. Lack of alignment of these fields may indicate the importance of eddy transport.

Little annual-cycle transport variability is observed in the NEC in the western Pacific (Wyrtki, 1961; Kessler, 1990); however, seasonal variability of the Mindanao Current and the tropical Kuroshio are relatively larger, and are oppositely phased. Qiu and Lukas (1996) attributed these annual cycle transport variations to changes in the bifurcation latitude driven by arriving Rossby waves and also by the annual cycle of Asian monsoon winds. The bifurcation partitions NEC inflow so that when the bifurcation moves to the south, Kuroshio transport increases and Mindanao Current transport decreases, and vice versa. On the other hand, the NEC (and the NECC) are observed to simultaneously increase their transport by up to a factor of two following El Niños when Rossby waves generated in mid-basin lift the thermocline ridge between the two currents and thereby increase the gradient across both of them (Qiu and Joyce, 1992). (The west Pacific NEC also shows interannual variability that is not associated with the ENSO cycle.) Both the Mindanao Current and tropical Kuroshio transports vary in phase with those of the NEC on ENSO timescales. These annual and interannual variations suggest that the NEC and its bifurcation are a mechanism by which low-latitude variability originating in midbasin can penetrate to the midlatitude North

Pacific and to the far western equatorial Pacific (and thus possibly to the Indonesian throughflow as well).

The Sverdrup transport of the NEC-NECC tropical gyre is about 30–35 Sv (Kessler and Taft, 1987), somewhat larger than most estimates of the transport of the Mindanao Current that closes the gyre, which are typically about 20–25 Sv (Lukas *et al.*, 1991). Complicating the linear Sverdrup picture is the recirculation associated with the cyclonic Mindanao Eddy a few hundred kilometers offshore (Fig. 6.5). Also, some of the Mindanao Current flows into the Suluwesi Sea and passes into the Indonesian Throughflow (Godfrey, 1996), which suggests that the Mindanao Current may partly be driven by the global thermohaline circulation.

The Mindanao Current is observed to flow close to the coast (usually within 100 km) with near-surface speeds above 1 m/s, and speeds of 10 cm/s or more extending to a depth of at least 700 m (Lukas *et al.*, 1991). At thermocline level ($\sigma_\theta = 24$), the Mindanao Current is seen as a tongue of relatively low salinity extending towards the equator. At $\sigma_\theta = 26$–27, the Mindanao Current is the path by which even lower-salinity intermediate water, formed in the far northern Pacific, exits the North Pacific (Bingham and Lukas, 1994). As the current passes south of the island of Mindanao (at about 5°N) drifter trajectories suggest a complex and probably highly temporally variable flow field in which some near-surface water flows directly into the NECC, some into the Suluwesi Sea and then out again and into the NECC, and some into the Indonesian passages (Lukas *et al.*, 1991; see Fig. 6.5). Also in this region the New Guinea Coastal Current exiting the South Pacific along the coast of New Guinea retroflects into the anticyclonic Halmahera Eddy and also passes eastward. Model solutions (Gu and Philander, 1997, among others) show that Mindanao Current thermocline-level water that flows eastward into the NECC moves into the EUC within a few thousand km (Fig. 6.1), and this is borne out by the sharp equatorial salinity front (Fig. 6.4) in which the fresh water on the northern side of the EUC has a North Pacific origin.

Measuring the Mindanao Current and the flows that feed the equatorial circulation from the western boundary is sure to be challenging. Moorings are the best proven, if expensive, approach. The steep and rugged bottom topography off the Philippines coast (the Mindanao Trench is more than 9000 m deep less than 80 km offshore), and the rapid current speeds over a long vertical extent means that moorings in the center of the current may not be feasible. Surface drifters have proven useful in resolving many of the complex features of the surface currents (Lukas *et al.*, 1991) and, since the area is near to the coast, regular reseeding of drifters should be relatively easy. Because of the climatic importance of thermocline-level flows that feed the equatorial circulation, sampling of water properties and transport through the thermocline will be needed. This will not be done well by Argo floats (Section 4.2), since they will not provide the spatial resolution needed (order 10 km), and there are no known opportunities for frequent VOS crossings. Autonomous gliders (Section 4.4) or moored profilers might collect repeated T and S profiles on either side of the current, thereby giving good measures of geostrophic transport, and gliders might observe sections through the currents. Coastal sea-level gauges have promise for monitoring the geostrophic flow of the Mindanao Current between Palau Island and Mindanao (Lukas, 1988), possibly in conjunction with inverted echo sounders to better sample the recirculation. Lukas (1996) suggested that acoustic tomography would be appropriate for monitoring fluctuations of NEC transport in the far west and its bifurcation. A careful study of the options is needed.

6.2.2 The Southern Hemisphere

Although much of the discussion of oceanic teleconnections from the subtropics to the equator has concerned the northern hemisphere, the main source of the South Equatorial Current (SEC) and the Equatorial Undercurrent (EUC), and a significant source of the North Equatorial Countercurrent

(NECC), are in the southern hemisphere. The southern counterpart to the NEC-NECC tropical gyre extends across the equator and includes as its westward limb the northern portion of the SEC and as its eastward limb the southern portion of the NECC (and the EUC at thermocline level). The observed transports of this gyre are reasonably well predicted by Sverdrup dynamics (about 10 Sv) (Kessler and Taft, 1987). Just as the Mindanao Current closes the northern tropical gyre by connecting the NEC to the NECC and EUC, in the southern hemisphere a western boundary path carries water from about 15°S, where the SEC bifurcates at the Great Barrier Reef, along Australia and the south coast of New Guinea, and then northwestward along the northern coast of New Guinea to the equator (cf. Lukas *et al.*, 1996). Figure 6.5 shows the western boundary connections of this circulation.

Along the northern New Guinea coast this equatorward transport is carried partly by a surface flow (New Guinea Coastal Current, NGCC), which seasonally reverses in the tradewind/monsoon wind cycle, and beneath this, focused near 200 m, by the New Guinea Coastal Undercurrent (NGCUC), which consistently flows toward the equator (Lindstrom *et al.*, 1990). Lindstrom *et al.* (1987) found that at 143°E the cores of the NGCUC and EUC had comparable transports of near 7 Sv, and inferred from water properties that the NGCUC was the main source of the EUC. Tsuchiya *et al.* (1989) traced the NGCUC waters back to a high salinity pool near 20°S and, confirming Tsuchiya's (1968) hypothesis, showed that it is the main source for the EUC where it originates west of 140°E. Tsuchiya *et al.* (1989) also identified a broad northwestward flow north of the Solomon Islands that is entrained into the EUC near 150°E, beginning the increase of the EUC's transport as it flows eastward (Fig. 6.1). By the time it reaches the central Pacific, the EUC has a transport of 30–35 Sv (Lukas and Firing, 1984). Its sharp salinity front (Fig. 6.4) shows the distinction between the southern high-salinity water and the northern low-salinity water that combine to make up the EUC.

The climatic significance of the New Guinea Coastal Undercurrent is its role in the shallow subtropical overturning circulation. The NGCUC transports subtropical water to the equator where it is carried to the east and eventually upwells. The upper part of the EUC, which is fed by the NGCUC, is the immediate source of the water brought to the surface by equatorial upwelling. This upwelling is a primary mechanism modulating SST where the atmosphere is most sensitive to SST. Thus the NGCUC is a conduit through which southern hemisphere oceanic variability can be carried to the equator where it can affect the atmosphere, and perhaps contribute to the modulation of the ENSO cycle (Section 6.4). Comparison of measurements of the transport of the NGCUC made by Lindstrom *et al.* (1990) over 6 months and the 1-year record obtained by Murray *et al.* (1995) indicates there are large interannual variations. In order to determine if the southern hemisphere can modulate ENSO through changing the mass, heat, and/or freshwater transports of the NGCUC, these transports must be measured for more than a decade.

The New Guinea Coastal Current and Undercurrent pass through a region of great bathymetric complexity and wind forcing that varies seasonally (with southeasterly trades peaking in July and westerly monsoon winds peaking around January) and interannually. Peak speeds in both the NGCC and NGCUC, which are roughly 200 km wide north of New Guinea, are in the range 50–100 cm/s. The latitude is low, so geostrophic calculations are noisy, and it would be desirable to extend observations across the equator, where direct velocity measurements are required. The site is remote with minimal local logistic support. Consequently, observing changes in the mass, heat, and freshwater transports at a cost that can be sustained will be challenging. On the basis of information gathered in the WEPOCS experiment (Lukas *et al.*, 1996; Lindstrom *et al.*, 1987, 1990; Tsuchiya *et al.*, 1989), two complementary strategies present themselves.

Off the east coast of New Guinea the path of the New Guinea coastal currents is interrupted by New Britain and New Ireland islands and the currents are forced though two narrow passages

(Fig. 6.5, inset). The main path, through Vitiaz Strait, is confined to a 40-km passage bordering New Guinea, while there is a smaller flow through St. George's Channel, a 30-km-wide gap between New Britain and New Ireland. These confined flows may be suitable sites for surface moorings (blow down of subsurface moorings makes them unsuitable for capturing the shallow flows). These passages might also be suited for measurement by pairs of moored profilers, pairs of pressure gauges or, if suitable ferries or inter-island packets could be found, repeated observation by XBT and/or ADCP. The 100 cm/s speeds found in Vitiaz Strait and island impediments to working back upstream probably preclude using autonomous gliders, and the current strength may also be a problem for moored profilers.

An alternate strategy employed in WEPOCS is repeated sections north from New Guinea somewhere between 143 and 148°E. Autonomous gliders sampling temperature and salinity could monitor the New Guinea Coastal Current and Undercurrent on a monthly basis using geostrophic calculations referenced by glider drift. (Lindstrom *et al.*, 1990, report general success but some problems with geostrophic calculations.) It would be desirable to add an ADCP to the glider and extend such sections north to capture both the Equatorial Undercurrent and its NGCUC source in a single section (cf. Fig. 1 of Lindstrom *et al.*, 1987).

6.3 Indonesian Throughflow

Transport of water from the Pacific to the Indian Ocean significantly impacts both global climate and diagnosis of climate processes occurring in the Pacific. As shown in Fig. 6.5, the convergence of the Mindanao and New Guinea western boundary currents feeds eastward equatorial flow in the Pacific Equatorial Undercurrent and surface and subsurface countercurrents. This convergence also feeds a complex flow westward and southward through the Indonesian Seas, which enters the Indian Ocean between Australia and the Indonesian Archipelago (the Indonesian Throughflow (ITF)). The mean mass, heat, and freshwater transports of the ITF are poorly known (Bryden and Imawaki, 1999; Wijffels, 1999), and the variability is high, but Godfrey (1996) estimates that the net transport of about 5 Sv leads to a net heat transport of about 0.5 pW from the Pacific to the Indian Ocean.

While the ITF is a fascinating oceanographic problem on its own, from the CLIVAR perspective there are three main reasons to sample at least its mass, heat and freshwater transport on a seasonal timescale:

- The ITF may affect Australasian climate. Consistent with the ITF's considerable heat transport, modeling by Hirst and Godfrey (1994) shows that weakening the Throughflow results in a distinct pattern of SST change that spans both the South Indian and Pacific Oceans. Roughly, a stronger Throughflow warms the subtropical Indian Ocean and cools the eastern South and equatorial Pacific. Allan *et al.* (1995) found similar patterns of SST variability in 20-year epochs over the 20th century, suggestive of a "strong Throughflow" state since the 1940s. While rainfall in northeastern Australia covaries with ENSO, western and southern Australia appear to be more closely tied, on interannual timescales, with Indian Ocean SST (Nicholls, 1989). Possible links between decadal SST changes and the recent 40-year trend of decreasing rainfall in southwestern Australia is under investigation. Although apparently by different mechanisms, the warm Indian SST and cold Pacific SST parts of the SST pattern resulting from ITF variability have reinforcing effects on the Australasian monsoon (Soman and Slingo, 1997).

- The ITF may be involved in ENSO and its predictability. An Indian Ocean precursor for initiation of El Niños has been suspected but not yet identified. Historical studies have clearly shown evidence of precursor anomalies in the Indian Ocean (from the surface up through the troposphere), but they are months prior to the onset of surface warm conditions in the Pacific (e.g., Harrison and Larkin, 1998) and no mechanism for such a slow coupled mode has been offered. Coupled modeling work by Schneider (1998) reveals that the ITF heat transport has a substantial impact on global climate. When the ITF moves heat from the equatorial Pacific to the Indian Ocean, deep atmospheric convection is moved westward with it. This shift drives changes in the global atmospheric circulation and affects mid-latitude winds. Coupled with Meyers' (1996) observations that variations in the thermal structure and transport of the ITF are linked to large-scale wind variations over the tropical Pacific and Indian Oceans, this opens the possibility of feedback loops in which the ITF transport affects ENSO. While this is highly speculative, CLIVAR must address the question of whether variability of ITF transport might not play a role in the growth and dissipation of El Niños.

- The ITF affects analysis of Pacific climate dynamics. Because of its topographic complexity, it is unlikely that climate models will soon be accurately predicting flow through the ITF. Thus ITF transports will enter PBECS assimilation analyses mainly through observations. Experience with GCMs (Hirst and Godfrey, 1993; Schneider, 1998; Rodgers *et al.*, 1999b) and with analysis of trans-Pacific hydrographic sections (Wijffels *et al.*, 1999) shows how important this transport is in understanding the climate-relevant circulation of the Pacific. Without measurements of the ITF transports it will be very difficult to diagnose variability of the heat budget of the western tropical Pacific on climate timescales.

Measuring the transport of the ITF is considered in depth by Imawaki *et al.* (1999). It is complicated by complex bathymetry, nearness to the equator where direct current observations are required, and strong variability on 40- to 60-day periods. While current meters and moored profilers may be required to apportion transport between the various inter-island passages, and to understand water-mass transformation during this odyssey, measurements of the net inflow between the Philippines and New Guinea and of outflow between Indonesia and Australia appear to be the most effective ways to measure inter-ocean transports. Hiroshima University has implemented a repeated XBT and ADCP section between Japan and Australia. The XBT/XCTD line IX1, which coincides with hydrographic lines during JADE and WOCE, has produced a geostrophic estimate of the upper ITF transport and heat flux since 1984, and is logistically attractive in providing VOS crossings roughly fortnightly.

Australian scientists now oversee low resolution XBT sampling on the IX1 line roughly every two weeks. The main limitations with the present sampling are:

- **Lack of synoptic salinity coverage.** This is a particular problem on the northern end of IX1, where variability is large. Fieux *et al.* (1996) report that use of climatological salinity data in the upper 800 m leads to transport errors of up to 5 Sv, largely because of the highly variable Java Current.

- **Limited depth of sampling**. Present XBT data cuts off at 750 m, too shallow to capture the entire ITF.

- **No direct velocity observations**. Ekman transport and any other ageostrophic flows are not captured by the present sampling.

An effective way to address the Indonesian Throughflow would be for the U.S. to cooperate with Australia in upgrading the IX1 VOS line by providing the following: deep XBTs to extend depth coverage; XCTDs, underway CTDs, and/or thermosalinographs to observe the salinity needed to make accurate geostrophic transports and to measure freshwater transport; and an ADCP on at least one ship to calibrate the geostrophic calculations and measure the Ekman transports of fresh water and heat.

6.4 The equator

Equatorial upwelling is the path by which the water in the subtropical overturning circulation returns to the surface along the equator, where it can be Ekman-advected back to the subtropical gyres (see Sections 2, 6.1 and 6.2). Upwelling is a central ingredient in the ENSO cycle. Indeed, model calculations (see Section 2.4) suggest that, because positive feedbacks between the equatorial zonal SST gradient and zonal winds can amplify small perturbations, changes in the properties of water upwelled at the equator can lead to significant changes in the ENSO cycle. Since the water that upwells from the equatorial thermocline was formed by surface processes in mid-latitudes and advected along isopycnals through the gyre, the possibility is raised that anomalies (either of temperature or volume transport) from the subtropics could be transmitted via oceanic pathways to the equatorial zone and influence interannual variability there.

Potential vorticity constraints inhibit water parcels from crossing the equator except in the presence of strong friction or mixing. This is often thought to occur mainly at the western boundary. But as Blanke and Raynaud (1997) point out, in OCGMs water upwelled to the surface at the equator is then redistributed by the surface Ekman transports, making upwelling a likely mechanism of interhemispheric exchange.

Complicating the picture of basin-scale thermocline-level convergence and Ekman divergence, Lu *et al.* (1998) found two meridional-vertical cells in their layer-model solution: the subtropical cells (Sections 2.1 and 6.1) that transport water over many degrees of latitude, and a smaller, shallower cell that recirculates warm near-equatorial upper water within about $\pm 5°$ latitude. GCM solutions reported by Blanke and Raynaud (1997) suggest that geostrophic convergence in the interior ventilates the upper EUC (Lu *et al.*'s shallow equatorial cell), while lower EUC water enters near the western boundary and exits into the east Pacific cold tongue.

The mean temperature signature of equatorial upwelling is seen in the upwardly bowed isotherms above about 20°C (at about 150 m depth) in Fig. 6.3, which suggest that upwelling is confined within about $\pm 2°$ of the equator. In the upwelling region, cool upwelled water rises to near the surface. It absorbs heat from above that has been estimated, from quite different data and assumptions, to be 50 to 100 W/m^2 (Bryden and Brady, 1985; Weare *et al.*, 1981), with an annual cycle amplitude on the order of half that (Wang and McPhaden, 1999). The water then diverges from the equator, mixing laterally with sun-warmed surface waters and vertically through the seasonal thermocline. Some of it may subduct into the shallow tropical overturning cells while the rest is warmed until it is no longer recognizable.

A simple view of equatorial upwelling was investigated by Bryden and Brady (1985) and Brady and Bryden (1987), who constructed a mass balance in a central-eastern Pacific box based mostly on hydrographic data, geostrophy, and Ekman dynamics. They estimated that at least 70% of the vertical velocity was associated with flow along isotherms (or, nearly equivalently, along the trajectory of the EUC) in the upward-sloping thermocline of the east-central Pacific. In this view, SST in the east Pacific Cold Tongue is primarily a horizontal cut across the thermocline rising to the surface (as opposed to being due to westward advection from the Peru coastal upwelling region).

Qiao and Weisberg (1996) measured the upwelling speed at 0°, 140°W from the divergence of the horizontal velocity components at moorings separated by 2° latitude and 4° longitude during 13 months of the Tropical Instability Wave Experiment. Their estimate, as well as most others (including those based on Ekman transport from observed winds) (Wyrtki, 1981; Wyrtki and Eldin, 1982) indicate upward speeds of one to a few m/day above the EUC core, but rather different patterns of flow are deduced from the different analyses. From the divergence of surface drifters Poulain (1993) found upwelling to be confined within a few tens of km of the equator, implying that upwelling of several m/day occurs in narrow filaments.

From a climate perspective, it is upwelling's role in determining SST that is important. Upwelling is both a local response to surface winds and a component of the gyre-scale circulation. Each aspect affects SST. The local wind probably determines how much water upwells while equatorially trapped waves and the general circulation advection determine what water is upwelled. The source of the upwelled water determines, in part, the SST. Indeed, the relationship between SST and thermocline depth used in simple ENSO models (Cane and Zebiak, 1987) is shorthand for a complex heat budget in which the upwelling temperature-transport (wT) is the main cooling effect. Since there is variation in how different ENSO cycles develop, and in how well specific models predict them, it appears essential to understand what varies in this full heat budget between different ENSOs.

One change may be in the upwelling volume transport—the w in wT. With better satellite wind fields, TAO and Argo large-scale T and S fields, and a few moored current observations, it may be possible to describe the upwelling transport on latitude scales of a few degrees and longitude scales of 10 to 20°. The current meters would be most useful on and near the equator, at a few longitudes where they could define the large-scale zonal currents. Other current meters close to the surface, off the equator, could measure transport and vertical shear in the Ekman layer.

To evaluate the effects on SST of changes in the equatorial thermocline water properties, it will be necessary to understand how thermocline water that has been carried into the equatorial undercurrent from the subtropics reaches the surface, how it is modified in the strong shears of the equatorial zone, and how the processes that bring it to the surface are affected by equatorial winds. For this purpose, simultaneous observation of the surface wind, and of vertical profiles of the horizontal velocities, temperature, and salinity at locations close enough to estimate meaningful gradients, would be most useful. Again, mainly a combination of improved winds, moored velocity profiles, and higher resolution T and S sampling are likely to be the main observing elements. Other techniques that have proven valuable in previous work might be included. Surface drifters can give high horizontal resolution velocity measurements, but tend to be quickly advected away from the equator (Poulain, 1993) and do not simultaneously measure the properties of the upwelled water. ADCP and CTD profiles taken during regular servicing of the TAO array can provide high meridional resolution, if infrequent, pictures of velocity and water properties.

Ultimately, however, the greatest hope for unraveling the equatorial processes affecting ENSO lies in exploiting models to both blend data of different types and to fill in the unmeasured. For example, combining the randomly positioned 10-day samples from Argo floats with the sparse but fixed-point and well-sampled in time TAO data may best be done inside an assimilation framework. Even more, vertical velocity will only be inferred indirectly from disparate data sources, something that almost demands assimilation. Of course, assimilation will be increasingly effective as the underlying dynamical models include better parameterizations of unresolved processes (see Section 9), are driven by better surface forcing (Section 5), and can run at higher resolution.

Much information is needed before an efficient system for sustained observation of the processes that govern SST in the equatorial zone can be implemented. The first step would be a well-designed process experiment with a broad range of techniques employed both to understand the component processes and to determine where and what type of observations would be most useful, and whether and how upwelling or its variability might be inferred from other types of sustained observations. Such a process study is discussed in Section 9.4.

7. Extratropical Climate Feedbacks

An oceanic link between the tropics and subtropics may be responsible for Pacific decadal variability, but it has also been hypothesized that processes confined to the extratropical North Pacific can drive decadal variability. While it is uncertain that the atmosphere is responsive to the extratropical ocean surface on seasonal timescales, it is much more likely that this coupling is active on decadal timescales. As described in Section 2.2, there are concrete hypotheses, confirmed by model runs, for decadal feedback loops in the extratropical North Pacific. PBECS must observe and describe these feedback processes in order to test the hypotheses. While there are no analogous hypotheses for climate feedbacks in the southern hemisphere, the physics of the two hemispheres are not so different that we can afford to let the South Pacific continue to be as sparsely observed as it is today.

7.1 The North Pacific feedback loop

There are three key coupled processes that figure in models of a North Pacific climate feedback loop: oceanic thermocline adjustment to changing wind-stress curl, SST response to the changing thermocline, and atmospheric response to changing SST. While elements of this loop are observed in the North Pacific, detection of the full feedback process will be difficult. Here we discuss approaches to the problem.

7.1.1 Basin-scale thermocline and current fields

Argo will provide excellent coverage of basin-scale thermocline variations in the subpolar and subtropical gyres. Surface drifters can provide information on the surface currents and SST in the subpolar and subtropical gyres. The Oyashio and Kuroshio boundary currents are important elements in the general circulation but will not be well observed with floats or drifters. Because boundary currents are so difficult to observe, it would be very helpful if past and future measurements of hydrography on the many lines off the East Asian coast were made available for scientific use.

Japanese scientists have aggressively expanded climate-relevant observations of Kuroshio transport. Imawaki *et al.* (1999) reviews methods for sustained observations of the Kuroshio transport. They report an extensive 3-year program of current meters (33) and repeat hydrographic sections (42) to directly measure changes in the Kuroshio transport and compare them to sea surface height differences from TOPEX/POSEIDON. The result is a tight linear regression such that SLH slope determines transport to within 5.6 Sv over a variability range of 60 Sv. While the Oyashio has not yet been “calibrated” in this fashion, there is reason to expect that this will be accomplished by international partners, thereby converting the long altimeter record into a record of Oyashio transport.

With Argo, extension of drifters to the subpolar gyre, and continued high-quality altimetry, the thermocline and boundary current parts of the North Pacific feedback loop will be adequately monitored.

7.1.2 Causes of SST variation

The region of maximum anomalous extratropical low-frequency large-scale SST variability extends roughly from 150°E to 160°W and 35 to 45°N (cf. Fig. 2.1). The western part of this region includes the Kuroshio-Oyashio extension (KOE) that figures prominently in theories of decadal variability (see Section 2.2). SST anomalies (SSTAs) over the larger area are correlated with U.S. winter weather patterns. Indeed, only the ENSO SSTA association is as strong.

The modeled North Pacific decadal feedback loop depends on ocean-atmosphere coupling that seems most likely to occur in the KOE, which is the site of the largest interannual-to-decadal SST variability in the Pacific (cf. Deser and Blackmon, 1995). Several recent studies have shown that this mid-latitude decadal SST variability has characteristics distinctly different from the tropical decadal ENSO signals (e.g., Miller *et al.*, 1994; Nakamura *et al.*, 1997; Zhang *et al.*, 1997). Cayan (1992b) and Iwasaki and Wallace (1995) both found that the time rates of change of SST anomalies and anomalous surface heat fluxes in the WBC outflow region are poorly correlated, suggesting the importance of the internal ocean dynamics in dictating the long-term SST changes. Farther to the east, seasonal variability is apparently driven by the atmosphere but coupled phenomena are more likely at lower frequencies.

PBECS must learn the processes that control SST evolution in this key region of the Pacific Decadal Oscillation (PDO). Devising a method to track SST changes in the KOE, and the heat transport processes that bring them about, will be difficult. The Kuroshio Extension System Study (KESS, described in Section 7.2 below) is a process experiment focused in the KOE that can provide a foundation for understanding the complicated oceanic processes that change SST in the KOE. Farther east the processes are evidently simpler and the scales larger, so design of a measurement array should be simpler. Yet to be learned is where SST is determined mainly by air-sea fluxes forcing one-dimensional mixed-layer processes, where large-scale currents contribute to SST anomalies, and how thermocline depth affects these. By combining various observational methods (TOPEX, moorings, tomography, in situ temperature and salinity profiles, etc.) it will be possible to devise an array from which we can learn, for example, if SST anomalies are formed in the KOE near the coast and then extend eastward under the force of ocean advection or by a coupled air-sea process.

While design of a full climate observing system for the North Pacific must await completion of KESS, it would be useful if the KESS array were supplemented with an eastward extension of buoys since it is presently rather localized in the eddying region of the KOE. These would focus on air-sea fluxes, elements of one-dimensional processes in the mixed layer, thermocline structure, and horizontal advection.

NOPP-sponsored mid-latitude moorings have been maintained at the Ocean Weather Station P site (50°N, 140°W) since October 1998 and in a data sparse region NNW of Hawaii (35°N, 165°W) since October 1999. The OWS P site is one of the few places in the open ocean with a multi-decadal record of upper ocean and air-sea flux observations. Decadal changes have been documented by Canadian oceanographers but recently the frequency of Canadian sampling is barely seasonal. Sustaining an observing effort at this site will provide the ability to understand how conditions in the first decade of the new millennium compare with those during the 1970s. The site NNW of Hawaii is near the maximum of PDO SST variability. It is the only reference site in the central North Pacific with which to evaluate the accuracy of air-sea forcing fields produced by operational meteorology centers or from NASA and NOAA satellite observations.

Maintenance of these sites and addition of new ones connecting them to the KOE will be very helpful to PBECS modeling and data assimilation studies while we await KESS results to design a full monitoring array.

7.1.3 Atmospheric sensitivity to KOE SST anomalies

Measuring atmospheric sensitivity to SST in the KOE will be a difficult task but several angles can be followed. Long-term surface heat-flux measurements are crucial for establishing ocean-atmosphere feedback processes on decadal timescales. KESS again can provide a base for these measurements, although additional heat flux buoys would make important improvements. Although it is unclear how mid-troposphere atmospheric observations can be made for these timescales over the KOE region, they should be attempted to determine the heating rates and vorticity balances. Otherwise we will need to rely on NCEP or ECMWF analyses to infer the atmospheric response to changing oceanic conditions. A 1-year or one-winter process study of atmospheric response in this region would be an excellent way to validate the NCEP/ECMWF analyses. (This type of experiment will have added benefits of improving weather and climate forecasting for Japan and the U.S.) Downstream feedbacks under the Aleutian Low may also be important in the decadal feedback loops. Surface drifters that measure sea-level pressure and surface currents in the subpolar gyre will help sort out these possible interactions.

7.2 The Kuroshio-Oyashio system

As discussed above, describing the large variations of SST in the Kuroshio-Oyashio Extension (KOE) and the causes of these variations is central to understanding climate feedbacks in the North Pacific. In addition to interannual changes in the energy level of the mesoscale eddy field, both the Kuroshio and the Oyashio undergo large-scale variations in their path and transport on the interannual-to-decadal timescales. These large-scale Western Boundary Current (WBC) variations can be regionally and/or remotely driven by wind and buoyancy forcings. They may also be a result of self-sustained, internal ocean dynamics associated with the recirculation gyres of the WBCs.

The KOE region can be separated into three sub regions: the recirculation region south of the Kuroshio Extension, the Mixed Water Region (including the Oyashio) to the north, and the Kuroshio Extension itself. Although these three regions (shown schematically in Fig. 7.1) are contiguous, they have distinctly different processes governing the regional SST changes.

7.2.1 Kuroshio Extension variations

North-south excursion of the Kuroshio Extension (KE) causes SST anomalies. Long-term satellite altimetric measurements reveal that the eastward surface transport of the KE and its zonal-mean position change coherently on the annual and interannual timescales; a larger eastward transport consistently corresponds to a more northerly position of the KE (Qiu, 1995). Both of these observed changes appear to be related to the strength of the southern recirculating flows of the current system, which may be determined by the following processes:

- Modulation of meanders and mesoscale instabilities that result in modulation of potential vorticity (PV) flux divergences (Jayne *et al.*, 1996) and of PV structure in the KE (Marshall and Marshall, 1992),
- Changes in regional buoyancy and wind-forcing (Huang, 1990),
- Changes in Pacific basin-scale (remote) forcing, exerted through the inflow of the upstream Kuroshio, and/or through Rossby waves downstream (Qiu and Miao, 1999).

7.2.2 Recirculation gyre and subtropical mode water

SST in the WBC outflow region can also be influenced by changes in the mixed layer structures and by the entrainment from below the mixed layer. Changes in the WBCs can impact subduction through mode water formation, and the mode water can serve as a sequestered reservoir for SST changes in subsequent winters. Estimates of the relevant terms by Qiu and Kelly (1993) using

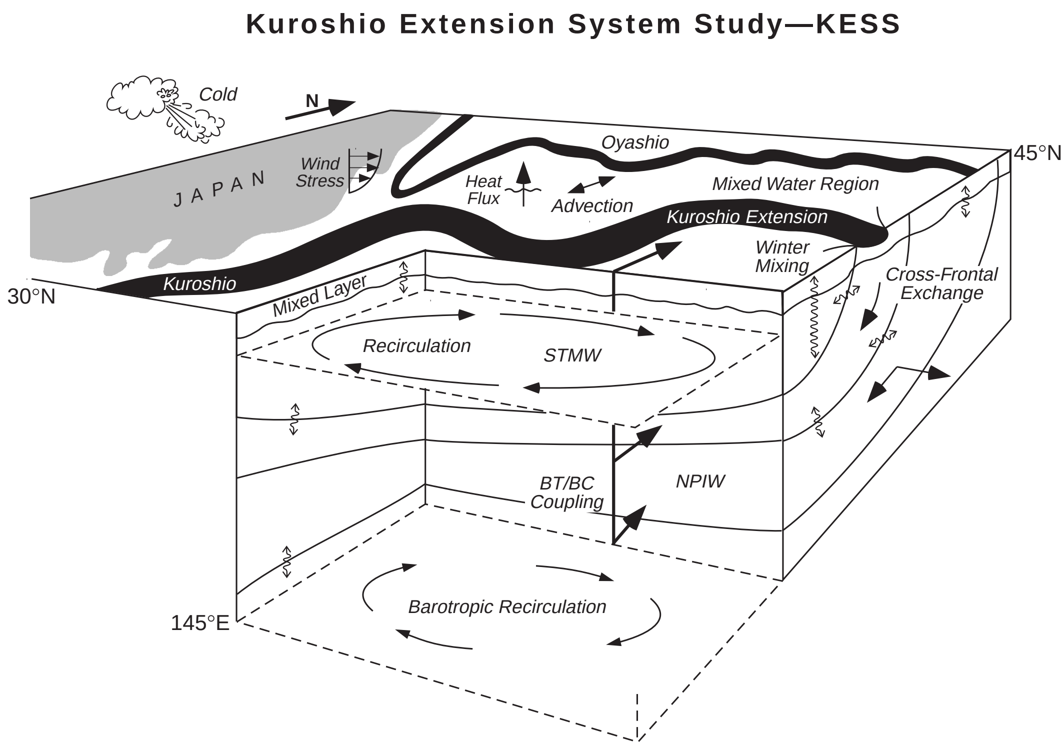

Figure 7.1: Schematic representation of the structure of the Kuroshio-Oyashio Extension mixed water region, including the dominant processes.

Geosat data for geostrophic velocities, showed that entrainment and advection each contribute fluctuations that are about one-third the magnitude of the seasonal variations in net surface heat flux. However, estimates were computed only over a 2-year period and interannual variations in mode water temperature were not included. We need to quantify the relative importance of the following terms: advected temperature anomalies in Kuroshio as basin-scale gyre strength varies; anomalous advection of heat by recirculation south of the KE; variations in net surface heat fluxes (atmosphere driving ocean); changes in mixed layer temperature by vertical entrainment from the slowly varying Subtropical Mode Water reservoir.

7.2.3 Mixed water region (or KOE region)

Decadal/interdecadal climate signals are strongly manifested in the Mixed Water Region to the north of the Kuroshio Extension (Nakamura and Yamagata, 1999). There are also corresponding subsurface temperature signals associated with the thermocline variations (Watanabe and Mizuno, 1994; Deser *et al.*, 1996; Miller *et al.*, 1998). In particular, Miller *et al.* (1998) showed that the thermocline shoaling during the 1980s was westward intensified and may be acting independently from the central Pacific SST variations. However, it is not clear how the SST and subsurface temperature couple with each other. We may hypothesize that the following processes can drive SST variations in the Mixed Water region: thermocline shoaling/deepening associated with wind-driven spin-up; anomalous Oyashio intrusions due to wind-driven spinup and/or anomalous density-driven current component; SST advection by time-varying Ekman transport; anomalously large cross-front exchange from ring-formation and meanders that brings near-surface warm waters to the mixed water region.

7.2.4 What is needed?

Enhanced monitoring and process-oriented studies in the WBC outflow region are crucial to understanding the Pacific decadal variability: how the ocean and the atmosphere are coupled and what determines the SST anomalies on interannual-to-decadal timescales. Diagnosis of the heat budget for the long-term SST changes in the KOE will depend on measurements of the oceanic state and of surface heat fluxes. These should be made for multiple years, covering several of the winter seasons when the air-sea coupling is most active. They should encompass the WBCs, their neighboring recirculation gyres, and the region to the east of the KOE where decadal variability is strong. All of the large-scale sustained measurements in PBECS (e.g., satellites, VOS, floats and gliders, drifters, and moorings) could be used for this. But how to proceed, and how elaborate a system will be needed, is not yet clear.

For the foreseeable future, coupled ocean-atmosphere models will not likely resolve oceanic mesoscale eddies, which affect SST and importantly affect the dynamics and cross-frontal exchanges in the WBC outflow region. Particularly, in-situ process-oriented observation in the WBC outflow region is crucial in three respects:

1. True mesoscale resolution is needed to understand the physics of strongly coupled barotropic-baroclinic eddies and associated cross-frontal structures to help improve parameterizations in climate models.

2. Measurements of the vertical structure of the recirculation gyres are needed to distinguish barotropic and baroclinic circulation changes and changes in the heat content, which are hard for satellites to measure.

3. Results from intensive observation systems can be used to design long-term affordable observing systems for the highly nonlinear, complicated regions.

An international field experiment planned between the U.S. and Japan, called the Kuroshio Extension System Study (KESS), may provide some of the high-resolution observations needed to design a sustainable observing system. A schematic of the proposed array is shown in Fig. 7.2. The following observations are now being planned:

KE and recirculation dynamics. High-spatial-resolution current, temperature, and salinity measurements are required to study eddy-mean interaction, recirculation dynamics, and cross-frontal exchange. An eddy-resolving array combining moorings, inverted echo sounders, acoustic tomography, and RAFOS floats is planned to achieve this ambitious goal. Concurrently, longer-term analyses of satellite SSH observations need to be combined with in situ current observations and tomographic measurements on larger scales to observe the strength and heat content of the recirculation. Regional wind and buoyancy forcing estimates are needed, and NCEP (or other) fields must be calibrated by in situ meteorological buoys. The strength of the inflow should be monitored.

Upper ocean heat budget. Simultaneous in situ measurements of the thermocline depth, integrated heat content, and SST are clearly needed. Lateral/vertical entrainment processes should be measured at the base of the wintertime mixed layer. By assimilating in situ and satellite observations into an ocean circulation model, we can evaluate the mechanisms and the accuracy of a model for the heat budget. Although the measurement period will be short compared with the

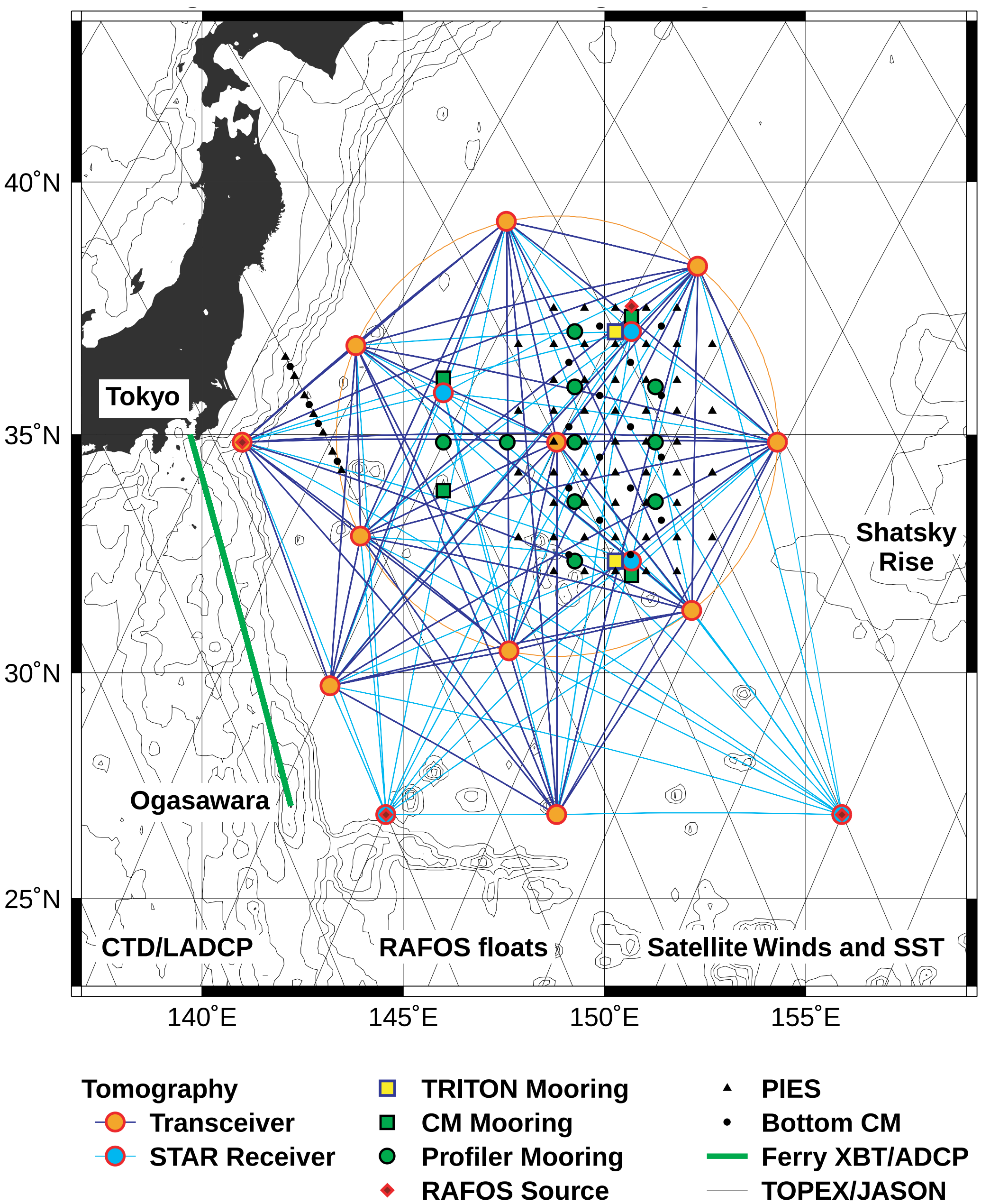

Figure 7.2: The proposed array for the Kuroshio Extension System Study (KESS) array. Blue lines are tomographic paths. Figure does not show proposed RAFOS floats or CTD stations.

decadal timescale, we will have a longer time series of satellite measurements and minimum in situ measurements with which to estimate the heat budget. KESS will provide some of the time series needed to begin studying the dynamics of SST fluctuations in the KOE. More importantly, it will elucidate the processes going on in this complex region so that affordable monitoring systems can be designed. In the meantime, PBECS scientists should consider, in cooperation with Japanese scientists, if there are augmentations to the KESS array that would significantly increase its utility for climate studies. Improved measurements of surface fluxes may be such an area.

7.3 The southern subtropics

The South Pacific subtropical gyre has strong interest for PBECS for several reasons. First, patterns of flow from the south are responsible for establishing characteristics of equatorial current systems (e.g., Johnson and McPhaden, 1999), and, therefore, may modulate the background upon which ENSO evolves, and they are the source of the cross-equatorial transport needed to balance the Indo-Pacific throughflow. Second, there are substantial hemispheric symmetries and asymmetries in the circulations of the North and South Pacific that are important elements of describing and understanding climate variability in the Pacific basin as a whole. Finally, the South Pacific has substantial internal variability on long timescales that are suggested (Sprintall *et al.*, 1995; Peterson and White, 1998) to affect climate on regional or global scales.

Equatorward interior flow in the southern subtropical thermocline is split into two parts. Waters east of 95°W at 20°S directly feed the Equatorial Undercurrent in the central Pacific (Johnson and McPhaden, 1999, Fig. 1) while west of 95°W the equatorward flow feeds the western boundary current system. Approximately equal volumes of thermocline water follow the two pathways, and variability in the system is poorly known.

Several datasets have been used to describe the westward (South Equatorial Current, SEC) transport and the boundary current system that it feeds. These include:

1. Historical hydrography and broad-scale XBT data have been synthesized in a series of papers by Ridgway and Godfrey (1994, 1996, 1997) describing the long-term mean and seasonal-through-decadal variability of currents in the region. The westward flow impinges on the western boundary between 10°S (the southern edge of New Guinea and the Solomon Islands) and 25°S, with transport in the upper 2000 m of about 27 Sv (Ridgway and Godfrey, 1994, Fig. 7). On reaching Australia, the flow bifurcates between 15 and 20°S, a direct analog of the NEC bifurcation in the North Pacific. The northern branch of this bifurcation becomes the low-latitude western boundary current (NGCC and NGCUC) discussed in Section 6.2. The NGCC is apparently fed directly from the SEC and also by flow north of the bifurcation, clockwise around the Coral Sea to the southern end of New Guinea (Ridgway and Godfrey, 1994, Fig. 2). Flow of the SEC south of the bifurcation becomes the East Australia Current (EAC) that, as it proceeds southward along the coast, gains transport and a substantial tight recirculation gyre. The upper layers of the EAC separate from the coast before reaching Sydney, flowing eastward across the Tasman Sea. Part of this eastward flow reattaches to the northern coast of New Zealand, as the East Auckland Current. Deep layers of the EAC continue along the Australian coast as far as Tasmania.

2. High Resolution XBT transects cross the East Australia Current off Brisbane (27°S, PX30 in Fig. 4.10) and off Sydney (34°S, PX34). A third transect (PX6/31) crosses the East Auckland Current as well as the westward flowing SEC. The mean and standard deviation of temperature in these three boundary current crossings is shown in Fig. 7.3. Seasonal to interannual variability in the subtropical gyre is discussed by Morris *et al.* (1996). It was

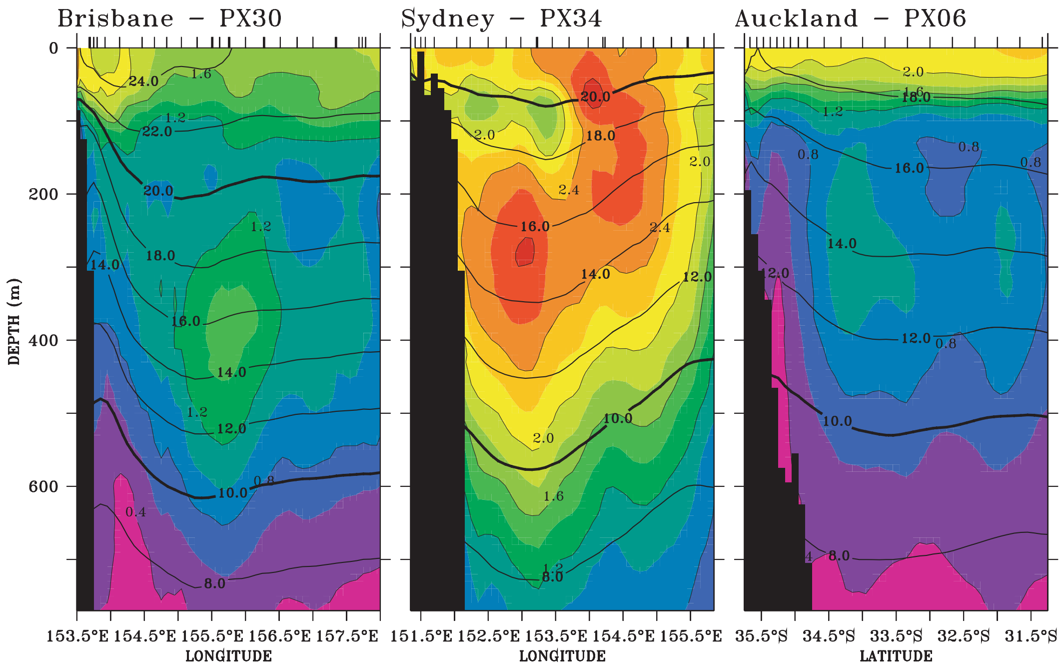

Figure 7.3: The East Australia Current off Brisbane and Sydney (1991–1999) and the East Auckland Current (1986–1999). Heavy contour lines indicate the mean temperature field; colors and light contours show the standard deviation of temperature.

suggested by Sprintall *et al.* (1995), that variability of geostrophic transport in the Tasman Sea is partly responsible for interannual climate fluctuations in New Zealand, including a damaging cold winter in 1992.

3. An array of six current meter moorings (WOCE array PCM3) was set in the EAC at 30°S from late 1991 to late 1993. Mean southward transport in the upper 1000 m was 15 Sv with standard deviation of 14 (Mata *et al.*, 1998). The high variability was attributed to Rossby waves propagating across the array. The array extended through the high velocity core of the EAC but ended only about 70 km from the shelf break and probably did not span all of the poleward flow. Ridgway and Godfrey (1994) estimate the through-transport in the upper 2000 m of the EAC at 31°S to be 17 Sv, with 10 Sv of tight recirculation offshore of the poleward flowing current.

The southwestern Pacific is an integral part of the climate puzzle of the Pacific Ocean as a whole. Little is known about variability of the strong currents, but this variability is clearly large on long timescales. Ridgway and Godfrey (1996) found that the westward inflow to the western boundary decreased by 15 Sv (compared to the 27 Sv mean) between 1975–79 and 1985–89. High variability in the supply of warm water and in its distribution between the equatorward and poleward boundary currents must substantially perturb the oceanic mass and heat balances. A further suggestion of the region's importance in the climate system was the finding by Peterson and White (1998) that coupled processes north of New Zealand initiate the Antarctic Circumpolar Wave. It then propagates southeastward into the Southern Ocean and circles the globe. Moreover, its phase appears linked with the ENSO cycle on the equator.

There is presently high-resolution VOS sampling that crosses the East Australia Current near Sydney and Brisbane and crosses the East Auckland Current north of New Zealand (see Fig. 4.10) on an approximate quarterly basis. Continuing these lines, and extending their depth and salinity coverage with occasional XCTDs and 2000 m XBTs, would be a significant step toward documenting the role of these boundary currents in climate variability. Reducing the aliasing of climate signals by short-term variability would substantially increase the value of these data. This could be done with glider sampling along the three XBT transects, extending from the shelf break out 500 km and sampling roughly once every 3 weeks. If possible, this sampling should include occasional transects upstream and downstream of the primary lines.

8. The Eastern Pacific

The eastern tropical and subtropical Pacific is the site of persistent stratiform clouds in the atmospheric boundary layer that are not well modeled and have an important influence on seasonal and interannual climate. The largest SST variability during ENSO is found in the eastern Pacific and yet we do not understand the heat budget of the cold tongue during either seasonal or ENSO cycles. Perturbations from El Niño events are seen to propagate along the eastern boundary toward higher latitudes, but the classical model of Kelvin waves serves to explain only a small part of this variability. Neither is the impact of interannual variability along the subtropical coasts to the adjacent landmasses understood. For these reasons and others, the eastern part of the Pacific must be included as a special region within PBECS.

8.1 Air-Sea coupling in the eastern Pacific

The climate of the eastern Pacific boundary current regime off the tropical west coasts of Central and South America consists of several related regimes:

- The subtropical southeast Pacific regime characterized by southeasterly trade winds, cool water, and extensive decks of boundary layer stratiform clouds,
- The eastern Pacific warm pool regime off the west coast of Mexico and Central America, and
- The cold-tongue/ITCZ complex (CTIC) in the equatorial region from 10°S to 10°N.

Each phenomenological regime is an important component of the Pan-American monsoon system with significant effects on the climate and weather in Mexico, Central America, and along the west coast of South America. Climate variability in the region is a mixture of strong seasonality created by the Pan-American monsoon and the basin-wide variability of ENSO (Wallace *et al.*, 1989; Deser and Wallace, 1990).

Because the atmosphere's intrinsic predictability time is so short, much of prospective climate predictability around the eastern tropical Pacific, as elsewhere, depends on forcing at boundaries that has the system's memory. The prediction of rainfall over the Americas, the response to ENSO anomalies, and the varying evolution of different ENSO cycles are all climate phenomena that are believed to be influenced by SST distributions in the eastern tropical Pacific. Coupled ocean-atmosphere simulations of the annual cycle of SST and other key climate processes are deficient in the eastern boundary current regime (Mechoso *et al.*, 1995).

The eastern Pacific warm pool to the north of the equator exists in a region of convergent wind-driven ocean currents with intense precipitation over the water and over the neighboring Central American landmasses. Divergent boundary layer winds over the warm pool and the intensity and location of the eastern Pacific ITCZ are associated with summer-season precipitation, and its variability over Mexico and Central America (Magaña *et al.*, 1999). Mixed layers in the eastern Pacific warm pool are shallow, and little is known about the relative importance of air-sea interaction and upper-ocean processes in setting the eastern Pacific warm-pool SST. Upper ocean mixing events involving large air-sea exchanges of momentum, heat, and freshwater are likely to be associated with mesoscale cloud systems, hurricanes, and perhaps synoptic-scale easterly waves crossing from the Caribbean into the eastern Pacific. Upper-ocean advection of temperature and salinity may be more significant than in the western Pacific warm pool.

The strongest regular feature of SST variability in the eastern boundary current regime is the annual appearance of cool water in the CTIC, extending westward from the coast of South America along the equator. March–April is the time of the heaviest rainfall in the equatorial belt; a double ITCZ, symmetric about the equator, is often observed. In contrast, September–October is marked by strong equatorial asymmetry, with the ITCZ in the northern hemisphere and extensive stratocumulus decks in the southern hemisphere. The annual cycle of SST is remarkably regular from year to year, even in the presence of El Niño (Mitchell and Wallace, 1992). The strength and pattern of this annual cycle is not explained by sun-earth orbital geometry, which would instead indicate a semiannual cycle.

The seasonal march of CTIC SST, rainfall, wind, and ocean currents are linked to the Pan-American monsoon, but the coupling mechanisms are still unclear. The oceanic seasonal march near the equator does not follow any simple model for the oceanic response to atmospheric forcing. The usual westward surface flow reverses from April through June even though the wind continues to blow from the east. The eastward flow in the subsurface equatorial undercurrent is at a maximum in May–June following the period of weakest surface easterly wind stress. Notably, the maximum SST in March precedes onset of the period of eastward flow that advects warm water eastward. The notion that a reduction in easterly wind stress reduces upwelling, leading to a deeper thermocline and warmer SST, does not appear to be applicable to the seasonal march. The seasonal variations in thermocline depth and sea-surface temperature do not exhibit the well-defined inverse relation characteristic of the ENSO cycle. Apparently the processes responsible for the SST variations in the seasonal march and in the ENSO cycle may be quite different.

A defining characteristic of the CTIC annual cycle in the eastern boundary current regime is the existence of southerly cross-equatorial flow during the season of coldest SST in the cold tongue (September–October). The boundary of the "southerly" regime, near 110°W, corresponds to the ridge in the equatorial sea-level pressure profile. Westward of 110°W, the zonal pressure gradient along the equator drives easterly surface winds that comprise the lower branch of the east-west Walker circulation.

On the basis of observations and theory, Mitchell and Wallace (1992) hypothesized that the onset of northward winds on the northern fringe of the oceanic equatorial wave guide plays a key role in development of the cold tongue, its self-sustaining nature, and its regularity from year to year. Observations show the meridional wind stress to be associated with the regular seasonal march of deep clouds and precipitation over Central America. Mitchell and Wallace (1992) suggest that the equatorial northward wind stress induces a remote response, similar to the theoretical results of Philander and Pacanowski (1981), bringing cooler water to the surface south of the equator and warmer water to the north. They suggest that a positive feedback involving the formation of relatively high atmospheric pressure over the cooling water may in turn intensify the northward boundary-layer winds and the atmospheric boundary-layer flow into the ITCZ. Through the pressure gradient feedback and the possible positive feedback of boundary-layer stratocumulus clouds, the CTIC would therefore tend to be self-sustaining.

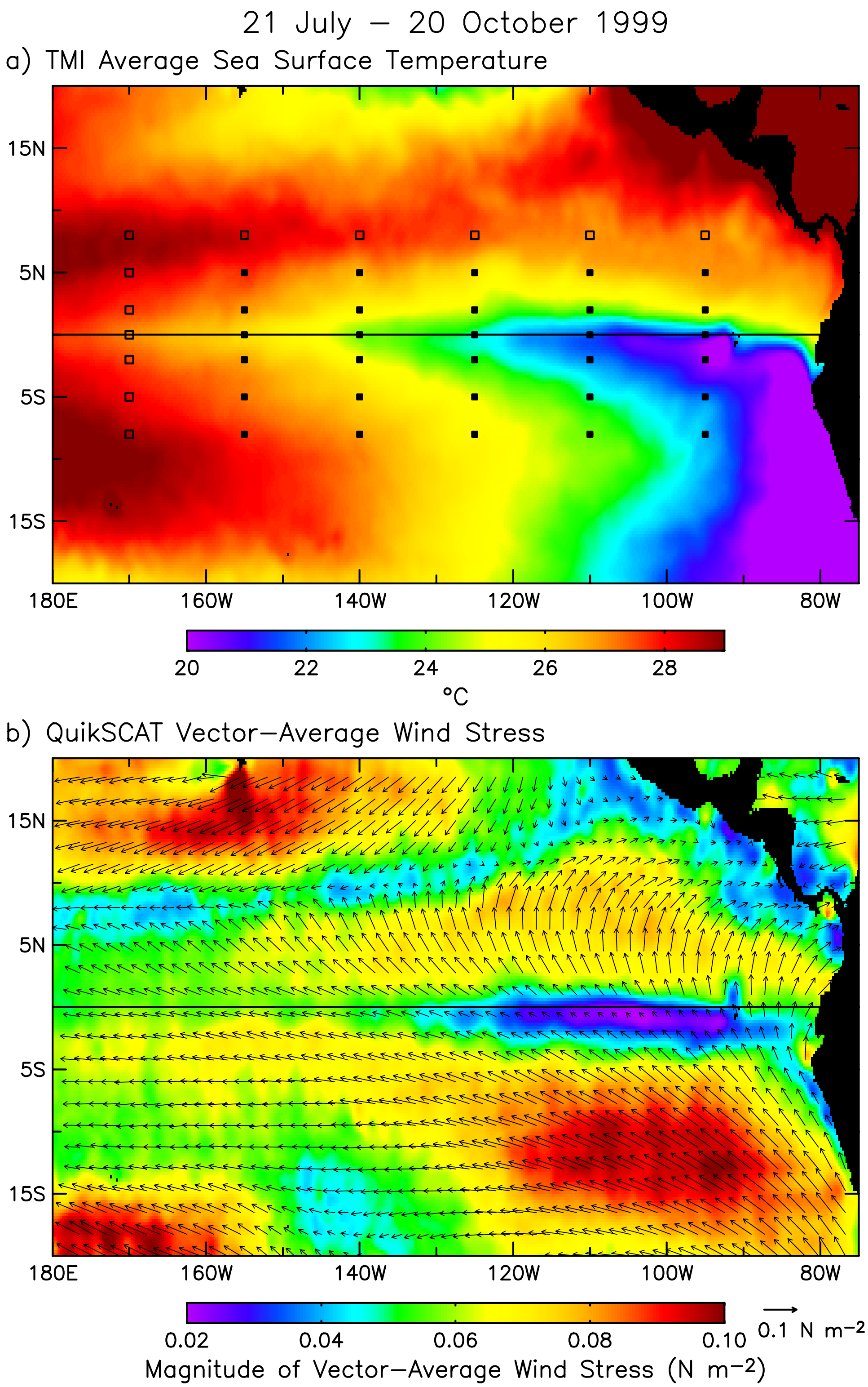

Figure 8.1: Three-month averages (21 July–20 October 1999) over 2° of longitude by 2° of latitude of (a) TMI measurements of sea-surface temperature with the locations of TAO mooring locations shown as squares, and (b) QuikSCAT measurements of surface wind stress.

Several aspects of the Mitchell and Wallace (1992) hypothesis require further study. Recent high resolution wind-stress analyses show that the strongest northward wind stress occurs on the northern fringe of the oceanic equatorial wave guide while the winds over the waveguide are quite weak (Fig. 8.1). This raises questions about whether the Philander and Pacanowski (1981) mechanism can be of quantitative importance in the seasonal onset and maintenance of the cold tongue. The acceleration of the winds to the north of the equator also involve air-sea coupling mechanisms in addition to pressure feedbacks. For example, the acceleration of these winds appears to be associated with vertical convective momentum transport in the atmospheric boundary layer as it is destabilized from below over the equatorial SST front (Wallace *et al.*, 1989). Furthermore, over the equatorial portion of the cold tongue, the ABL becomes relatively stable and cloud free, while north of the equator the ABL becomes unstable and cloudiness increases as air from the southern hemisphere moves over the ST front and approaches the ITCZ. The expected pattern of turbulent heat-flux and surface radiative exchanges could provide a significant negative feedback, tending to weaken the SST gradient.

8.1.1 What can PBECS do?

The eastern Pacific is one of the most data sparse areas over the global ocean. PBECS observations will make substantial contributions to understanding air-sea coupling in the eastern boundary current regimes in the following ways:

- **Upper-ocean temperature and salinity.** PBECS increases in sampling the large-scale temperature and salinity structure of the eastern Pacific are critical to understanding the phenomena of interest. By better defining currents and heat storage, these data will make it possible to validate coupled ocean-atmosphere models and to sort out the mechanisms responsible for the warm pool, subtropical Southeast Pacific, and CTIC phenomena.

- **SST and surface currents.** Stratocumulus clouds cover much of the eastern boundary current regime. In situ observations are sparse, and infrared satellite measurements cannot "see" the sea surface. Surface drifters will provide essential regional "calibration" data for both infrared and microwave satellite SST retrievals. The same drifters will provide direct observations of the surface currents associated with advection of upper ocean heat and SST. Augmentation of current measurements on TAO moorings will add valuable time-series data on upper-ocean currents and the effect of advection on SST.

- **Wind stress and fluxes.** Presently, the primary sources of data for estimating wind stress and air-sea heat flux in the eastern tropical Pacific are TAO moorings, VOS, and satellites. The in situ observations have inadequate spatial resolution to define wind-stress and surface-flux features that are important to the structure and variability of the region, while NWP analyses of heat flux have large biases and inadequate resolution. Continuation of satellite scatterometers and microwave wind measurements is therefore essential, as are the PBECS efforts to improve analyses of stress and surface fluxes.

- **Assimilation studies.** PBECS can support understanding in the eastern tropical Pacific by increasing observations and understanding of the ocean-atmosphere processes responsible for the structure and evolution of the large-scale atmospheric heating gradients in the equatorial and northeastern Pacific portions of the cold-tongue/ITCZ complex. Hopefully, the PBECS approach of long-term observations integrated within a model framework will contribute toward this both directly and by providing a framework inside which process experiments can be particularly effective.

- **Planetary boundary layers.** Progress in understanding air-sea coupling in the eastern tropical Pacific depends on observations and understanding of the dynamical, radiative, and microphysical properties of the extensive boundary-layer cloud decks in the southeasterly trade wind and cross-equatorial flow regime, their interactions with the ocean below, and the evolution of the upper ocean under the stratocumulus decks. These are currently being explored inside the EPIC process studies, which will study the atmospheric boundary layer, upper ocean, and convection along 95°W in 2001.

- **The Future.** Three additional TAO moorings will be deployed along 95°W in 1999–2000 and the surface meteorology on these and other 95°W TAO moorings will be upgraded. Beginning in Fall 2000, a surface mooring to measure surface fluxes of heat, freshwater and momentum, and high vertical resolution upper ocean currents, temperature and salinity will be deployed under the stratocumulus cloud deck. The first year of data from these deployments will be used to examine the extent to which local air-sea interaction explains the evolution of SST and upper-ocean structure, cloud-SST feedbacks, and the extent to which existing coupled models can replicate this evolution. Based on what is learned from the comparison of models and new observations in 2000–2001, more detailed field studies may follow in 2003 or later. More surface moorings will be deployed along the South American west coast by Peru (funded) and Chile (planned). This emerging effort should be coordinated with PBECS and the long-term components included into PBECS growth.

8.2 The California Current System

Meridional current systems are observed along all of the west coast of all of the Americas. Within these circulation features are strong upwelling and downwelling centers that exchange ocean properties vertically. These currents and their poleward undercurrents are also the conduits of information and exchange of ocean properties over great distances between the tropics and the subpolar gyres. They are not steady circulation patterns. Beside ubiquitous mesoscale eddies, the California Current System (CCS) exhibits seasonal, interannual, and decadal timescale variations that today are not well modeled and not well observed beneath the ocean surface.

The coastal sea-level records in the CCS show rapid propagation of subinertial-frequency signals from the tropics northward that are consistent with the dynamics of coastal-trapped, forced topographic Rossby waves (Allen and Denbo, 1987; Davis and Bogden, 1989). The SST anomaly patterns that spread northward during ENSO, however, have offshore scales much larger than coastal-trapped waves. Air-sea interaction could also cause these ENSO timescale SST changes over the eastern Pacific (Liu *et al.*, 1998). The COADS air-sea interaction data indicate that the CCS, like the equatorial cold tongue, is a region of significant net heat flux into the ocean. Small fractional changes of this large net flux can produce significant upper ocean temperature changes. One such large-scale event was the Pacific Ocean "regime shift" of 1976 that warmed the eastern Pacific. Whether this was the result of air-sea fluxes or oceanic processes cannot be determined from the current COADS data because the net insolation under the low cloud deck which overlays much of the CCS is not well known. Instrumental observations of the net heat flux in this area have not been made on a systematic basis. Salinity patterns exhibit the longest timescales and the deepest vertical scales of evolution (Chelton *et al.*, 1982). Both salty and fresh ENSO events occur in the eastern Pacific. A complete picture of the changing climate of the Pacific requires understanding the eastern boundary processes about which we now have only a rudimentary view.

Observing the subsurface climate signal in the California Current system requires sampling that does not alias the large mesoscale eddies into the climate signal. The CalCOFI program carries out quarterly surveys of the southern California Bight with 100–300 km longshore resolution. Quarterly monitoring lines, with higher offshore resolution, have been established westward off the coasts of Monterey Bay, Oregon, Vancouver Island, and southward from Anchorage. High-density XBT lines cross segments of the eastern boundary current system on great circle routes between Hawaii, Los Angeles, San Francisco, Anchorage, and Peru.

PBECS could, in cooperation with other programs concerned with the California Current System, enhance observations in two specific ways to understand climate variability:

First, it is proposed that an array of moored, air-sea interaction buoys carrying a standard set of ocean and atmosphere observations be established between Vancouver Island and Baja California, Mexico. These would not only serve to determine the relationships between the local forcing and the local subsurface response, but also to ascertain whether, and how much of, the subsurface temperature and salinity signal propagates northward by ocean dynamics, and how much propagates northward due to direct air-sea interaction. These buoys would carry comprehensive surface meteorology and radiation instruments and closely spaced temperature, velocity, and salinity sensors down through the seasonal thermocline. Initially, five buoys are proposed to be moored, for example, at the latitudes of Punta Eugenia (Baja California), CalCOFI Line 90, Monterey Bay, southern Oregon, and Vancouver Island. They would be placed offshore at the middle of the historical axis of the baroclinic structure of the California Current System. The data would be available for operational marine forecasters in real time.

Secondly, autonomous gliders would occupy hydrographic sections at the same latitudes as the buoys and at additional lines placed between the buoys, for a total of up to sixteen sections. These sections would be sampled on spatial and temporal timescales appropriate to reducing the aliasing of the climate signals by mesoscale variability, and to provide a more complete view of the temporally and spatially evolving subsurface field than is now possible. Initially, temperature and salinity profiles would be gathered and transmitted via satellite. Eventually, bio-optical observations could be added to the instrument suite. The details of instrumentation, sampling, and the precise locations for moorings and autonomous hydrographic lines are not yet determined. A meeting has been called to organize a joint proposal in January 2000 for submission soon after that.

8.3 Southeast Pacific Ocean boundary currents

Over a large region off the west coast of South America, very low SSTs are associated with prominent decks of stratus and stratocumulus clouds. This association suggests a positive feedback mechanism for ocean-atmosphere coupling. Because cold SST stabilizes and thereby isolates the saturated atmospheric boundary layer from the much drier air above, the formation and persistence of stratus decks under the subtropical subsidence regime is enhanced. The clouds, in turn, promote cooling of the ocean surface by reducing shortwave radiation, and may play an indirect role in strengthening the trade winds and evaporative heat loss through the effects of enhanced radiative cooling on the large-scale atmospheric circulation. These feedbacks probably play a role in the evolution of the climatically important cold pool of surface waters in the southeast tropical Pacific. If the positive feedback is essential to the maintenance of the stratus decks, then the system may be sensitive to external perturbations. Understanding the factors that contribute to the cold SSTs (including surface fluxes and offshore advection of cold upwelled water), as well as the factors that allow the formation of stratus decks (the strong subsidence inversion that caps the layer, and the vertical distribution of water vapor) is important to investigating this coupling feedback and its role in climate. These phenomena, and studies of them, were discussed in Section 8.1.

Coastal upwelling is prominent off the coasts of Peru and Chile because the coast is aligned with the low-level Hadley circulation winds (the height of the Andes close to the coast probably contributes to this alignment). Some of this upwelling may contribute to evolution of the equatorial cold tongue through advection of cold upwelled water from the coast towards the equator in the South Equatorial Current (Reverdin *et al.*, 1994). A strong seasonal cycle has maximum equatorward wind and minimum sea surface temperature (SST) in Austral winter. The upwelling raises shallow thermocline waters at the coast, which are then advected offshore by the Ekman flow. Of course, there is also strong interannual variability in SST, with significant warming during El Niño, even while the equatorward winds that drive the coastal upwelling off South America persist. This SST modulation could be accomplished in the face of upwelling-favorable winds through a deepening of the thermocline during El Niño, so that warm water is upwelled instead of cold, and/or by a strengthening of the offshore meridional surface pressure gradient so that geostrophic flow toward the coast opposes Ekman flow away from the coast (Huyer *et al.*, 1987).

The current system offshore of the continental shelf probably provides cold water for the upwelling and may also provide a significant route for oceanic equatorial to subtropical teleconnections. The Peru Undercurrent (PUC) flows rapidly (∼18 cm/s) poleward off the shelf break below near-surface equatorward flow (Huyer *et al.*, 1991). Water mass analysis suggests that the PUC is fed by the equatorial undercurrent (Lukas, 1986). The PUC is 300 m thick, 50 km wide, and transports ∼1 Sv. This transport can supply coastal upwelling as far as ∼15°S (Strub *et al.*, 1998).

The Peru-Chile Countercurrent (PCCC) is a larger poleward surface flow, located between 100 and 300 km offshore through altimetry and water-mass analysis (Strub *et al.*, 1995, 1998). The PCCC appears to be supplied with cold water from the southern subsurface countercurrent, an eastward jet on the south flank of the deep equatorial thermostad (Tsuchiya, 1985; Johnson and McPhaden, 1999). Satellite altimetry suggests a strong seasonal cycle to the PCCC, with flow peaking in Austral spring, velocity variations of 50 cm/s, and transport variations of ∼8 Sv (Strub *et al.*, 1995). This transport magnitude is twice that of the southern equatorial subsurface countercurrent (Rowe *et al.*, 1999), and is sufficient to supply all coastal upwelling poleward of 15°S (Strub *et al.*, 1998). This magnitude is also significant from a general circulation standpoint, being half that of the interior ocean equatorward flow in the South Equatorial Current (Johnson and McPhaden, 1999). The PCCC could be a significant route for mass, heat, and freshwater exchange from the equatorial regime to the subtropics, the only other being poleward Ekman transport.

The upwelling, PUC, and PCCC vary with season and phase of ENSO. These features have sufficient scale and magnitude to play a significant role in the basin-wide climate system. Because of its origin near the coast and its narrow offshore scales, assessment of the upwelling, its source waters, and its role in air-sea coupling will require a targeted array of in situ measurements, as will any attempt to quantify the potential equatorial to subtropical ocean route for mass, heat, and freshwater fluxes in the PUC and PCCC. The broad interior-ocean coverage provided by satellites and Argo profiling floats will not resolve these features. A minimum observational array for PBECS should allow determination of the sources of the upwelling, and the poleward fluxes of mass, heat, and freshwater in the PUC and PCCC. It would resolve variability of the system on seasonal timescales.

An array to observe the ocean components of this climate system should augment the surface moorings and improved VOS for air-sea interaction studies and coastal mooring array planned by Peru and Chile that were mentioned in Section 8.1. Additional ocean observations would be an appropriate combination of high resolution XBT/XCTD sections, autonomous CTD gliders sections, and current and temperature measurements on available moorings. Up to three glider sections perpendicular to the coast between 5°S and 25°S would allow study of upwelling along the coast, and assessment of the current magnitudes and variability from the equator into the subtropics. Steep continental slopes here allow gliders to approach close to the shelf break while profiling to about 1000 m. A horizontal glider speed of 20 cm/s would allow a 300 km section with 5–10 km horizontal resolution to be made in 17 days. Sections would be subject to noise from internal tides and waves; this might be reduced by temporal filtering of records from the slowly moving vehicles or perhaps by using glider pairs to allow geostrophic shears to be calculated between simultaneous profiles. Sections might also be subject to aliasing by coastally trapped waves that are energetic in the 8–11 day band (Cornejo-Rodriguez and Enfield, 1987), but this could be reduced using filtering that exploits the nonuniform sampling intervals associated with alternating sampling directions, and coherent filtering using altimetric records and data from a few moored current meters to define space-time patterns of variability.

9. Process Experiments

A fundamental characteristic of the ocean-atmosphere climate system is its nonlinearity. So the interannual and decadal variability that is the focus of PBECS cannot be understood and predicted (Lorenz, 1969) without an understanding of higher frequency processes. It is impractical to monitor the ocean on a scale adequate to resolve all of the processes that ultimately affect climate. It is, therefore, necessary to perform focused experiments to examine these processes and build parameterizations for inclusion in models, including assimilating models. We propose a sequence of process experiments as an important adjunct to PBECS.

The focus of the process experiments is the shallow (less than 1000 m) meridional overturning circulation that links the subtropical and equatorial Pacific. The meridional circulation can be considered very broadly to have four components: (1) the movement of surface water downward out of the mixed layer in mid-latitudes, (2) equatorward geostrophic flow to the tropics, (3) upward transport to the surface near the equator, and (4) near-surface poleward flow in boundary and wind-driven currents. PBECS monitoring is designed mainly to quantify the equatorward (2) and poleward (4) branches of the meridional cell. Process experiments are suggested for all four of the branches to complement PBECS.

Consider the first branch of the meridional cell, the subduction of surface water in the subtropics. Atmospheric conditions are imprinted on the ocean as temperature-salinity anomalies at mid-latitudes. A key issue to be addressed by process studies is the generation of these anomalies. The T-S anomalies are then subducted into the ocean's interior, acting as a record of past atmospheric conditions. The process of subduction is itself dependent on air-sea fluxes. Annual and interannual variations of mixed-layer depth are crucial (Stommel, 1979), as are wind-driven flows. Process studies are needed to clarify the vertical structure of turbulent fluxes of heat, salt, and momentum in the upper ocean as anomalies are formed. After being subducted, anomalies of both PV and T-S properties are propagated by some combination of advection, wave dynamics, and eddy fluxes. Issues to be addressed in process studies are the rates of dispersion and dissipation along equatorward pathways. One critical element of this question is how anomalies are propagated and/or generated along western boundaries, particularly where low latitude interior currents bifurcate at the western boundary. A process study in the bifurcation of the North Equatorial Current would help to both understand the role this process plays in climate variability and to improve models of western boundary currents.

The third branch of the meridional cell involving upward fluxes near the equator has been poorly observed. Upon reaching the equator, the water that originated at the surface in mid-latitudes is finally upwelled. Quantification of this upwelling is an important goal of a process experiment. The upwelled anomalies affect sea-surface temperatures and thus are important in the equatorial ocean-atmosphere system. Such T-S anomalies have been observed to influence the atmosphere in coupled general circulation models, but the observational basis for understanding the process is weak and deserves attention. Particularly in equatorial latitudes where vertical gradients and shears are strong, diapycnal mixing clearly supports important vertical fluxes. Process studies with the goal of improving mixing parameterizations need to quantify the internal wave spectrum at low latitudes, evaluate the role of salt fingers, and resolve daily cycles.

The overturning cell's fourth, poleward, branch involves wind-driven currents in and just below the mixed layer. Better understanding these flows will help to improve the design of the monitoring observations for PBECS. For example, a better quantification of the vertical structure of wind-driven currents will aid in designing the program to measure near-surface velocity and the Ekman heat transport, which is greatly affected by the depth to which the currents penetrate.

Process experiments will benefit from the monitoring and assimilation activities in PBECS. The processes occurring along the meridional circulation have all, to some extent or another, been observed. In all of the cases summarized below, we have a clear idea of the problem to be solved. The technology to solve the problem is also in hand. What has often been lacking in past attacks on climatically relevant local processes is the simultaneous observation of the large-scale ocean. For example, the Subduction Experiment in the North Atlantic in 1991–1993 focused on the equatorward branch of the meridional cell. Detailed observations of the process of subduction were made, but it was difficult to relate them to basin-scale conditions. A great advantage of embedding process experiments in PBECS is the very existence of the broad-scale observing system, which will allow scientists to make the connection between local processes and climate.

The process experiments, discussed in more detail below, are strongly linked by their importance to the meridional cell and, thus, to climate. The overarching goals are to improve parameterizations for use in models and to aid the design of the observational system. We anticipate that small groups of scientists will form to develop and carry out these experiments during the 15-year PBECS time frame. In the following, we simply highlight scientific objectives of importance to climate. Details of implementation are left to the scientists who will one day address the objectives.

9.1 Diapycnal fluxes in the daily cycle below the equatorial mixed layer

Accurate estimation of sea-surface temperature (SST) is essential for realistic climate predictions, and nowhere is it needed more than on the equator. For example, Wang and McPhaden (1999) studied the seasonal heat balance in the equatorial western Pacific and found that the surface-layer balance results from a "complex interplay between surface fluxes, advection, and mixing." Their residual heat flux, representing the sum of turbulent entrainment at the layer base and downward heat diffusion, had the same magnitude as the net surface heat flux less the penetrative flux passing through the layer. When predicting climate, the mixing component cannot be calculated as a residual. Instead, it must be calculated from other variables in numerical models.

Owing to its recognized importance, near-surface mixing has been studied intensively on the equator during several process-oriented programs: TROPIC HEAT, TOGA/COARE, and TIWE. All of these have found surprisingly different average mixing levels, as illustrated by the mean profiles of diapycnal diffusivity, K_ρ, in Fig. 9.1. These differences result from changes in the intensity of a daily cycle of mixing in the stratified water below the surface layer (Gregg *et al.*, 1985; Moum and Caldwell, 1985). Figure 9.2 shows several cycles of the deep daily K_ρ cycle during TIWE, the Tropical Instability Waves Experiment. For tens of meters below the surface layer K_ρ varied by factors of 10–100, peaking during the night. By comparison, at most open-ocean sites, convective mixing extends only a few meters below the surface layer. The different behavior on the equator is a consequence of the high mean shear and low mean Richardson number below the surface layer on the equator in the central and western Pacific. Variations in the easterly winds and

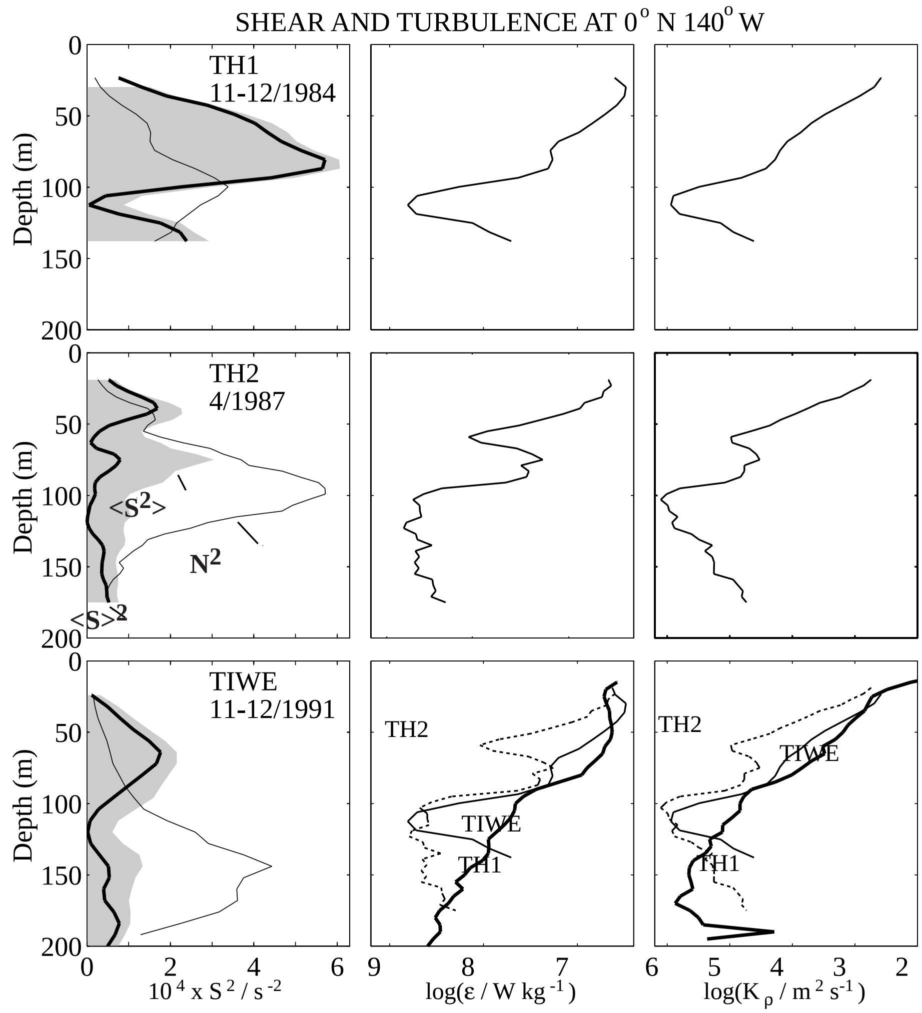

Figure 9.1: Average profiles of stratification, shear, turbulent dissipation, and K_ρ on the equator at 140°W during TROPIC HEAT I (Nov–Dec 1984), TROPIC HEAT II (Apr 1987), and TIWE (Nov–Dec 1991) from Gregg (1998). Panels on the left show stratification N^2, square of the mean shear $< S >^2$, and shear variance $< S^2 >$).

the intensity and depth of the undercurrent modulate the magnitude of the deep cycle and hence the mean diapycnal fluxes of momentum and heat into the surface layer (Lien *et al.*, 1995).

A daily cycle of finescale shear variance and small-scale internal variability accompanies the mixing cycle. One line of studies attributes the turbulence to breaking of the internal waves. Two mechanisms have been explored for the generation of internal waves beneath a convecting surface layer. One is the oscillation of the mixed layer base by convective plumes impinging on it. The other is the response of the stratified shear flow of the undercurrent sweeping across the base of the surface layer roughened by convective plumes. In both cases, the internal waves become dynamically unstable and break while propagating downward through a mean shear flow having a gradient Richardson number of about $\frac{1}{4}$. Another line of studies (e.g., Clayson and Kantha, 1999), concludes that the turbulence is generated directly at the base of the mixed layer and extends downward into a profile already close to instability owing to the high mean shear. According to

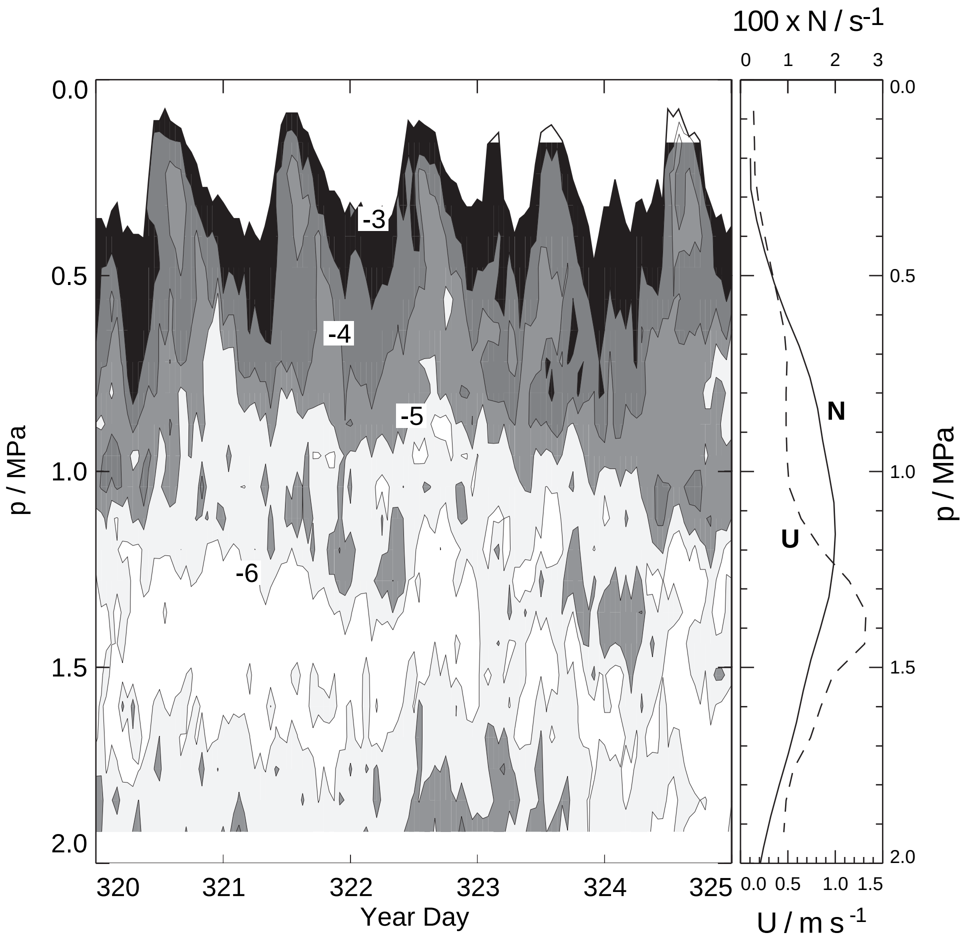

Figure 9.2: Shaded contours of $\log_{10} K_\rho$ begin at the base of the surface layer in the left panel, and the mean profiles of eastward current U, and stratification N are shown to the right for 5 days during TIWE (Gregg, 1998). The diapycnal diffusivity waxes and wanes with a daily cycle lagging the deepening and shoaling of the surface layer.

this scenario, the internal waves are produced by the stratified turbulence and are a result, rather than a cause, of the mixing.

Three conclusions are evident: (1) diapycnal fluxes produced by stratified turbulence below the equatorial surface layer must be parameterized for use in climate models; (2) no satisfactory parameterization exists and even the mechanisms responsible are not agreed upon. Moreover, it may not be possible to parameterize the mixing adequately with variables now in climate models. (3) Further piecemeal work by individual investigators, whether cruises or models, is unlikely to converge on accepted answers soon enough for climate modeling.

Developing satisfactory methods for including deep-cycle fluxes into climate models probably requires oversight from a CLIVAR working group on models and process experiments. Such a working group could be charged with developing coordinated numerical studies, process experiments, and monitoring upon which to base flux representations. This would put necessary focus into the exploration of ocean mixing studies.

9.2 Understanding and parameterizing diapycnal mixing at low latitude

Diapycnal mixing refers to transfers across mean density surfaces. Because the mean slopes of isopycnals are usually very small (10^{-5}–10^{-4}), diapycnal fluxes can be considered vertical but must be distinguished from the vertical components of lateral currents and mixing along isopycnals. These lateral motions are responsible for the large vertical eddy coefficients, $K_Z \sim 10^{-4}$ m^2/s, estimated from budgets (e.g., Munk, 1966), but their horizontal scales are much too large to alter water masses or dissipate kinetic energy. As representations of lateral currents and mixing become more realistic in numerical models, obtaining accurate parameterizations of true diapycnal fluxes is essential for predicting climate variability.

Three processes are known to produce diapycnal fluxes in the upper kilometer of the open ocean: breaking internal waves, salt fingers, and thermohaline intrusions. Of these processes, breaking internal waves are the most common and the best understood. They generate microstructure, allowing their diapycnal fluxes to be estimated from observed variances of temperature and velocity microscale gradients

$$K_T = k_T \frac{\langle |\nabla\theta'|^2 \rangle}{\langle \nabla\theta \rangle^2} \tag{3}$$

(Osborn and Cox, 1972) and $K_\rho \leq 0.2\ \varepsilon/\mathrm{N}^2$ (Osborn, 1974), whereas ε is the turbulent kinetic energy dissipation. Application to the large-scale, low-frequency general circulation is ad hoc (Davis, 1994a, 1994b), but its utility is demonstrated by the factor-of-two agreement with the rate of thickening of a tracer cloud during the North Atlantic Tracer Release Experiment (NATRE) (Ledwell *et al.*, 1993; Toole *et al.*, 1994; Sherman and Davis, 1995).

Microstructure cannot be measured over large areas, but the Garrett and Munk (1972) internal wave spectrum provides a reference linking diapycnal eddy coefficient to statistics of the internal wave field. Gregg and Sanford (1980) found that $K_\rho \sim 7 \times 10^{-6}$ m^2/s, with no N-dependence, when internal waves are at their background state. Henyey *et al.* (1986) constructed a simple analytical model for estimating K_ρ when internal waves differ from background. Gregg (1989) observed levels at mid-latitudes within a factor of two of these predictions, and Polzin *et al.* (1995) improved the agreement by using strain to estimate the average frequency of the waves. These agreements are good enough to make the problem of estimating K_ρ at mid-latitude one of determining the space and time structure of the internal wave field. Consequently, such an effort should be an important adjunct to PBECS.

How K_ρ from breaking internal waves behaves at low latitudes is not well understood. Henyey *et al.* (1986) give the dependence as $f \cosh^{-1}(N/f)$ with f the Coriolis frequency. As seen in Fig. 9.3, the sparse available data are consistent with this dependence, but there are large and unexplained differences between sites at different longitudes. Alford *et al.* (1999) obtained a time series showing varying K_ρ with the 4.5-day inertial period at 6.5°S in the Banda Sea. This variability is consistent with the observed near-inertial modulation of the Froude number, $(S/N)^2$. If this is a general condition at low latitude, then the inertial period establishes the length of sampling required to measure K_ρ, greatly complicating determination of the latitudinal dependence. Owing to the importance of low latitudes in modulating climate, process experiments to parameterize mixing rates there are given the highest priority. The first step is to determine if K_ρ is modulated at the inertial period in the open ocean, as it is in the enclosed Banda Sea. Then sampling strategies can be developed to adequately test Henyey *et al.*'s latitudinal dependence, and to develop alternative parameterizations if necessary.

Salt fingering is possible over wide areas, particularly equatorward of 25°, but there are no observations showing how often they occur or what diapycnal fluxes they produce. If K_ρ from breaking internal waves does vanish toward the equator, salt fingering may be the major agent of

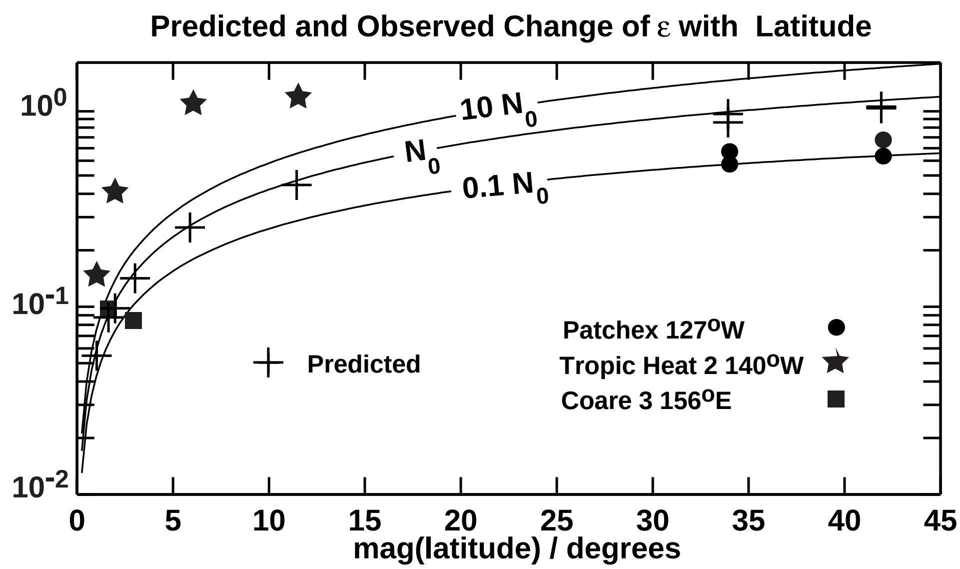

Figure 9.3: The predicted f-dependence of kinetic energy dissipation ε. Varying with stratification and latitude, the term is shown by three solid curves for a 100-fold range of N values. The crosses mark predictions for the particular N and f values where observations have been made, and the symbols show the corresponding observations after correction for the other terms in the Polzin *et al.* (1995) formula.

diapycnal transport at low latitudes. Comparison of microstructure levels with K_ρ from a tracer release should be included in at least one program measuring low-latitude microstructure and internal waves. Significantly faster thickening of the tracer would indicate that salt fingers are important and their fluxes should be included in models.

Thermohaline intrusions, particularly salt-stabilized temperature inversions, are common near fronts and are found to some degree nearly everywhere. They are often observed sloping across isopycnals and often contain signatures of double diffusion, but neither their generation nor their evolution is understood. Owing to the large magnitudes of thermohaline intrusions in equatorial fronts and apparently low levels of background mixing, intrusions may dominate diapycnal fluxes at low latitudes, even though they do not always contain strong microstructure. We cannot be confident of flux parameterizations until the contribution of intrusions are evaluated, and such an experiment should be one of the PBECS process studies.

9.3 Generation of subducted temperature-salinity anomalies

The downward branch of the meridional circulation involves the subduction of mixed-layer water in the subtropics. Subduction is an important element in global ocean-atmosphere interaction. The surface layer of the ocean exchanges heat, fresh water, and gases with the atmosphere. This exchange stops once the surface water is subducted into the interior of the ocean, and the properties set by the atmosphere are carried away with the water. In this way heat and greenhouse gases, for example, are sequestered in the ocean. The properties thus set in the mixed layer and subducted into the geostrophic interior act as an effective tracer of ocean circulation and a record of past atmospheric conditions. A clear understanding of this downward branch of the meridional cell is crucial to our ability to predict climate.

The main focus of a process experiment on the downward branch of the meridional cell is to determine how temperature-salinity anomalies are created and injected into the geostrophic circulation. While it is clear that these anomalies are due to atmospheric forcing, the mixed layer is an energetic region in which the anomalies are stirred and mixed before they pass into the deeper geostrophic ocean. For example, horizontal gradients of temperature and salinity in the mixed layer tend to compensate in their effect on density (Rudnick and Ferrari, 1999). That is, fronts in the mixed layer tend to be warm and salty on one side and cold and fresh on the other such that density fronts are minimized. These observational results suggest that horizontal diffusion in the mixed layer is an increasing function of density gradients, in agreement with the theoretical results of Young (1994). A process experiment should observe the establishment of T-S anomalies starting from the initial atmospheric forcing, as by rainstorm, to its eventually more compensated state. It has been known for some time that properties set in the winter mixed layer are transported downward along surfaces of constant density (Iselin, 1939), and that the annual cycle of mixed-layer depth is important (Stommel, 1979). Tracking specific anomalies from the mixed layer would be an essential objective of a process experiment. T-S anomalies on an isopycnal may be conveniently described using the variable called "spice" or "spiciness" (Munk, 1981), which quantifies the component of temperature and salinity that has no effect on density. Spice variability is observed to be largest in the mixed layer, and to decay with increasing depth (Ferrari and Rudnick, 1999). In view of the weakness of vertical mixing in the thermocline (Gregg, 1989), this decrease in spice has been attributed to salt fingers (Schmitt, 1999), although horizontal mixing may also be important (Ledwell *et al.*, 1998). An objective is to observe this decay in spice and determine the responsible mechanism. In summary, the objectives of a process experiment on the establishment of T-S anomalies are to observe and quantify: (a) initiation of anomalies due to atmospheric forcing; (b) evolution of anomalies through stirring and mixing in the mixed layer; (c) propagation of anomalies from the mixed layer to thermocline; and (d) decay of anomalies in the thermocline.

The observational capabilities already exist to address these objectives. It is possible to measure air-sea fluxes, horizontal and vertical structure of the mixed layer and thermocline, and turbulent mixing. The challenge of this experiment is that it is truly three-dimensional: both vertical and horizontal processes are crucial. The experiment is based on well-established theory, and theory is required in the context of this experiment to develop parameterizations for use in numerical models. Finally, the experiment will benefit greatly from the existence of PBECS monitoring. After anomalies have been tracked down into the geostrophic circulation, they can be handed off to Argo floats to follow their long-term propagation.

9.4 Processes of subtropical-tropical exchange

After they are subducted, it is unclear how anomalies are transported from the subtropics to the tropics. Variable subtropical-to-tropical exchange has been conjectured to induce decadal variation in low-latitude stratification and, in turn, the character of El Niño events, as discussed above. The modification of meridional exchange can be manifested as meridional mass transport anomalies, meridional heat transport anomalies, and/or meridional transport of potential vorticity (PV) anomalies. These are all related in one way or another to variations in air-sea heat, buoyancy, and momentum fluxes. A process experiment is envisioned within the large-scale PBECS observing system to explore the physics of the time-varying subtropical meridional overturning cell.

The experiment would observe the evolution of a PV anomaly as it moves south in the low latitude central North Pacific to, and through, the thermocline ridge marking the boundary between the NEC and NECC at about 10°N. The specific objective would be to tag a PV anomaly and measure the rates at which the anomaly is dispersed and dissipated as it translates equatorward. The long-range goal of the study is improved understanding of how off-equator PV anomalies influence low-latitude sea-surface temperature. Of particular interest is how mixing influences the water passing through the PV barrier driven by upwelling at the ITCZ. Quantification of mixing within the upper Pacific Ocean's subtropical-tropical meridional overturning cell is needed to link dynamically the large-scale circulation observations of PBECS to the dynamics of low-latitude stratification change, SST variations, and decadal ENSO variability.

9.5 North Equatorial Current bifurcation study

Important pathways from subtropics to tropics involve low-latitude western boundary currents. The physics of western boundary currents are extremely complex, and present climate model resolutions are too coarse, and parameterizations too crude, to give confidence in the results of numerical experiments involving advection and mixing in this regime. Sparse enhanced monitoring of this regime will not provide a sufficient basis to improve models, nor will they overcome deficient model physics during data assimilation analyses. For PBECS, it is crucial to obtain accurate analyses of the cross-gyre exchanges that occur primarily in the western tropical Pacific when the North Equatorial Current (NEC) encounters the Philippine coast and splits into the Kuroshio and Mindanao Currents. The bifurcation of the NEC is affected by remote forcing from the interior of the Pacific and from the north along the western boundary, by local wind and buoyancy forcing, and by mesoscale eddies. The correct modeling of the interaction of these processes is essential to the correct modeling of the western boundary currents of the North Pacific. An intensive process study is required to provide the observational basis for assessing existing models, for improving deficiencies, and for determining the minimum long-term observing elements needed to support accurate analyses.

We have little observational or numerical basis for designing the long-term observations of the Pacific low-latitude western boundary currents and their variability. Ocean circulation models do not agree on the mean and variable transport of the Mindanao Current (MC), which is an important factor in the heat and salt budgets of the western Pacific warm pool. Because adequate time-series observations are lacking in the Pacific low-latitude western boundary currents, we cannot tell if any of these models is correct. Not even a complete annual cycle of MC transport has been measured. Fortunately, Japanese and German current time series have been obtained for the New Guinea coastal currents, which give us some guidance for the southern branch of the LLWBC system. A NEC bifurcation study would provide the intensive observations that are required.

9.6 Equatorial upwelling and the emergence of subtropical anomalies

Equatorial upwelling is the choke point of the meridional overturning circulation, where much of the water subducted into the geostrophic flows of both subtropical gyres emerges to the surface. Once upwelled, this water interacts with the atmosphere, and coupled feedbacks can occur because of the sensitivity of equatorial winds to SST gradients. The poleward Ekman divergence of upwelled waters can also be a vehicle for interhemispheric exchange. Since it is not feasible to directly measure vertical velocities for sustained periods, monitoring this feature of the overturning circulation must ultimately be conducted through models constrained by assimilated data of diverse types, probably including horizontal velocities and water properties. A principal goal of a process study should be to learn what proxy observations will be most useful for inferring upwelling and its role in the overturning circulation within such data-assimilating models. The second purpose is to use an intensive observing period to investigate the processes by which upwelling influences equatorial SST, either through changes in vertical velocity itself, or of the thermocline stratification that affects the properties of the upwelled water. The third goal is to observe the sensitivity of the coupled feedbacks to the path and emergence region of advected anomalies, and the signal sizes of these emerging anomalies compared to internal tropical variability.

This process study would observe, for a limited period of time including at least one complete (non-warm-event) annual cycle, the full range of processes that bring thermocline-level water to the surface, including how water properties are modified in the strong mixing regime above the undercurrent core. One focus is how ocean-atmosphere feedback is triggered by emergence of subducted anomalies, which likely depends on the anomalies themselves. A potential vorticity anomaly can affect the upper heat budget when the associated anomalies of temperature upwell into the upper layer (i.e., mean advection of anomalous temperature), or if the associated horizontal and vertical velocities change the temperature field (i.e., anomalous advection of the mean field) or a combination of both. A spiciness anomaly, on the other hand, does not affect oceanic density and therefore only enters the surface heat budget by changing the temperatures of upwelled water. The process study will require simultaneous observation of winds and surface radiation fields along with vertical profiles of u, v, T, and S at a spatial resolution fine enough to distinguish the effects of horizontal versus vertical advection, mixed layer deepening and entrainment, and heating due to penetrative solar radiation. Ideally estimates of rainfall (probably from satellites referenced to moored rain gauges) would add the ability to confirm the processes deduced from the temperature balance with a salinity budget. Since we have only rather general ideas about the zonal or meridional scales on which upwelling takes place, an array of velocity moorings to measure divergence of the horizontal velocity components (which could also be sampling temperature, salinity, and winds) would likely require on the order of ten sites, in order to be able to take derivatives over a variety of separations and to span the region over which upwelling takes place ($\pm 3°$ latitude). Because much of the Ekman flow occurs in the upper 25 m that is poorly sampled by acoustic current profilers, deployment of point current meters at 10–15 m depth on several TAO picket lines would greatly contribute to our ability to understand the wind-driven divergence that drives upwelling.

If, in the course of the PBECS program, a particular low-frequency subducted thermal anomaly is identified and is predicted to emerge in the equatorial upwelling regions, a study that goes beyond the standard PBECS observations should be mounted to examine how it emerges. This would provide an observational basis to test the ideas and hypothesis about the coupled response to emergent anomalies.

9.7 Vertical structure of horizontal currents

Determination of ocean circulation within PBECS will depend largely on an essentially geostrophic methodology using temperature and salinity observations under the constraints of hydrodynamic models. The upper ocean, as demonstrated by shipborne ADCP observations (Chereskin and Roemmich, 1991; Wijffels *et al.*, 1994), however, is not in geostrophic balance. The ageostrophic circulation extends well below the temperature mixed layer. Models used to describe this upper-ocean circulation parameterize mixing processes that have been well observed in the planetary boundary layer, or mixed layer (Large *et al.*, 1994). Estimates of the vertical diffusivity in the mixed layer during wind mixing conditions are 0.01–0.1 m^2/s. The cross-isopycnal diffusivity 300 m below, in the center of the main thermocline, is 1–3 $\times 10^{-5}$ m^2/s, three orders of magnitude smaller. Existing models of upper-ocean mixing do not adequately parameterize the transition of vertical diffusivity from the geostrophic thermocline circulation to the ageostrophic surface layer.

This is because observations of the forcing of the three-dimensional structure of the upper ocean currents, together with the baroclinic density structure, have not been made in any mixed-layer experiment.

A complete horizontal-momentum, heat, and salt budget experiment of the upper ocean would provide the data set needed to unravel the mixing process below the mixed layer. In such an experiment the pressure gradients and Coriolis forces could be separated and the vertical convergence of turbulent momentum fluxes would be isolated. This experiment is now possible because of autonomous techniques for measuring temperature and salinity, and ways of profiling ocean currents with individual moored instruments or moored acoustic profilers. The uncertainties of in situ air-sea flux estimates have been reduced five fold in the past 10 years so that heat and salt budgets below the mixed layer are now feasible. No longer do we need to measure heat content change to check the surface flux in tropical conditions.

Accurate estimates of motion and fluxes in the upper-ocean are required to achieve the PBECS objectives and this will require improvements of models of upper ocean mixing between the mixed layer and the main thermocline. Connecting satellite altimeter data to Argo subsurface observations also requires understanding the transition of dynamics in this region. The tropical current system south and east of Hawaii has a significant difference between geostrophic currents, which flow to the southeast, and the observed mixed-layer circulation, which is to the north and west. This trade-wind driven system is a prime candidate for study and would be a test bed for the methodology of the PBECS data and models.

10. Integrated Atmospheric Observations

The success of coupled ocean-atmospheric models, as well as the utility of assimilated data sets for research purposes from PBECS, will depend crucially on detailed atmospheric observations with sufficient coverage in both time and space. The existing atmospheric observing system, especially over the extratropical Pacific, has large gaps in coverage. PBECS, in conjunction with other observational efforts, could substantially improve this situation through the optimization of emerging observing and data assimilation systems.

As discussed in Section 2, accurate estimates of the fluxes of momentum (surface wind stress) heat, and the input of fresh water are the cardinal requirements for driving ocean general circulation models (OGCMs). The boundary forcing of an atmospheric GCM (AGCM) is, at least in principle, somewhat simpler, and can be achieved by specifying the SST and the spectrum of incoming solar radiation at the top of the atmosphere. The latter field can be quite accurately calculated. SST is relatively well-observed, compared to the fluxes and precipitation (Section 4), although the sensitivity of the atmosphere to the diurnal cycle and other high frequency fluctuations in SST can not be discounted, particularly in the tropics (e.g. Weller and Anderson, 1996; Shinoda *et al.*, 1998; Woolnough *et al.*, 2000). Thus the boundary forcing of AGCMs might appear to be well-specified compared to the requirements of OGCMs. However, in practice, even the best AGCMs are still severely hampered, since even the best models are not capable of producing the correct ocean-to-atmosphere flux and precipitation fields, even given perfect boundary forcing. Improvements in AGCM climate simulations are therefore focused on improving model physics, such as parameterization of convection and radiation.

Historically, the development of OGCMs for climate studies has been hampered more severely than that of AGCMs by a lack of accurate boundary forcing data, particularly wind stress (Section 3.3). While substantial improvements in the surface wind stress fields have been realized using satellite scatterometers (Section 4.1), these observations are by no means continuous in time or space. Accurate surface fluxes of heat and moisture are even more difficult to obtain. Better surface flux measurements could also aid the diagnosis and improvement of AGCMs, since those models can be forced directly by such fluxes, without a need to rely so heavily on boundary layer parameterization schemes. It follows that in order to test and improve both AGCMs and OGCMs, and therefore coupled models, and to aid in process studies of ocean-atmosphere interaction, better estimates of surface fluxes and wind stress fields are critical.

There are relatively few direct measurements of surface fluxes of momentum, heat, or moisture at points across the Pacific. The PBECS strategy to use additional estimates of these quantities depends not only on increasing the coverage of direct surface observations, but is necessarily also concerned with improving observation of the entire atmosphere. One reason for this emphasis is that the present state-of-the-art data assimilation systems can, through the use of derived fields obtained from AGCM output, produce estimates of fluxes which can then be used as "best guess" boundary forcing for OGCMs (see Sections 3 and 5). In addition, in coupled models used for seasonal-to-interannual forecasting, well-specified initial conditions are mandatory. In all cases, upper level atmospheric as well as surface observations are crucial to obtaining the best possible derived fields for model initialization.

The spatial and temporal resolution of the atmospheric observations required to achieve adequate AGCM or coupled simulations for seasonal-to-interannual forecasts has not been determined. It is, however, becoming evident that intraseasonal variability at relatively high frequencies, such as the MJO at the 30-60 day time scales, is critical to such forecasting, especially where ENSO is concerned (Section 2.3). Even higher frequency fluctuations may be important to such forecasting. Takayabu *et al.* (1999) presented evidence that the sudden demise of the 1997-98 El Niño event was related to a trade wind surge over the eastern equatorial Pacific which upwelled a reservoir of cold subsurface water to surface over a period of a few days. This surge was attributed to the MJO, but in fact appears to be linked even more closely to the passage of a convectively coupled Kelvin wave, which moved rapidly eastward over the Pacific within a matter of a few days. Such waves move at a phase speed of around 17 m/s, and have easterly anomalies to the east and westerly anomalies to the west of their convective centers (Wheeler and Kiladis, 1999; Wheeler *et al.*, 2000).

Since one primary focus of PBECS is ENSO variability, atmospheric observations within the equatorial wave guide are of prime importance. These observations must, to the best degree possible, be able to resolve not only the several thousand kilometer scale of MJO variability, but also the somewhat smaller scale and higher frequency phenomena such as convectively-coupled Kelvin waves and cyclone pairs related to equatorial Rossby modes (Kiladis and Wheeler, 1995; Numaguti, 1995), which are linked to the occurrence of "westerly wind bursts" and subsequent oceanic Kelvin modes (Harrison and Larkin, 1998). Thus a strategy is needed that takes maximum advantage of Pacific geography, the TAO array, and volunteer observing ships (VOS), to improve coverage of both surface and upper-level observations along the equator.

In addition to equatorial Pacific observations, extratropical Pacific atmospheric observations should also be given high priority. In particular there are critical questions concerning the role of ocean-atmosphere feedbacks in decadal atmospheric variability over the North Pacific (see Section 2.2 and Section 7). The extent of the role of North Pacific SST in determining the subsequent evolution of the atmosphere on a variety of time scales from intraseasonal to decadal is still an open question (Robinson, 2000). Conversely, the combined effect of radiative fluxes and wind stress forcing of the North Pacific on ocean circulation is important for modeling interannual and decadal fluctuations (e.g. Deser *et al.*, 1999; see Section 2). Unfortunately the lack of land-based observational platforms over the extratropical North and South Pacific requires a heavy reliance on mobile platforms, such as VOS and aircraft, and on remote sensing from satellites. It is of concern that key island stations, such as Midway and Wake, do not appear to be currently making routine observations. Obtaining routine reports from these stations is considered to take highest priority, since they would provide direct observations of fluctuations in the Pacific jet and storm track systems deemed ultimately critical to modeling variability of the North Pacific oceanic circulation (Section 7).

In order to optimally initialize coupled ocean-atmosphere models, an efficient methodology of data collection and assimilation is necessary to assure the utilization of all available data with adequate quality control. This is a primary consideration of PBECS and is discussed in some detail in Section 3.

The expansion of Pacific atmospheric observations for CLIVAR is a substantial undertaking that will necessarily involve the cooperation of multiple international agencies. Fortunately, several initiatives and organizational efforts that have already been put forth can bear directly on the problems of data gathering and assimilation over the Pacific. Some of these efforts are being organized under the auspices of the World Meteorological Organization (WMO) World Weather Research Programme (WWRP), and the WMO Working Group on Numerical Experimentation (WGNE). By aiding implementation of these initiatives, PBECS could in turn benefit such programs by providing additional impetus for extended observations in the context of a CLIVAR ocean/atmosphere research effort.

10.1 The Status of the Existing Atmospheric Observing Network

At present, there are large gaps in both the surface and upper-atmospheric observing networks over the Pacific (as is also true for the rest of the earth's oceans and large portions of some land masses). The Global Climate Observing System (GCOS) has advocated a network of atmospheric observations designed to monitor the state of the global atmosphere and ocean primarily for purposes of climate change detection and attribution. Details of GCOS can be obtained at `http://www.wmo.ch/web/gcos/gcoshome.html`

The existing surface observing system consists of direct observations from island stations, VOS, telemetered observations from both drifting and moored buoys such as the TAO array, and satellite measurements of surface wind from the ERS-2 and QuickScat scatterometers. Existing upper-level observations are obtained from radiosonde profiles from island stations and occasional ship launches, pilot balloon launches from islands, data collected from wind profilers, and aircraft observations. These upper-air observations are supplemented by remote measurements from satellites, which provide cloud vector winds and temperature and humidity sounding data. For the purposes of PBECS, an extension of this network utilizing a suite of pre-existing and expanded platforms is necessary to maximize the temporal and spatial sampling of the atmospheric state for modeling and research purposes.

A portion of the atmospheric monitoring data currently collected is disseminated to operational forecast centers via the Global Telecommunications System (GTS). Examples of the amount and types of real time data getting into the NCEP data assimilation system can be seen at `http://lnx40.wwb.noaa.gov/` by clicking on "Data Distribution". A comparable view of ECMWF operational data is available at `http://www.ecmwf.int/services/dcover/` by clicking on "data coverage maps".

From these maps it is obvious that much of the upper-air data available over vast reaches of open ocean comes from satellite estimates. While extremely valuable, these data sources do not measure atmospheric state variables directly, nor do they provide high vertical resolution. Consequently, they suffer from inaccuracies in both the algorithms used to retrieve the desired parameters and uncertainties in the assignment of a given observation to an altitude above the earth's surface.

The GCOS Global Surface Network (GSN) is shown in Figure 10.1. These sites are primarily

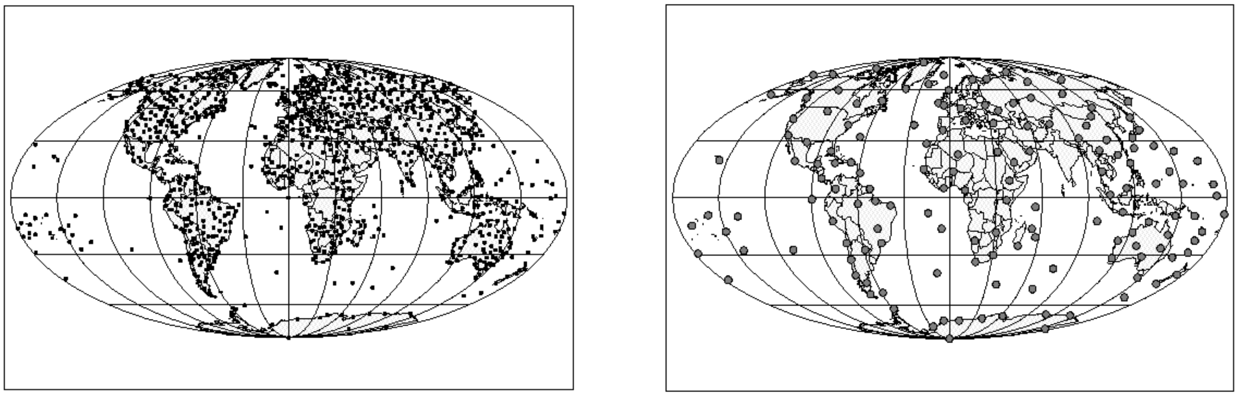

Figure 10.1: The Global Climate Observing (GCOS) Surface Network (GSN) (left) and the GCOS Upper-air network (GUAN). Source: www.wmo.ch/web/gcos/gcoshome.html – click on “Networks” then on GSN-map or GUAN-map.

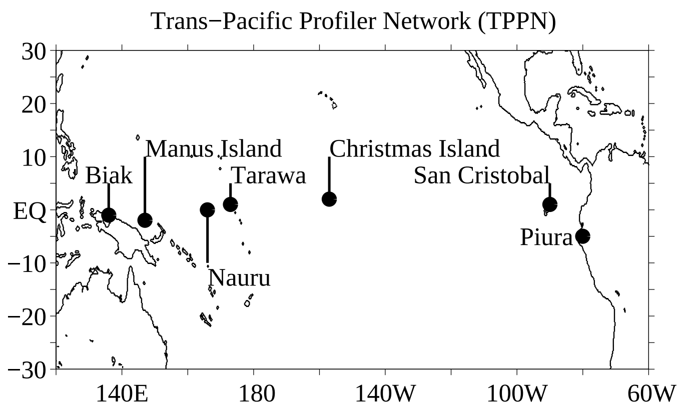

Figure 35: The Trans-Pacific Profiler Network (TPPN). Source: www.al.noaa.gov/WWWHD/pubdocs/ElNino.html

land-based surface observation stations reporting temperature and precipitation. Many also provide humidity and radiation measurements. In the Pacific, the region of best coverage lies over the warm pool and just south of the equator to about where island stations are plentiful. However, island surface stations suffer from local effects and may not best represent conditions over the open ocean. Thus for the purposes of initialization of forecast and coupled models, supplemental surface observations from buoys, ships of opportunity, and satellite are crucial. As can be seen at the NCEP and ECMWF sites, the real time surface data coverage provided by these supplemental observations is quite extensive, especially over the open ocean.

For data assimilation, short- and medium-range forecasting, and research, radiosonde and profiler data still provide the most reliable data sources for upper-level information on wind, temperature and humidity. The GCOS Global Upper-Air Network (GUAN) is shown in Figure 10.1. This network consists of a set of relatively reliable land-based stations launching radiosondes at least once a day. However, it is obvious from the NCEP and ECMWF real time observation reports cited above that many of these stations are not getting into the GTS data stream for one reason or another.

Figure 10.2 shows the locations of profilers within the Trans-Pacific Profiler Network (TPPN), maintained by the NOAA Aeronomy Laboratory. The locations at Biak, Christmas Island, and Piura are sites where 50 MHz profilers, which measure wind and vertical motion from roughly 2 to 15 km, are located. 915 MHz profilers are located at the other sites, including Christmas Island. These platforms measure wind and vertical motion from near the surface up to about 5 km. The array is meant to provide accurate (to within 1 m/s) measurements of wind at high vertical (<500 m) and temporal (30 minute or better) resolution along the equatorial wave guide. The 915 MHz profilers also provide information on droplet size distribution, cloud base elevation, and freezing level (Gage *et al.*, 1994) and are currently being used as ground-truth validation for the Tropical Rainfall Measuring Mission (TRMM) project. From a comparison of the locations of observations getting into the NCEP and ECMWF data assimilation systems with the locations of TPPN sites, it appears that the operational centers do not now utilize all of the profiler data on a regular basis. The Aeronomy Lab is working with the centers to correct this.

The ocean observing component of the TAO array is discussed in Section 4.4 (see Fig. 4.12). Atmospheric observations from the TAO buoys include surface wind, air temperature, and relative humidity at all of the sites, and rainfall and shortwave radiation are also measured at some sites. At present surface wind, air temperature, SST and humidity are telemetered to the GTS, and these data appear to be used regularly for operational assimilation at both NCEP and ECMWF.

In summary, as can be seen at the GCOS web site and maps of the GTS data stream at NCEP and ECMWF, at present surface atmospheric conditions in the Pacific are relatively well-monitored along the equatorial wave guide from the warm pool eastward to the Line Islands, thanks to a combination of land based stations and the TAO array. Surface observations are less dense in the subtropics, but provide reasonable coverage over the western Pacific and the central South Pacific. Land-based surface observations are severely lacking over the entire eastern Pacific and mid-latitudes of both hemispheres, although VOS and satellite-derived scatterometer winds fill in some gaps over the North Pacific.

Coverage of upper air observations by radiosonde is extremely poor in all but the western tropical and subtropical South Pacific. The bulk of the information on upper level winds over most of the Pacific comes from commercial trans-Pacific aircraft observations in the upper troposphere and cloud tracking satellite estimates of wind. Temperature and pressure are also measured by aircraft, although these data sources are nearly absent over most of the South Pacific except in the extreme west. Spatial coverage of satellite-derived wind and temperature is quite good, but, as mentioned above, at the current state of this technology these are difficult to assign to a specific atmospheric level.

10.2 General Recommendations

For the atmospheric observing component of PBECS, an optimization of the existing land-based observing platforms, as well as the addition of some radiosonde sites at key locations, would be most useful.

As discussed in Section 5., enhancement of surface meteorological observations should be considered by upgrading the instrumentation on selected VOS lines, by upgrading 3–5 TAO moorings to measure all parameters needed to infer fluxes, and deploying and maintaining 2–3 additional buoys outside the equatorial zone. If feasible, radiosonde ascents from VOS should also be considered.

The continued maintenance and possible expansion of the TPPN into the northern Line Islands is a cost effective way of increasing automated boundary layer observations of wind and vertical motion in a key region of tightly coupled air-sea interaction and strong interannual variability.

The possibility of aerosonde observations from land- or ship-based locations should be seriously considered. These light weight, robotic aircraft appear to have the potential for quality, high-resolution, targeted observations over remote regions of the Pacific at relatively little cost. Aerosondes have been used successfully in several recent field programs, and recently developed models have a nominal range of 5000 km and 50 hours endurance per flight. Models under development are expected to extend that range considerably. More detail on aerosondes can be obtained at `http://www.aerosonde.com/company_structure.htm`

A concerted effort to efficiently collect and disseminate data from all platforms in near real time, from one location, would provide a cost effective means of increasing the spatial and temporal coverage of atmospheric observations needed for operational data assimilation into models and for reanalysis products suitable for case study and statistical research applications.

10.2.1 Island-Based Observations

An effort to re-establish radiosonde sites at stations with pre-existing historical records is encouraged. For the statistical study of variability from the diurnal and synoptic time scales out to interannual and decadal time scales, it is imperative to have observations that are taken regularly and in a consistent manner at the same site for extended periods. This is also desirable for model data assimilation purposes, where coding and quality control of the data can be optimized for maximum impact on the analysis.

While the distribution of islands in the Pacific is unfavorable for land-based observation over vast reaches of the mid-latitude oceans, certain geographical accidents actually are quite favorable for monitoring the equatorial wave guide and the transition from the equatorial dry zone into the core of the ITCZ.

Specifically, the fortuitous alignment of islands along the equator provides for reasonably spaced coverage for surface and radiosonde observations except for the longitudinal range from 157°W to 90°W. In addition, the alignment of the Line Islands from Christmas Island (2°N, 157°W), north to Fanning Island (4°N, 159°W), Washington Island (5°N, 160°W) and Palmyra (6°N, 162°W) offers a natural and convenient transect for observations from the dry zone at Christmas, which receives less than 500 mm of precipitation per year, to Palmyra, which averages more than 5000 mm of precipitation per year, over a distance of only 650 km (Morrissey *et al.*, 1993). This configuration of islands was effectively exploited to study the circulation of the ITCZ-to-equator transition in 1968 during the Line Islands Experiment (e.g. Madden and Zipser, 1970). This region is strongly influenced by ENSO, and would provide a cost effective means for observing changes in the structure of the ITCZ associated with SST fluctuations over the tropical Pacific. In addition, the Line Islands lie close to the location of greatest discrepancy between different methods of satellite based precipitation estimates within the ITCZ (Janowiak, 1995).

While the ITCZ is relatively well sampled over the western warm pool at very reliable reporting stations such as Truk (7°N, 152°E) and Yap (9°N, 138°E), the character of the eastern Pacific ITCZ is different and much more influenced by ENSO (Kiladis and van Loon, 1988) and by Rossby wave activity originating in the extratropics (Kiladis, 1998).

At least one instrumented site should be considered at either Palmyra or Washington Island to sample the active convective region of the ITCZ. It is not clear what the logistics of a radiosonde operation at one of these islands would be. Barring a radiosonde, an automated 915 MHz profiler should at least be considered to sample boundary layer winds and the structure of precipitating convective and stratus clouds at this location. Surface observations would also be extremely valuable.

In order to take maximum advantage of Pacific geography, efforts should be taken to assure that radiosonde and/or profiler observations are also obtained at the following locations:

San Cristobal, Galapagos (1°S, 90°W): This location is listed as a GCOS, GSN and GUAN site, although the observations appear to be quite intermittent and do not regularly appear on the GTS data stream at NCEP or ECMWF. The Aeronomy Laboratory's 915 MHz profiler has been in operation reliably since October, 1994, and is being utilized for boundary layer research (Hartten and Gage, 2000). Given the location of the Galapagos within the equatorial cold tongue, and the strategic importance that regular profiles of wind, temperature, and humidity observations would be to programs such as PBECS, EPIC, and PACS, every effort should be made to return the Galapagos radiosonde and surface station to regular operation. The supporting infrastructure is already in place at this site, which has an airport with the necessary trained personnel on hand for observations. The main problem at present appears to be the communications equipment used to send data to the GTS, and a lack of funding to provide for the necessary expendables for radiosonde launches.

Midway Island (28°N, 177°W): Given the paucity of atmospheric data over the North Pacific, this location as the northernmost island south of the Aleutians should be a high priority for instrumentation. Although Midway has a lengthy historical record of surface and radiosonde observations, and is still a U.S. protectorate, upper level observations ceased in 1997 when the U.S. Navy vacated the island, and it was taken over by the U.S. Fish and Wildlife Service. Now a National Wildlife Refuge, Midway has an airport that sees frequent eco-tourism flights from Hawaii, with several personnel certified as NOAA weather observers. Regular surface and tide observations are still being taken and are reported to the National Weather Service and the FAA, although the surface observations do not appear regularly as GTS observations at the operational centers. The management of the airport is very willing to cooperate to reestablish upper air observations at this site. Funding appears to be the main issue. The possibility of Midway acting as a monitoring site for other PBECS activities, such as a base for aerosonde or ship operations, should be explored.

Wake Island (19°N, 168°E): Observations from this site would fill an important gap in subtropical observations between Guam (13°N, 145°E) and Hawaii. As with Midway, Wake has a long historical radiosonde record, although the more recent record in the NCAR sounding archives is quite discontinuous, but still has some data from as recently as 1998. If the site is still operating the soundings do not appear to be getting onto the GTS data stream at present. A relatively modest effort could be made to assure that such observations are transmitted correctly to be used for data assimilation and archived for research purposes.

There appear to be many island stations taking observations over the Pacific in conjunction with the GCOS, GSN and GUAN for climate monitoring purposes, but currently not reporting in real time on the GTS (this can be seen through a comparison of Figure 10.1 with surface and upper level observations shown being assimilated at the ECMWF and NCEP sites). The possibility that these stations could report those observations in real time should be investigated. If cost effective, these data could fill in important gaps in the network, particularly over the subtropics of both hemispheres but especially over the subtropical South Pacific, where islands abound.

10.2.2 Ship-Based Observations

Ship-based ocean state observations are described in detail in Section 4. Enhancements of atmospheric observations from the existing VOS lines should be given high priority. An effort to upgrade shipboard surface observations from VOS could, at relatively little cost, increase the coverage and accuracy of surface temperature, wind, pressure, humidity, flux, and radiation measurements. These parameters are crucially important to driving ocean models, and to the initialization of atmospheric and coupled models, yet are at presently poorly represented, at best, in reanalyses and operational analyses over data sparse regions.

The possibility and feasibility of radiosonde ascents from high-density XBT lines should be considered. One potential drawback of this approach would be the necessity of training shipboard personnel in the launching of these devices. As an alternative, automated shipboard balloon launching is also a possibility. However, given the potential expense of these systems, VOS lines which would utilize them should be carefully assessed to avoid redundancy with other enhancements by the proposed atmospheric observing programs to be discussed below. Nevertheless, the implementation of regular GPS sonde ascents from the NOAA R/V Ron Brown and Ka'imimoana has the highest priority. Efforts to continuously maintain C-band radar, S-band wind profiling, cloud radar and air-sea flux instrumentation aboard these two vessels are also critical.

10.2.3 Satellite Observations

Existing and proposed satellites designed to improve monitoring of the ocean and atmosphere are described in Section 4.1 of the PBECS implementation plan. As seen at the NCEP and ECMWF data assimilation sites, large data sets of remotely sensed wind, humidity, and temperature, as well as radiation parameters, are continuously being updated for use in operational forecasting. It is important to continue research by groups at NASA and universities developing algorithms to improve the quality of these data. These observations are readily available from the GTS in real time and after the fact at several data centers such as NESDIS, providing a valuable resource for data assimilation and research for PBECS. In addition, the TRMM project, administered jointly by NASA and NASDA (Japan), promises to greatly improve the quality of remotely sensed rainfall, cloud height and droplet size distribution data that will become available operationally within the time frame of PBECS. More information on TRMM can be found at `trmm.gsfc.nasa.gov/`

10.2.4 Other Observational Efforts

In addition to the above platforms, several field programs involving significantly enhanced upper atmospheric observations over the Pacific and the rest of the globe are being considered for the next few years. These programs would greatly improve the amount of data on the state of the stratosphere and troposphere particularly over the currently data sparse regions of the extratropics outlined above. Entrainment of the products produced by these programs into any PBECS data assimilation system should be encouraged. A formal collaboration between these observing programs and CLIVAR would almost certainly be beneficial for a broad variety of applications, among them data assimilation for synoptic-to-medium range forecasting, initialization of global coupled models, and ultimately for research on weather and climate variability on all time scales. Among those programs, the following appear to be particularly relevant:

The Hemispheric Observing System Research and Predictability Experiment (THORPEX). THORPEX is a program of observing system research and development aimed at improving skill of weather forecasts for Northern Hemisphere nations on time scales ranging from nowcasts to 10 days. The program will draw upon current and future data impact studies, observing system simulation experiments, and data assimilation development. The program would deploy in situ observing systems such as aerosondes, driftsondes, and buoys under operational conditions, in an effort to determine the best feasible mix of platforms to improve the initialization of numerical weather prediction models for forecasts over the Northern Hemisphere. THORPEX would also evaluate the impact of satellite observing systems on model initialization, as well as the use of "targeted" observations.

The initial experimental design involves oversampling the North Pacific atmosphere with as many observational resources as possible for several months. The effect of such observations will

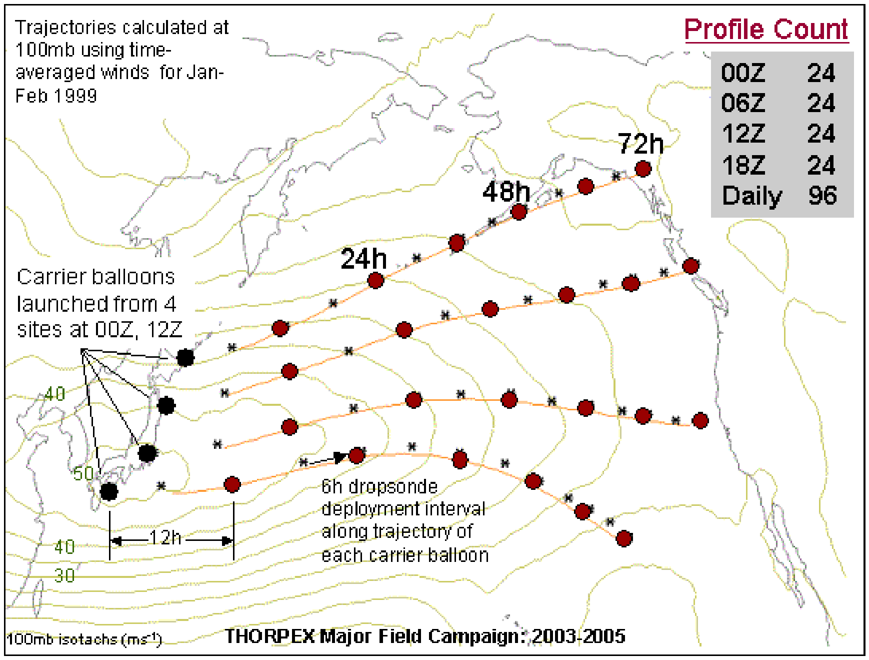

Figure 36: Driftsonde profile coverage at one assimilation time, 3 days after deployment from 4 launch sites in Japan. Each do represents a GPS dropsonde profile location at 00Z or 12Z from one of four carrier balloons/gondolas. Stars are profile locations at 06Z or 18Z. Source: Figure 5 of the THORPEX plan at: www.nrlmry.navy.mil/~langland/THORPEX_document/Thorpex_plan.htm

then be objectively determined by numerical model hindcasting where observations are withheld to determine their net impact on the forecasts.

A key aspect of THORPEX of particular relevance to PBECS is the use of driftsonde balloon systems currently being developed by the Atmospheric Technology Division at NCAR. These would provide atmospheric sounding data over the North Pacific (and the Atlantic) by launching large but expendable balloons into the stratosphere from eastern Asia and North America which carry payloads of several dozen dropsondes. The dropsondes would be deployed at pre-determined intervals to gather detailed data on temperature, wind, humidity, and perhaps ocean state data as the balloons are advected across data-poor regions. An example of the spatial coverage that could be obtained by this system is shown in Figure 10.3. These data would be transmitted to operational centers via the GTS and would be available for data assimilation to operational centers, PBECS model initialization, and researchers in real time.

Other platforms being proposed for THORPEX include:

- Aerosondes to be launched from Hawaii and western North America.
- Upper-air observations delivered by a weather-rocket buoy system, which will hold up to a year's supply of 400 rocketsondes. A network of up to 50 buoys designed for heavy seas is envisioned.
- Targeted observations by aircraft provided by the U.S. Air Force to be deployed from Hawaii to Japan to measure the atmospheric state, especially at jet stream level.
- Moored surface buoys provided by PMEL to measure surface pressure and sea surface temperature at hourly intervals, transmitted to the GTS.

- Radiosonde ascents from ships in conjunction with the ACE-ASIA atmospheric chemistry program.
- Atmospheric soundings over remote oceanic locations delivered by shear-directed super pressure balloons (see GAINS below).

An observing system test and evaluation phase of THORPEX is being planned for early 2001. The THORPEX major field campaign is currently slated for several seasons within the 2003-2005 time frame. The WWRP will facilitate the field and research activities, and the USWRP will act as a coordinating agency for U.S. participants in THORPEX. More detailed information on THORPEX can be found at `www.nrlmry.navy.mil/~langland/THORPEX_document/Thorpex_plan.htm`

Global Air-Ocean in-situ System (GAINS). GAINS would provide several hundred long-lived balloons cruising along constant latitude circles in the stratosphere near 60 hPa. These would deploy dropsondes for meteorological, atmospheric chemistry, and ocean state measurements at either predetermined or operationally determined intervals. The balloons would be recoverable after several months of service, and are intended to provide data from observation-sparse regions of the ocean for synoptic to climate time scales.

As with THORPEX, GAINS data would be relayed to the GTS in real time for operational and other data assimilation purposes. Since GAINS data will be global and available to worldwide meteorological services, it is anticipated that the costs will be shared multinationally, and the data will be distributed under WMO auspices.

GAINS is being proposed as a 50-year operational program which would eventually utilize up to 400 balloons. A prototype proof-of-concept balloon is being planned for 2001, and a pilot study using up to 3 balloons is slated for 2003.

Focused Research on Intraseasonal Tropical-Midlatitude Interactions to Improve NOAA's Weekly to Seasonal Forecast Capabilities. This initiative by NOAA's CDC and ETL would focus on improving weather and climate forecasts on weekly to seasonal time scales through improvements in observations, analyses, understanding and modeling of intraseasonal tropical-midlatitude interactions and their regional impacts on the U.S.

The efforts would begin in 2002 with focused studies over the Pacific, with an initial emphasis on identifying key mechanisms of tropical-extratropical interactions that can lead to improvements in forecasts in the 8-14 day range.

The program would result in dense targeted observations over regions of strong tropical convection such as the Madden-Julian Oscillation and adjacent subtropical jet regions known to be affected by such forcing. The project would also likely involve efforts to improve the assimilation of data into the operational forecast system, and produce an archive of detailed atmospheric observations over the tropics and subtropics of the Pacific for research purposes.

10.2.5 Data Assimilation and Archiving

CLIVAR and the other programs mentioned above can best benefit from an expanded data base only if such observations are made available in a timely fashion for data assimilation, and properly archived in consistent formats for future use in research applications (Section 3).

While the expansion of the various types of observations is a priority, close attention must be paid to the ways the data are collected and disseminated. A central location for the dissemination of all oceanic and atmospheric data related to CLIVAR and related programs would be ideal for this purpose. As an absolute minimum, there should be a "virtual data center" with pointers to all available data resources resulting from CLIVAR programs. For real-time access of both point measurements and gridded output, data distribution by user-initiated "pulls" from a web or ftp site has proven to be very practical. Consequently, data sets and software can be effectively distributed worldwide with standard formats. If a large volume of accesses is anticipated, the Unidata concept of scheduled "pushes" through a local data manager may be more appropriate. The importance of these relatively inexpensive measures has been demonstrated in many past field programs such as GARP, TOGA COARE, and TEPPS, among others. Management of the potentially enormous volume of oceanic and atmospheric data generated by future programs will necessarily require implementation of emerging technologies for data archival and processing. Serious consideration should be given to a centrally located effort of data management to avoid the loss or unavailability of such data to the community.

11. Summary and Recommendations

The Pacific Basin Extended Climate Study (PBECS) is a basin-wide study of interannual-to-decadal climate variability in and over the Pacific Ocean. This implementation plan lays out the main elements of PBECS as defined in a series of community planning meetings. It also points to areas where additional planning is needed to complete PBECS. Here we summarize the plan and enumerate recommendations to begin implementing PBECS.

11.1 Scientific objectives

The PBECS scientific objective is to increase understanding and predictability of ENSO, decadal modulation of ENSO, and decadal variability such as that embodied in the Pacific Decadal Oscillation. While these phenomena involve potentially distinct climate-scale processes, from a logistical point of view it would be inefficient to design separate attacks on each. Each study would depend on much of the same observational and modeling infrastructure. Rather, PBECS has combined the elements needed to study all of the climate phenomena into a single effort.

The scientific basis for PBECS was described by Lukas *et al.* (1998) and is summarized in Section 2.

The observational record shows that individual El Niños have developed differently and that the amplitude and frequency of the ENSO cycle have varied on decadal timescales. PBECS will explore the thesis that this is the result of changes in the oceanic background, particularly the depth, east-west slope, and structure of the thermocline. Several hypotheses have been advanced for the causes of decadal variability of the near-equatorial thermocline, involving a variety of interactions extending outside the equatorial zone. These include slower ENSO-like oscillations that span a wider latitude band, effects of North Pacific decadal variability on tropical winds that affect the thermocline slope, and oceanic anomalies generated and subducted in the subtropics and propagated to the equatorial zone through the interior ocean and low-latitude western boundary currents.

The North Pacific exhibits basin-scale ocean-atmosphere variability with decadal timescales that has a significant impact on seasonal climate forecasts over North America. The leading hypotheses for the mechanism of these fluctuations are that the ocean is passively filtering random weather fluctuations, that the subtropical signature is a response to an atmospheric teleconnection from the tropics, and that there is an ocean-atmosphere feedback within the subtropics. The subtropical feedback is hypothesized, mainly from model experiments, to involve a local reinforcing air-sea interaction near the Kuroshio-Oyashio Extension and delayed negative feedback, in which SST-caused wind-stress anomalies spin up a subtropical ocean circulation that later destroys the initiating SST anomaly.

Despite the significant predictive skill that has been achieved, there is much that needs to be better understood and predicted about the ENSO cycle. The main issues concern the processes governing initiation and termination of the El Niño warm phase, and how "background" oceanic conditions affect evolution of the ocean-atmosphere instability that sustains El Niño. Conjectures abound about nonlinear interactions with higher-frequency forcing, like the Madden-Julian Oscillation, and with lower-frequency trends, including decadal variations involving the mid-latitude Pacific and global warming. The oceanic role in ENSO involves the equatorial thermocline response to wind stress and the connection of thermocline depth to the sea-surface temperature that affects the atmosphere. The thermocline-SST relation, which involves diapycnal mixing as well as upwelling, is not well understood and is treated quite differently within various models. Understanding the dynamics connecting thermocline structure and SST evolution is a PBECS goal.

11.2 Assimilation and modeling

The PBECS strategy to understanding these phenomena will be comprehensive, integrating a diverse suite of in situ and remote-sensing observations inside data-assimilating models to describe the evolution of much of the Pacific Basin. The observational objective is to describe the Pacific Basin well enough that the evolution of climate-scale patterns can be diagnosed dynamically so that the various hypotheses for climate variability can be tested. This will allow improvement of models to simulate ENSO, its decadal modulation, and decadal variability itself. The rapidly developing capabilities of data-assimilating ocean models are described in Section 3. Several U.S. groups are exploring assimilation methods to produce ocean state estimates. The PBECS strategy is to work with these groups, providing data in near real time, and supporting improvements of assimilation methodologies and dynamical models upon which assimilation is based. Discussion of the interaction of data, assimilation, and model development and experimentation in Section 3 leads to the following recommendations:

1. Additional computer resources to support the extensive calculations of assimilation procedures should be found or developed.

2. Ways should be found to encourage observers to work with assimilators to prepare in situ and remote ocean and surface forcing data for inclusion in analyses, to test the assimilation products, and to use empirical knowledge to improve the statistical information upon which the assimilation procedure is based.

3. Under PBECS, the U.S. CLIVAR Pacific Sector Implementation Panel or the Science Steering Committee, a working group of modelers and experimenters, should be established to develop process experiments, numerical experiments, and model development activities to improve dynamical models.

11.3 Basin-wide observations

The backbone of the observing array for PBECS will be the suite of broad-scale in-situ and satellite observations described in Sections 4 and 10. These observations will provide basin-wide coverage of the interior ocean and broad-scale atmosphere in support of studies of the variability phenomena of concern to PBECS. Three of these elements are essential and must be continued for 15 years for PBECS to succeed:

4. Satellite altimetry with coverage and accuracy comparable to TOPEX/POSEIDON will be a primary tool for assimilation analyses and linking together in situ observations like Argo.

5. An array of 1200 Argo profiling floats should be maintained across the basin north of 40°S to profile temperature and salinity to 2000 m on 10-day intervals (see Section 4.3). These will

map the 3-D structure of heat and freshwater storage and lateral transport by broad currents, processes that figure centrally in all aspects of ENSO, its decadal modulation, and decadal variability.

6. The ENSO Observing System, including the temperature and wind measurements of the TAO array, is central to observing equatorial processes. TAO velocity sampling of the shallow currents most directly affecting SST should be augmented (see Sections 4.4 and 6.4). Additional salinity sampling should be considered.

The status of remote-sensing measurements in general, and altimeters in specific, are discussed in Section 4.1 where we are reminded that altimeters are scheduled to be transitioned to NPOESS by 2010 but the present NPOESS system is incompatible with high-accuracy altimetry. As Argo builds, the XBT sampling of the ENSO Observing System should migrate onto high-resolution lines. Augmentation of TAO salinity depends on clear demonstration of the importance of high-frequency salinity sampling that Argo cannot supply.

Recommendations for other elements of the basin-scale observing systems that are needed to achieve the goals of PBECS are:

7. The existing high-resolution XBT/XCTD sections (see Fig. 4.10) from VOS should be continued on a roughly quarterly basis to observe lateral oceanic mass, heat, and freshwater transports, which are inadequately sampled by broad-scale observations. These should be augmented by meridional lines in the far eastern and western Pacific (see Section 4.3).

8. Satellite measurements of surface vector winds with coverage and accuracy comparable to QuickSCAT are needed to improve fluxes, to force ocean models, and for diagnostic studies (see Section 4.1).

9. High-quality infrared and/or microwave sensing of sea-surface temperature (Section 4.1), supplemented by in situ reference time series, are needed to support diagnostic studies and estimation of basin-wide air-sea surface fluxes (see Section 5).

10. Approximately 200 surface drifters are needed to map surface velocity and provide accurate SST measurements in the subpolar North Pacific north of the ENSO Observing System (see Section 4.6).

Air-sea fluxes of momentum, heat, and water play a central role in the coupled phenomena of PBECS concern. A complete strategy for improving the availability of high-quality surface flux fields is outlined in Section 5. The following steps are recommended to implement this strategy:

11. Add high-quality time series of surface observations from moored buoys and specially instrumented VOS. Discussions at the St. Raphael OceanObs99 conference suggested U.S. support for five high-quality VOS lines, upgrading three to five TAO moorings, and maintaining two additional buoys.

12. Compare these data with NWP fluxes and remote-sensing fields to define biases and error characteristics. Develop basin-scale flux fields and understand how fluxes are involved in climate fluctuations.

13. Help develop AGCMs that can successfully assimilate all kinds of in situ and remotely sensed data to produce better routine surface meteorology and surface flux fields.

Improved atmospheric obervations are important to PBECS for four reasons: (1) to produce atmospheric analyses accurate enough for initializing coupled models and supporting diagnostic studies of climatic variability; (2) to provide long records of climate variability from intraseasonal to decadal time scales for statistical and empirical studies; (3) to improve knowledge of the atmospheric boundary layer and the estimation of air-sea fluxes of heat and momentum, from which air-sea interaction can be diagnosed; and (4) to provide verification data for inversion schemes used to deduce atmospheric structure from satellite measurements. The following recommendations are discussed in detail in Section 10.

14. Re-establish regular radiosonde sampling at stations where historical records have been interrupted. San Cristobal (Galapagos), Midway Island and Wake Island are high priorities.

15. Maintain upper air soundings from at least one site in the Line Islands for the study of the climatology and variability of the ITCZ. If radiosondes are impractical, a 915-MHz profiler could at least track fluctuations in the boundary-layer.

16. Implement regular GPS-sonde sampling from the TAO deployment cruises and investigate the feasibility of deploying radiosindes from high-resolution VOS lines (see item 7 above).

The basin-wide oceanographic array discussed above will resolve climate processes in the interior ocean (see 6.1). These broad-scale observations, however, are not well suited for measuring the transport and processes occurring in boundary currents or the highly zonated currents in the eastern tropical Pacific. Some of these regions are centrally involved in the climate processes of concern to PBECS and, therefore, special observations are recommended to complete the PBECS observing array.

11.4 Western boundary currents

Oceanic connections between the subtropics and equator play important roles in theories of both decadal modulation of ENSO and decadal variability. Interior pathways will be well observed by the broad-scale arrays, but in both hemispheres much of the communication occurs through low-latitude western boundary currents.

17. The Mindanao Current is the main conduit between the North Equatorial Current (NEC) and the North Equatorial Counter Current (NECC) and hence the equator. Long-term measurements of mass, and if possible heat and fresh water, transport east of Mindanao should be maintained using an appropriate combination of moorings, autonomous gliders, altimetry, and pressure gauges (see Section 6.2). These might be combined with measurements off Luzon to more fully monitor the NEC bifurcation.

18. The New Guinea Coastal Current and New Guinea Coastal Undercurrent are the main sources of southern hemisphere water to the western Equatorial Undercurrent and NECC. Suitable long-term transport measurements should be maintained in Vitiaz Strait or north of New Guinea's north coast (see 6.2).

19. The Indonesian Throughflow is another conduit of tropical waters that affects local climate and, potentially, the evolution of ENSO (see Section 6.3). The U.S. should work with Australia to upgrade the repeated VOS section I1.

The Kuroshio and East Australia Current are central elements in theories of subtropical ocean-atmosphere feedbacks that may be responsible for decadal variability (see Section 7). The Kuroshio is crossed by high-resolution VOS lines off Taiwan and Japan. It is hoped that scientists from Japan, Taiwan, and China will continue and expand other measurements of the Kuroshio in low latitudes and along Japan. The East Australia Current is presently crossed by high resolution VOS lines off Brisbane and Sydney.

20. It is recommended that the existing VOS lines crossing western boundary currents be maintained and upgraded with increased use of 2000 m XBTs, XCTDs, and underway CTDs. These lines should be augmented with autonomous gliders to increase sampling frequency. Two or three U.S. glider sections are suggested.

11.5 Special regions

The maximum variability of the Pacific Decadal Oscillation (PDO) is in the subtropics. The observing elements described above will well describe gyre circulation, thermocline variability, and large-scale subducted anomalies of concern for ENSO modulation and decadal variability. The KESS experiment (Section 7.2) will serve as a pilot study to prove methods for measuring important aspects of oceanic heat transport near the Kuroshio-Oyashio Extension (KOE). The key remaining issue is the cause of SST variability, specifically whether it is a response to anomalous air-sea fluxes or generates those fluxes.

21. One or two moorings with high-quality air-sea flux sensors and upper-ocean instrumentation should be maintained eastward from the KOE (see Section 7.1). The present NOPP mooring at 35°N, 165°W is a candidate for the eastern end of this array. One of these could serve as a flux reference site (see item 11 above).

The eastern tropical Pacific is the site of strong air-sea coupling associated with the equatorial cold tongue, the intertropical convergence zone, coastal upwelling, and persistent stratus cloud decks, all of which affect the local climate of the Americas and are the largest terms in the heat budget of the eastern Pacific (see Section 2.3 and 8.1). At the same time this is the site of a complex zonated oceanic circulation and strong fronts. The EPIC experiment is examining the coupling processes in the atmospheric and oceanic boundary layers and this may lead to design of long-term measurements that might be added to PBECS. For the time being, PBECS and its predecessor, the Consortium on the Ocean's Role in Climate, will serve mainly to support process experiments in this region.

The ocean along the west coast of North America is the site of substantial ocean and atmosphere variability, particularly resulting from the ENSO cycle (see Section 8.2). The role of local air-sea interaction in propagating El Niño signals along the coast, and their affect on local weather, is not known, but their broad offshore scale and the fact that they propagate at speeds quite different from free coastally trapped waves suggests a coupled phenomenon.

22. In cooperation with coastal modeling, weather, and ecology studies, it is recommended that an array of air-sea flux and upper-ocean moorings and offshore VOS or autonomous glider sections be established and maintained along the U.S. west coast. Two moorings and three lines would be appropriate for PBECS use.

The eastern boundary region along South America is the site of coastal upwelling and prominent decks of stratus and stratocumulus clouds that have substantial effects on radiation. The cloud decks are thought to involve a positive feedback by blocking solar radiation and cooling SST with this, in turn, fostering more stratus (see Section 8.3).

23. It is suggested that repeated offshore autonomous glider sections be maintained between 5°S and 25°S to develop a data set for studying the upwelling source of cold water, its alongshore transport, and their interaction with the ENSO cycle. One to three sections would be appropriate.

11.6 Areas for further study

The community has not had an opportunity to develop and evaluate plans for all suggested parts of PBECS. The plans from other nations that might wish to contribute to PBECS are not yet known. We mention here some suggestions for additional components of PBECS that deserve to be explored to the point where their utility and the needed resources can be evaluated. This list is likely to grow as the community becomes aware of PBECS and how it can accelerate study of climate variability in the Pacific sector.

As discussed in Section 4.7, the Argos array and VOS sections will not obviate the need for repeat hydrography in tracking slow climate-related variability. If the nations of the Pacific were to collaborate on this, it is likely that a basin-scale survey of five to seven lines might be repeated on a roughly decadal period with reasonable resources.

While the general strategy for improving air-sea flux fields over the Pacific is clear, there has not been enough study to determine exactly how many reference flux sites should be maintained or how the connections with the satellite and operational modeling communities are to be made. A careful examination of the options is needed.

Time-series stations also provide a unique perspective on ocean climate change. The HOT project north of Hawaii is the longest time series now being maintained. The reasons for maintaining HOT, and perhaps implementing new time series, need to be enunciated and reviewed.

The dynamics of the equatorial processes like diapycnal mixing, upwelling, frontogenesis, and the shallow equatorial overturning circulation are central to the evolution of the ENSO cycle but are poorly represented in climate models. A strategy for describing these processes during the ENSO cycle should be developed and implemented.

12. Acknowledgments

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