# A process study to observe upwelling in the equatorial Pacific

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## 1. Introduction

The purpose of this document is to focus discussion on the goals and methods of a process study to observe vertical exchange in the upwelling regime of the equatorial Pacific. By upwelling, we mean the entire complex of processes that connect the thermocline to the surface and thereby allow interaction of ocean dynamics and property transports with the atmosphere. Upwelling includes vertical velocity, along the upwards-sloping isopycnals along the equator as well as diapycnal transport induced by surface divergence and balanced by turbulent vertical exchanges and surface fluxes. Heat transport is of particular interest because of its role in ocean-atmosphere feedbacks, but vertical mixing of momentum is also a vital parameterization in ocean models that plays a major role in their low-frequency evolution. An ability to correctly model the vertical structure of the ocean response to varying equatorial winds would also foster accurate simulation of upper ocean biological productivity and its connection to $CO_2$ uptake/outgassing by the ocean. In both simple and GCM ENSO models, the subsurface memory, communicated to the surface by (possibly parameterized) upwelling, is the dominant source of interannual oscillation. Finally, mass and property transport from the thermocline into the diverging equatorial surface layer is a crucial link in the meridional overturning circulation in the Pacific.

As briefly reviewed below, previous observations have sampled the horizontal divergence on one hand, and the equatorial microstructure on the other. A novel experiment would place turbulence observations in meso- and large-scale context, allowing diagnosis of the complete set of processes for a limited period of time, and contribute to development of model parameterizations of vertical exchange that take into account all relevant factors.

## 2. Observational background

### *a. Vertical velocity*

Upwelling has been identified as a fundamental element of the circulation of the equatorial Pacific (and Atlantic) since the pioneering investigations of Cromwell (1953), Knauss (1963) and Wyrtki (1981). By 1981, its signature in the three-dimensional distribution of water properties was described, its dynamical origin in a meridional circulation forced by zonal winds and Ekman divergence was understood, and its significance to the heat balance was recognized. Early estimates of upwelling speed above the equatorial undercurrent (EUC), based on displacement of isotherms, were a few $\mathrm{m\,day^{-1}}$, a value entirely consistent with modern estimates.

Several attempts have been made to estimate vertical velocity at the equator from continuity, based on divergence of moored horizontal current measurements. While such observations allow excellent temporal sampling, a caveat is that errors accumulate in the downward integration, in practice making conclusions about w in the EUC core and below only tentative. Halpern and collaborators deployed triangle arrays of moorings at 110°W and 152°W for partly overlapping periods of more than a year during 1979-81, with zonal and meridional spacing of 1° or somewhat less. Mean w for a 94-day record in early 1979 at the 110°W triangle was $2.2\,\mathrm{m\,day^{-1}}$ above the EUC core, and was due both to the zonal and meridional components (Halpern and Freitag 1987). Taking du/dx between the 152°W and 110°W moorings, and averaging dv/dy from the two arrays, resulted in maximum mean upwelling of nearly $4\,\mathrm{m\,day^{-1}}$ over 15 months, due almost entirely to the v-field (Bryden and Brady 1987). An array of four moorings in a 2°

latitude/longitude diamond during the TIWE experiment at 140°W for 13 months in 1990-91 gave mean w above the core of about 2 m $day^{-1}$; while it may not be significant, these TIWE observations also suggested equally strong downwelling in the lower EUC and below (Weisberg and Qiao 2000). Although dv/dy (divergent above 50m and convergent below) was the principal determinant of dw/dz during TIWE, the divergence of u above about 150m, where the zonal speed at a given depth increases eastward as the EUC rises, also contributed. Combining the TIWE velocities with large-scale isotherm slopes dZ/dx between the TAO moorings at 170°W and 125°W gave an estimate of effective cross-isopycnal velocity ($w_c \sim w - u dZ/dx$); this was near zero at the EUC core (about 21°C) and grew upward to more than 2 m $day^{-1}$ near the surface. The TIWE dataset has provided the only opportunity to date to produce time series of vertical velocity profiles. These time series showed fluctuations an order of magnitude larger than the mean, in many cases varying coherently over the upper 250m. These variations were associated with remotely-forced intraseasonal Kelvin waves and tropical instability waves (TIW), as well as with variations of local zonal wind at low and high frequencies.

Repeated sections using shipboard acoustic Doppler current measurements can also produce estimates of w from horizontal divergence that (barely) rise above estimated errors, and then only above the EUC core. The significant difficulty with infrequent shipboard surveys is aliasing by TIW, which make it very difficult to adequately sample v or its derivative. On the other hand, shipboard sampling shows the complex meridional structure that has not been available from moorings. Terms of the zonal momentum balance were estimated from 15 monthly cruises (1979-80) in the Hawaii-Tahiti Shuttle near 150°W. These implied that upwelling was centered 1°-2° north of the equator with largest values of about 1.5 m $day^{-1}$ (Johnson and Luther 1994). The ships servicing the TAO array have accumulated considerable ADCP data along regular lines repeated about twice per year each. 85 of these cruises during 1991-99 were combined to estimate horizontal divergence as a zonal average over 170°W-95°W (Johnson et al. 2001). TIW aliasing meant that averaging over smaller longitude ranges was unable to produce meaningful v values. Time-mean vertical velocity stood out above the errors near 50m depth, with a value of about 2 m $day^{-1}$ on the equator, but could not be estimated with confidence in the EUC or below. This unique data set, with repeated direct current measurements over a wide region, gave total vertical transport of 60 Sv across 50m depth over the east-central equatorial Pacific. Both the studies based on shipboard ADCP data showed a strongly banded structure in latitude, with off-equatorial downwelling, that suggests the need to carefully consider the spatial sampling of future observations.

Drogued surface drifter velocities have also been used to estimate horizontal divergence, and generally come up with similar few meter/day values for w into the surface layer, with transports of 30-50 Sv (Hansen and Paul 1987; Johnson 2001). One study binned drifter velocities on a 5 km meridional scale, averaged over 150°W to 90°W, and suggested that the decade-average upwelling occurred in a 20 km-wide band right at the equator (Poulain 1993). If the long time and space average values are this narrow, presumably individual events have even smaller scales. This result has not been confirmed by other techniques, but is another cautionary note regarding spatial sampling strategy. We note that high-resolution satellite microwave SST does not show a variance maximum at the equator that might be expected if upwelling was in fact this narrow.

Both shipboard ADCP and drifter velocities show a region of surface convergence and inferred downwelling at about ±4° latitude, a result that is also commonly seen in models. This occurs because Ekman transport decreases rapidly with increasing Coriolis parameter. The dowwelled water flows geostrophically towards the equator and then upwells to form "tropical cells" (McCreary and Lu 1994) apparently within the relatively homogeneous water mass above the thermocline. This suggests that some fraction of equatorial upwelling is in fact a narrow recirculation of warm water and may not be involved in the large-scale overturning of the subtropical Pacific. It also means that estimates of upwelling from box models based on geostrophic and Ekman transport across, say, ± 5° latitude (discussed below) are not resolving this complex meridional structure.

Vertical transport averaged over a region can also be estimated (in the mean or low frequency) by indirect methods, based on divergence of geostrophic and (assumed) Ekman transports. The simplest type of estimate is a box surrounding the equatorial region. Geostrophic and Ekman transport across the poleward edges is estimated from zonal isotherm slopes and zonal winds, and some assumptions or estimates made of the zonal flows at the east and west edges. Wyrtki (1981) assumed that all the meridional imbalance went into upwelling, and found a net vertical transport of 50 Sv within 170°E-100°W. Bryden and Brady (1985) additionally used equatorial geostrophy to estimate zonal currents at the east and west edges of their box (150°W to 110°W), and also assumed that the flow was horizontally nondivergent poleward of ±0.75° latitude. That gave a profile of vertical velocity averaged over 0.75°S-0.75°N (about 3 m $day^{-1}$ upward above the EUC and about 1 m $day^{-1}$ downward below); this results in a maximum upwelling transport of about 22 Sv. Most of this was due to poleward Ekman divergence. Comparing the ratio w/u with the zonal isotherm slopes gave velocities normal to isotherms smaller than the along-isotherm upwelling by a factor of at least three, and generally in the same sense as total w. Meinen et al. (2001) took advantage of the measured (ADCP) currents from the TAO service ship to correct zonal transports estimated from geostrophy. The upwelling transports within their 165°E-95°W, 5°S-5°N box agreed reasonably well in magnitude and vertical structure with Bryden and Brady's estimate, scaled for the area of the box. With eight repeated meridional sections, they were able to define, as well, the varying depths of isothermal surfaces, and thus to find effective diapycnal transports, which were downward across the EUC and upward above 100m. Time series of vertical transport within their box showed reasonably good correlation of upwelling with local easterly winds on both seasonal and interannual timescales. Internnual SST was well correlated with inferred w, but at the seasonal cycle it was not, implying that $w\partial T/\partial z$ is not the dominant term of the surface layer heat budget in the annual cycle. The Meinen et al. vertical transports were roughly consistent with those required to explain the depleted distribution of $^{14}C$ above the thermocline at 150°W (Quay et al., 1983). However, the profile of diapycnal transport remains difficult to assess; a recent inverse model calculation combining CTD and ADCP data from the TAO servicing cruises implied that diapycnal transport was upward across the whole depth range of the EUC, and the source of thermostad water from below contributed to the downstream increase of EUC transport in the central Pacific (Sloyan et al., 2003).

Clearly, though there is general agreement that total vertical velocity above the EUC core is a few m $day^{-1}$, and that cross-isopycnal velocity $w_c$ below the surface layer is somewhat smaller than that, we have a wide range of estimates of the $w_c$ profile, and considerable doubt as to its sign except at very shallow levels. Cross-isopycnal transport requires either heating from above (for example through penetrating radiation), or turbulent exchanges, which we are just beginning to measure and understand.

*b. Turbulent mixing*

The earliest microstructure measurements in the equatorial undercurrent region showed low mixing in the current core but strong mixing in the high shear zones above the EUC (Gregg 1976). This zone often produces low mean Richardson numbers in stratified water below the surface layer. There have been three sets of measurements of microstructure in the central equatorial Pacific in recent decades, which sampled different regimes of the ENSO and seasonal cycles. Tropic Heat I took place in late 1984, during neutral ENSO conditions with a strong EUC and TIW (Gregg et al. 1985, Moum and Caldwell 1985). Tropic Heat II took place at the end of the 1986-87 El Niño (Peters et al. 1991), and was during the warm season (April 1987), so TIW were weak. The TIWE experiment at 140°W, whose divergence measurements are discussed above, also included a microstructure survey during Nov-Dec 1991 (Lien et al. 1995). The onset of the El Niño of 1991-92 meant that TIW were absent, and Kelvin waves produced thermocline oscillations of 50m or more within a few days, with correspondingly large zonal current fluctuations. Unfortunately, the TIWE divergence measurements and microstructure were not analyzed as a whole. Diapycnal diffusivity inferred from all these microstructure measurements agreed in showing low mixing at the EUC core, but above the EUC the profiles of $K_\rho$ were signficantly different, for reasons that are not understood, but could easily reflect the quite different regimes sampled. Other observations have taken place in the quite distinct regime of the west Pacific warm pool during the TOGA-COARE experiment.

In the central Pacific, because of the complexity and time-dependence of the circulation in a short vertical range above the sharp, shallow thermocline, including strong vertical shears, strong meridional oscillations due to TIW, large, rapid vertical motion of the thermocline due to remotely-forced Kelvin waves, a large solar radiation term that can involve significant penetration through the wind-mixed layer, a strong and variable diurnal cycle that can change the surface convecting layer thickness by a factor of two or more, and the order(1) SST changes associated with El Niño events, it has proven difficult to draw unambiguous conclusions from the small number of regimes sampled. In addition, measurements to date have mostly taken place right at the equator, though some theory and sparse observations suggest that internal wave properties may change rapidly with latitude because of the gradient of f. We have much less information about either the shear regime or the microstructure just off the equator.

Although we have the tools to measure microstructure from ships for the duration of a research cruise, the wide diversity of turbulence regimes to be looked at in the equatorial region poses a challenge that is just beginning to be met. It is unlikely that we will be able to measure mixing everywhere it is important. Thus, the task of developing useful parameterizations of diapycnal fluxes in the equatorial upwelling region will require a combination of sampling internal wave properties within a detailed observational context, and the use of internal wave models tuned by these observations.

## 3. Role of upwelling in Pacific climate:

From a climate perspective, it is upwelling's role in determining SST that is important. Upwelling is both a response to local winds and a component of the gyre-scale circulation. Each aspect affects SST. In general, the local wind determines the rate of vertical exchange, and how deep it extends into the thermocline, while the gyre-scale circulation determines the background stratification and the properties of the water that is upwelled. These properties are a boundary condition for the SST budget. Indeed, the relation between SST and thermocline depth used in simple ENSO models is shorthand for a complex heat budget in which upwelling temperature transport is the main cooling effect.

*a. Basin-scale*

Equatorial upwelling is the path by which water in the subtropical overturning circulation (subtropical cells or STCs) returns to the surface, where it can be Ekman-advected back to the subtropical gyres. The historical studies reviewed above suggest that the net vertical transport averaged over the equatorial region (and over some long zonal distance) is given by Ekman divergence which is reasonably well-described by the wind. Given that gross picture, important questions remain concerning the role of upwelling in the overturning circulation, and the constraints that the STCs impose on water available for upwelling. The vertical mixing parameterizations in OGCMs determine how the models produce the divergence required by Ekman transport. Typically models are tuned so the equatorial temperature and velocity profiles resemble observations, which does not necessarily result in the correct near-equatorial meridional structure. This has implications for the types and properties of upwelled water, and for model sensitivity to wind anomalies.

Layer models of the STCs, an equatorward extension of ventilated thermocline ideas, have dynamics dominated by Rossby characteristics distorted by the Sverdrup circulation (Lu et al. 1998). These models suggest that surface water transported poleward across a subtropical latitude by Ekman flow is replenished by upwelling from the equatorial thermocline, which is the principal means for producing new surface water in the tropics. The properties of the upwelled water are determined by subtropical subduction injecting water into the thermocline, and advection follows Rossby characteristics starting from the subtropical mixed layer, eventually ending up at the equator via possibly-circuitous routes. Thus, while the slope of the equatorial thermocline is in balance with the local zonal winds, its stratification and zonal mean depth is due to winds and surface fluxes over a much broader range of latitude. The layer models suggest that changes in either subtropical subduction, or the ray paths of subtropical Rossby waves, or the magnitude of gyre transport, can affect the properties of the equatorial thermocline and thus the water available for upwelling. The picture is complicated by the presence of the much narrower tropical cells, that may recirculate a substantial fraction of the equatorially-upwelled water above the thermocline, but their role in the heat and momentum balances are poorly known.

Potential vorticity constraints inhibit water parcels from crossing the equator, except in the presence of strong friction, mixing or nonlinearity. Models (Blanke and Reynaud 1997) and observational diagnoses (Sloyan et al. 2003) show that the redistribution that takes place when equatorial water is upwelled, mixed laterally across isopycnals at the surface, and then transported poleward can be a mechanism for interhemispheric exchange, which is part of the Indo-Pacific circulation around Australia. Better understanding of the influences on this exchange would advance diagnosis of how changes in equatorial winds affect the intergyre and interbasin exchanges.

The meridional shear between the EUC and SEC branches is strong enough to modify the effective $\beta$ ($Q_y = \beta - u_{yy}$), which may in fact change sign at thermocline level near ±2°-3° latitude. Does this need to be taken into account in estimating the Ekman transport? A related issue whose resolution would improve models is the thickness of the “Ekman layer” near the equator and the vertical structure of directly wind-driven ageostrophic flow in the region. Because we have only rudimentary understanding of how wind momentum is communicated from the surface to the thermocline, especially near the equator where concepts like the Ekman layer are uncertain, it is difficult to judge the adequacy of present-generation model parameterizations of the vertical structure of vertical diffusivity.

*b. Local*

Local aspects of vertical exchange include processes that maintain the surface mixed layer in an order (m $day^{-1}$) vertical velocity field, mechanisms of vertical penetration of surface fluxes of heat and momentum, and interaction of upwelling-induced SST changes with the wind. The strong turbulence observed in stratified water below the isothermal layer is distinct from other open-ocean areas, and suggests that the equatorial surface “mixed layer” may have different dynamic and thermodynamic balances than seen elsewhere.

One of the most important observations has shown the occurrence of deep daily cycling of equatorial turbulence driven by surface fluxes (Lien et al. 1995); this implies that models of vertical exchange must either resolve the diurnal cycle or learn to parameterize it. The region of high $K_\rho$ can vary by tens of meters, deepening a few hours behind the nightly thickening of the mixed layer (Gregg 1998). Several mechanisms have been proposed for this turbulence. Internal waves can become unstable and break while propagating downward through the zone where the high shear lowers Ri to 1/4. The generation of the internal waves could be due either to pumping of the mixed layer base by nightly convective plumes, or to strongly sheared horizontal flow along the mixed layer base distorted by these plumes. Another argument maintains that the turbulence can be due not to the diurnal heating cycle but to lowered Ri with increased near-surface shear generated by easterly winds. Results to date, however, suggest that breaking internal waves, however generated, are the key contribution to mixing in the upwelling region.

Since internal wave properties may depend on latitude, but most of the observations have been right at the equator, it will be important to learn the meridional structure of turbulence and its causes. Since at least one analysis of drifter observations (see section 2a) suggests that upwelling may occur very narrowly on the equator, the environment of mixing and vertical exchange may be a strong function of latitude, which is not presently indicated in OGCM solutions, and would have large implications for the response to wind events.

In addition to the mixing down of wind momentum, OGCM solutions show that upwelling advects eastward momentum from the EUC towards the surface. The term $(wu)_z$ is an essential part of the zonal momentum balance, leading to a weaker SEC than would otherwise be the case. The seasonally-varying vertical profile of zonal flow depends sensitively on the balance among westward wind forcing, upwelled eastward momentum (both strongest in boreal fall), and the zonal pressure gradient; we note that the EUC is strongest in boreal spring, when it surfaces east of about 140°W (Yu and McPhaden 1999).

Horizontal mixing, especially across the sharp front near 1°N, must be a significant component of the heat (and salt) budgets that contributes to the maintenance of stratification by mixing in lower-density water from off the equator. TIW are a prominent part of this exchange, but small-scale frontal processes may also play a substantial role, as has been seen in other parts of the ocean where sharp fronts are prevalent. The importance of small-scale gradients suggests that significant improvements in our ability to describe and model the upwelling regime will require much denser sampling than previous 100-km scale buoy programs have provided.

*c. Coupled processes*

Because equatorial zonal winds are sensitive to the zonal SST gradient on short time and space scales, upwelling events have the potential to interact with the wind and thus produce coupled feedbacks. These scales are on the order of 1 day and a few tens of km, illustrated by the large windspeed changes as southeasterly trades blow across the SST front north of the cold tongue (Chelton et al. 2001). This sensitivity is the basis for “SST modes” (Neelin et al. 1998), in which SST anomalies are produced by wdT/dz due either to anomalous upwelling itself or to anomalous thermocline depth that changes the temperature of the upwelled water. Such modes can propagate either eastward or westward, depending on the relative importance of these two processes, and probably contribute to the evolution of El Niño events.

Coupled models consistently show the cold tongue extending too far west, leading to a double ITCZ, as air subsides and diverges in both directions from the equator (Mechoso et al. 1995). Another common CGCM problem is a too-large amplitude of the equatorial semi-annual component compared to observations, and delayed initiation of the westward propagation of the annual cycle (Latif et al. 2001). A possibly-related error that has received relatively little attention is the too-warm equator east of 90°W (Meehl et al. 2001). Observed SST in this region is relatively cool, appearing as a continuous cool surface water mass from the Peru coastal upwelling region to the equatorial cold tongue, and there have been a variety of speculations concerning a possible connection between the two regions (Kessler et al. 1998). Model problems in this region may be due to insufficient spatial resolution near the coast, to poor simulation of the extensive stratocumulus decks in this region, to incorrect horizontal currents on the eastern equator, or to inability to generate cool SST under the weak zonal winds east of the Galapagos; all of these likely have an association with modulation of SST by upwelling. The fact that the westward propagation of the annual cycle originates close to the South American coast suggests that coupled models must simulate oceanic vertical exchanges not just in the open Pacific but include the far eastern region as well, and that observation of the influences on the vertical structure there would be useful. Since the winds east of about 90°W are principally meridional, the near-surface dynamic and thermodynamic balances in this region may be quite distinct from those further west.

The relation between SST and thermocline depth is the key parameterization in simple ENSO models, with the memory carried in thermocline depth the dominant source of oscillation, whether this depth is explicitly due to waves or considered to be in equilibrium with the wind. In coupled GCMs, the vertical diffusivity is found to be a principal factor in the amplitude of their ENSO oscillations, with low background diffusivity producing a sharper thermocline and realistically more intense El Niño events (Meehl et al. 2001). In all these models, the essential subsurface memory is communicated to the surface through variations of either component of wdT/dz. Observations of upwelling in full three-dimensional context would spur advances in validating and improving this aspect of OGCM physics.

**4. Strategies for a field experiment**: **P**acific **U**pwelling and **M**ixing **P**hysics

Since we will not be able to monitor upwelling or vertical exchange continuously in the way that Argo, TAO and altimetry (will) let us monitor the gyre circulation, the ultimate goal of the process study is to provide the observations and interpretation that will let models accurately represent vertical exchanges near the equator. There are two elements to improving models: first, to improve parameterizations through more precise diagnosis of the situation at and near the equator, and second, to learn how to use sparse sustained observations assimilated into models to infer and diagnose upwelling events.

Both horizontal divergence and vertical mixing contribute to vertical exchange, and horizontal mixing (which may be partially resolvable) is also essential for the property and momentum balances; measurements of all these will be necessary. Unlike previous experiments, the process study should integrate its microstructure observations into as full a context as possible, to avoid the difficulties of interpretation that occurred in the past. While it seems clear that resolving mechanisms and challenging modern models requires finer-scale sampling than the 100-km scales of previous efforts, the need for context and representativeness demands substantial spatial and temporal coverage. The balance between these two goals will determine the shape of the experiment.

**A workshop should be held to begin planning a process study. Issues to be discussed include:**

1. Quantities/processes to be sampled:
   a. Vertical mixing, turbulence and property fluxes: mechanisms of diapycnal transport
   b. Horizontal divergence: spatial structure of vertical velocity forced by time-varying winds
   c. Heat/salt budgets: provide large-scale constraints; useful for comparison with GCMs
2. Sampling requirements/implications/technologies for the above:
   a. Temporal
      - Is it necessary to sample an entire annual cycle? If not, what period of the year is most important? What happens if we get an El Niño during the experiment?
      - TIW are a potential major aliasing factor (but also a major contributor to horizontal and probably vertical exchanges).
      - The diurnal cycle will need to be resolved through at least part of the experiment.
   b. Spatial
      - Make best use of the TAO array: historical context, augmentation of existing buoys, regular cruises.
      - East Pacific (110°W) or central Pacific (140°W)? Is this an EPIC follow-on or is central Pacific upwelling more important? (In the east Pacific the southerlies shift upwelling to the south; in the central Pacific zonal winds are stronger).
      - Meridional extent: Need to determine if upwelling is a small-scale filamentary phenomenon or a slow response to spatially-averaged winds. Need to resolve the tropical cells. Does the entire TIW region need to be sampled?
      - Must sample the near-surface (Ekman diverging) velocity!
   c. Uses/constraints of remote sampling: scatterometer, TMI SST, met profiles, altimeter.
3. Models:
   a. Internal wave models to diagnose microstructure observations and test parameterizations
   b. High-resolution (eddy-permitting) models (for OSSEs?) and for checking basin models
   c. Sampling designed to effectively compare/verify/challenge models (10s of km)
4. How to get the most out of integrating the differing types of measurements? CPT approach?

**Straw-man observing plan (to be knocked down and torn apart):**

Place experiment at 110°W. Although the situation may be more complicated than further west, the coupled models suggest that the real problem with the coupled annual cycle is in the east.

<u>Strawman</u> observational elements:

1. Moored measurements:
   - Meridional line of ADCP moorings spanning 3°S-3°N.
     - Spacing: Eq, ±1/3°, 2/3°, 1°, 1.5°, 2° latitude (13 moorings)
     - Additional 4 moorings along equator at ±1° and 5° longitude east and west.
     - (Existing temperature moorings are ±15° longitude away).
     - These moorings must be subsurface to avoid fish and vandals, however sampling of the near-surface velocities is crucial. Will shallow (200m) upward-looking ADCPs do this, or will it be necessary to also instrument surface moorings with point dopplers?
   - Corresponding (?) line of temperature/flux moorings (surface).
     - With point dopplers at 10 or 15m?
     - Do these need to be as dense as the ADCP moorings?
   - These moorings to remain in the water for at least one year.
2. Shipboard measurements:
   - During the cooling time of year (July-September).
     - Fine-structure surveys, concentrating on SST front just north of equator.
     - Observe meridional structure of velocity within mooring array.
     - Observe fine-structure of wind/humidity in relation to SST front.
     - Microstructure sampling program on equator and also around SST front.
3. Drogued surface drifters deployed during ship program.
   - Resolving TIW variations (in concert with TMI SST).
   - Resolving surface Ekman layer (?). Might use drifters drogued at varied depths.
4. Other technologies?
   - ADCP gliders might do a better job of sampling the near-surface velocities.
   - Gliders might provide a broadbrush picture similar to the shipboard obs above, but the observations could be extended for a longer time.
   - Isopycnal floats deployed in EUC core?
   - Tracer release experiments?
   - Remotely sensed ocean color?

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