Intense Currents in the Deep Northeast Pacific Ocean

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ABSTRACT

Observations of deep currents in the northeast Pacific Ocean are reported that indicate that although the eddy kinetic energy level is, as expected, generally low, the deep northeast Pacific is subject to occasional intensely energetic events. These events are energetic enough to dominate the distribution of kinetic energy in the water column and the depth-averaged kinetic energy. The assumption can no longer be made that deep flows are weak when estimating near-surface flows.

1. Introduction

Eddies are now known to be a ubiquitous feature of the circulation systems of all of the world's oceans. A general understanding of the distribution of eddy scales and energy levels is necessary to allow a parameterization of horizontal mixing processes and the transport of heat in the atmosphere/ocean system. Of all of the oceanic regions of the world, it is expected that the northeast Pacific deep water should exhibit the lowest eddy energy levels because of its remoteness from intense boundary flows and the absence of deep convection. However, here I report some observations of the oceanic eddy field in the northeast Pacific that indicate an eddy energy level that is generally weak but exhibits occasional periods of very high energy level. The peak speeds observed near the bottom can exceed near-surface current speeds. This type of intermittency in the energy level was unexpected and is presently unexplained. Some possible mechanisms for generation of these events have been conceived, but at the present time we have no obvious means for testing these hypotheses.

The velocity field in the deep northeast Pacific Ocean has received little attention in the past from physical oceanographers. A comprehensive summary of current meter observations worldwide to January 1989 (Dickson 1989) vividly demonstrates the dearth of observations in the northeast Pacific. To a large extent this is because the area is remote from the active western boundary and even remote from the Kuroshio Extension region. Furthermore, the Pacific Ocean has few marginal seas. The densest water formed anywhere within the Pacific Ocean and its marginal seas is probably formed in the Sea of Okhotsk or off the coast of Kamchatka. In neither case is the water formed particularly dense, and after formation probably ventilates the Pacific only down to depths of around 800 meters (Talley 1991). In contrast, the Mediterranean outflow ventilates the Atlantic Ocean down to depths in excess of 1500 m. Thus, it is naturally anticipated that the eddy field will be extremely weak and that all we need to do is map the general circulation using temperature and salinity observations. This has been done (Mantyla and Reid 1983; Tabata 1965), and normally the geostrophic circulation maps are computed relative to a depth of 1000 m (in the Atlantic 2000 m is more commonly used) because of a general prejudice that deep currents are weak in the northeast Pacific (e.g., see Fig. 2 of Mantyla and Reid 1983). Of course, the expectation that the general circulation can be mapped with hydrographic observations alone derives from an assumption that the barotropic velocity component is small. If it is not small, then the velocity field must be measured directly. The results of Cummins and Freeland (1993) suggest that the northeast Pacific can support a significant, but not dominant, barotropic component of flow.

The few direct observations of current speeds that have been made tend to support the low energy view of the northeast Pacific. Observations near Hawaii (Taft et al. 1981) in 1974 are remarkable in that Aanderaa current meters were below the threshold at which the instruments can measure currents (i.e., less than 1.5 cm s^{-1}) for typically 50% of the deployment time, and in one case a Geodyne current meter was below threshold for 97% of the deployment time. More recent observations on 35°N and 39°N on longitude 152°W (Niiler and Koblinsky 1985; Schmitz 1988) verify this general picture to some extent and suggest eddy kinetic energy (per unit mass, hereafter understood) levels of $K_E = \frac{1}{2}(u^2 + v^2) = 1$ cm^2 s^{-2} in the abyssal waters of

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the northeast Pacific. However, the arrays deployed are strongly biased toward observations west of the date line and do not extend very far north, staying entirely within the California Current system south of the west-wind drift and the subarctic front. Two studies (Holloway 1986; Freeland 1987) of eddy transports using satellite altimetry present figures that indicate that the rms meridional velocity in the northeast Pacific is more like 7 cm s^{-1}. This was derived by relating surface height fluctuations to a near-surface streamfunction field. Assigning a fraction of the variance observed to surface intensified flows and part to flows distributed over the water column leads to an estimated rms value of the depth-averaged streamfunction ψ of 3.8×10^7 cm^2 s^{-1}. This is related to rms speed via "a characteristic eddy scale." Picking a reasonable value of 50 km for the eddy scale leads to a depth-averaged rms meridional velocity of 7 cm s^{-1}. Though this is averaged over the whole water column, it is considerably larger than the values suggested by earlier observations. This is an important discrepancy to clear up. It is well known (Bennett and White 1986; McBean 1991; Michaud and Derome 1991) that the northward transport of heat in the atmosphere–ocean system, derived from external radiation budgets, is apparently not equal to the sum of the transports observed in the atmosphere, North Pacific, and North Atlantic oceans. Given the paucity of knowledge of the circulation in the North Pacific it seems appropriate to search there for the source of the imbalance.

2. Observations

In June 1989 a single mooring was deployed at 49°33′N and 138°38′W. This lies in an extensive region of very flat and featureless topography 1000 km off the west coast of Vancouver Island (see Fig. 1) with

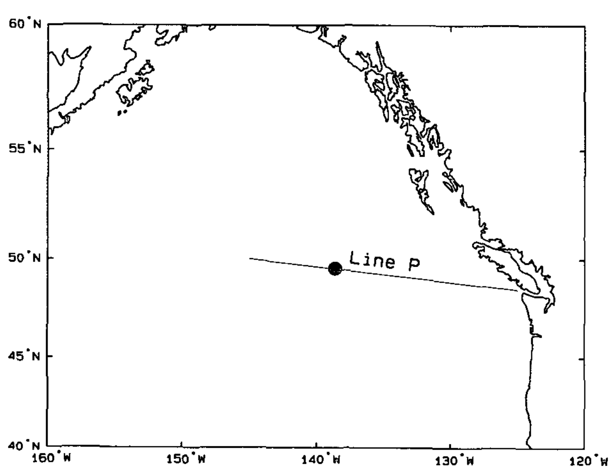

FIG. 1. Map of the northeast Pacific Ocean showing the location of the current meter mooring site, marked with the dot, along the line of CTD observations known as Line P, which extends from the mouth of the Juan de Fuca Strait to ocean weather station Papa.

an average water depth of 3980 m. There are no seamounts within 100 km of the mooring site. The mooring was equipped with Aanderaa RCM4 and RCM5 current meters at depths of 200, 600, 1500, 2200, and 3000 m. Since the initial deployment the mooring has been serviced on five occasions. At those times the mooring was recovered and then replaced with previously prepared current meters. The data return has not been impressive; however, at two depths (1500 and 3000 m) we have unbroken time series of hourly observations extending over more than two years. The last servicing operation was in October 1991 and the mooring remained in place while this was written. As data were acquired from each deployment, the time series were joined. Small gaps do occur at the servicing times; gaps of less than 48 h were filled by an interpolation scheme that fits a mean, trend, and diurnal and semidiurnal oscillation to data extending 48 h on either side of the gap. The hourly observations were then filtered using a Lanczos-cosine filter (Freeland 1983). This is a short filter having a half-width of 60 h that efficiently eliminates the diurnal and semidiurnal tides but leaves motions at periods longer than 40 h effectively unmodified. In the data to be presented here no other data processing of any kind has taken place. At the latitude of this mooring the inertial period (2π/Coriolis parameter) is 15.7 h and so lies between the semidiurnal and diurnal tidal bands. Thus, the water column cannot support internal motions at diurnal periods and the dominant diurnal tides are expected to be barotropic. Tidal analyses have been completed, using software reported by Foreman (1978), and the observations at this mooring site verify that about 95% of the energy at the K_1 tidal line falls into the barotropic mode. The result is that the diurnal tides extend to the bottom of the ocean and keep the rotors on the deep current meter turning. For this reason few of the hourly current observations fall below the threshold for these instruments (1.1 cm s^{-1}). The observations at all depths for the first year of deployment have been reported (Cummins and Freeland 1993) and a decomposition into barotropic and internal modes compared with the predictions of a numerical model. The comparison of observations of overall energy levels and the partition into modes with the predictions is excellent.

Figure 2 shows a plot of the E–W and N–S components of velocity at the deepest (3000 m) level. For most of the deployment time the low-passed current speeds are typically 1.5 to 2 cm s^{-1}. However, near the end of July 1990 current speeds increased rapidly to a peak value of over 10 cm s^{-1} (low passed). This event lasted about 50 days and, as can be seen in Fig. 2, appeared as a burst of flow to the south. Later, in January to March 1991 other smaller events were observed including a strong northward flow. It is tempting to imagine that we are seeing an eddy pass over the mooring site producing the southward flow in August followed by northward flow several months later as

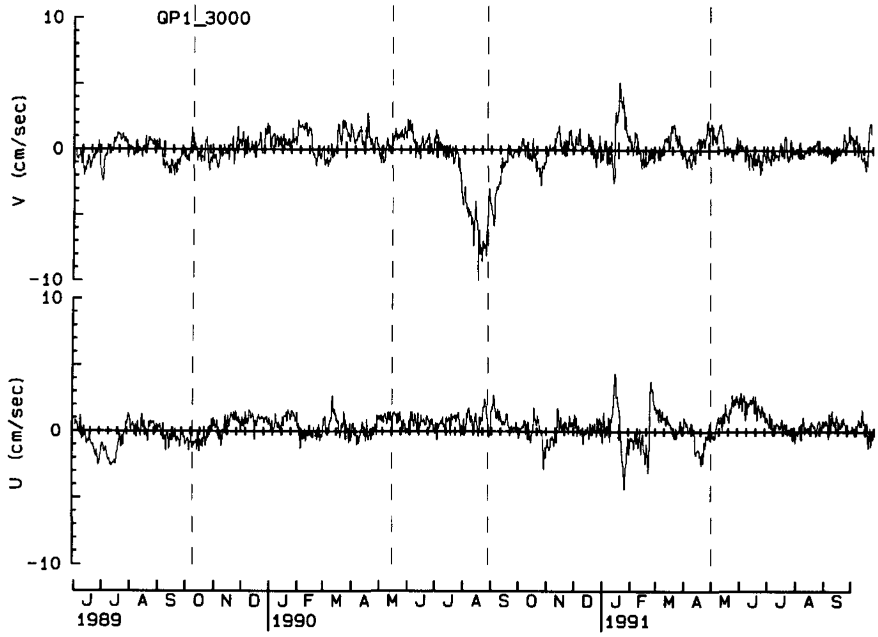

FIG. 2. Components of velocity measured at the mooring location identified in Fig. 1 at a depth of 3000 m. The component v is in the north–south direction (positive to the north) and the component u is in the east–west direction (positive to the east). The magnitude of each component is plotted against time over a period exceeding two years. The vertical dashed lines indicate the times of mooring recovery and servicing operations.

first one side and then the other passes over the mooring site. However, we have no evidence that suggests that these two events are associated with each other in any way. The first response when this event was seen was total disbelief. However, as indicated in Fig. 2, the mooring was serviced during the high speed burst. The current meters before and after the mooring operation were different instruments and yet the event is continuous. We have not been able to find any error of calibration in the instrument and can conceive of no error that would produce this anomaly. As a check on calibration of speed, the amplitudes and phases of the K_1 tidal line were computed at all depths and for each deployment. One calibration error was detected (and corrected) by this method and the resulting K_1 fit appears to be barotropic, within reasonable bounds, and essentially identical between deployments. Temperature was recorded on both of the instruments that recorded this event. However, three problems arise: first, deep temperature gradients are very weak in the central North Pacific; second, the temperature sensors on Aanderaa RCM5 instruments are extremely coarse (in fact, the total range of temperature variations observed at 3000 m is only about 4 bits); third, the calibration of instruments is imprecise. Conductivity measurements were not made successfully until the fourth deployment. Thus, there may be a sigma t or a temperature signal associated with the passage of this event, but we are unable to detect it.

Figure 3 shows currents at a depth of 2200 m at the same current meter mooring. Data return at this site was actually quite poor; during the early deployments, and so the horizontal scales of Figs. 2 and 3 are different. Once again current speeds are low for most of the time, near the 1.5 to 2 cm s^{-1} expected. However, for a period of 80 days between April and June 1991 another current event occurred. As with the event at 3000 m this event was also interrupted by a mooring recovery and redeployment, and the instrument at 2200 m was replaced. Thus, we again have corroboration that this was a real event rather than an instrumental malfunction. The detailed behavior of the velocity variations is strongly suggestive of an eddy moving directly over the current meter mooring site. For example, an anticyclonic eddy (clockwise rotating) passing from north to south over the mooring with the eddy center somewhat to the east of the mooring site will produce the observed pattern of flows that are initially to the west, then north, and finally to the east. (Of course, the same pattern of currents will occur with a cyclonic eddy passing from south to north over the mooring with the eddy center somewhat to the west of the mooring.) At the 2200-m depth we were more successful with temperature and conductivity measurements. There is a drop in salinity of about 6 bits (or 0.16 psu) and a rise in temperature of about 3 bits (or .06°C) as the eddy passes. This is consistent with water parcels being drawn down to the 2200-m depth level. However, the obser-

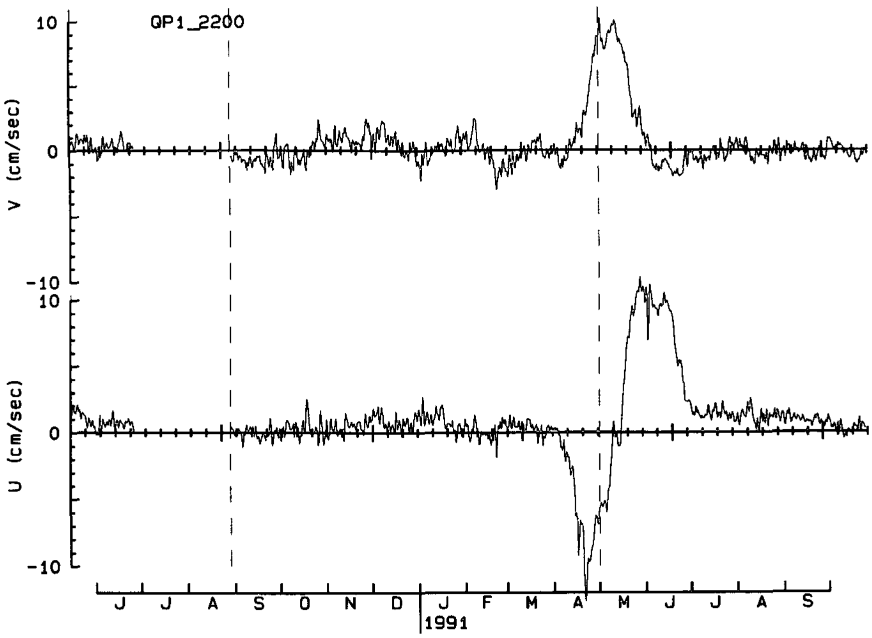

FIG. 3. As Fig. 2 but at a depth of 2200 m.

vations are extremely imprecise and cannot reasonably be expected to yield quantitative estimates of the distance parcels might have been drawn down.

A velocity disturbance is evident at the 1500-m depth level that matches the timing and shape of the big event of April to June 1991; however, the amplitude is much reduced. This is, however, the only incidence we have of the deep current bursts being observed at this level. We conclude therefore that the bursts have a very short vertical scale of 800 m or less.

3. Discussion

It appears that the kinetic energy density in the Pacific Ocean is generally weak but is occasionally disturbed by episodes of extremely high kinetic energy. We cannot generally assume that motions at 2000 or 3000-m depth are weaker than those near the surface. The observations at the site in question give values for the eddy kinetic energy K_E of 3.7, 1.0, 0.8, 7.2, and 1.6 $cm^2 s^{-2}$ at 200, 600, 1500, 2200, and 3000 m, respectively, where $K_E = \frac{1}{2}\overline{(u^2 + v^2)}$. These values do not fit with the pattern for the North Pacific exemplified by Schmitz (1988). If we had missed observing these unusual deep events, then the K_E at 2200 m would be reduced from 7.2 to 0.8 $cm^2 s^{-2}$. At the present time we know only that occasional events do occur; however, we have no idea how often they occur at any particular depth. This is an important omission. The objective of the work mentioned earlier by Holloway (1986) was to use the estimated depth-averaged rms current speeds to estimate a depth-averaged eddy diffusivity via a mixing length theory. This was then applied to estimates of the depth-averaged temperature field to estimate the contribution that barotropic eddies might have on large-scale heat transports. The figures given above indicate estimates of the depth-averaged kinetic energy of 2.6 or 0.7 $cm^2 s^{-2}$, depending on whether or not the bursts are included. The larger number is large enough that it might be necessary to consider the depth-averaged fluctuations in estimates of the large-scale heat transport, but it is substantially smaller than the value implied by Holloway's estimates.

Finally, we should give some thought to the possible mechanisms by which such motions might be generated. Superficially, the observed northeast Pacific events look a lot like the "Meddies" reported in the North Atlantic (Armi et al. 1988); however, there is no marginal sea adjacent to the North Pacific that generates dense water masses the way the Mediterranean does. As mentioned earlier, the densest water formed anywhere in the North Pacific probably ventilates to a depth of only 800 m. Thus, deep injection of dense water will not serve here as a credible mechanism. To generate an intense eddy we require a large amount of vortex stretching. The northeast Pacific has a generous scattering of seamounts with a wide variety of summit depths. An eddy might be formed when a steady flow is initiated past a seamount or conceivably by the collapse of a volume of fluid abruptly mixed over a seamount. In the former case abrupt vortex stretching as a column of fluid is swept off a seamount will produce a cyclonic eddy. In the latter case, if we mix an isolated volume of water in the center of a stratified water column, this mixed fluid will collapse back to the mean density level producing vortex shrinking and generating an anticyclonic eddy. An alternative mechanism, however, is suggested by the observations of a deep silica maximum in the North Pacific reported by Talley and Joyce (1992). In this paper they describe the ventilation of the deep northeast Pacific by buoyant, silica-rich water vented from the hydrothermal vents of the Juan de Fuca Ridge. In their model we apparently see buoyant water injected at the bottom in 2500 m of water rising to a neutral density level 300 m above. Thus, we have a deep buoyancy source that behaves like the inverse of the Mediterranean ventilation of the northeast Atlantic Ocean. The same vorticity dynamics would apply, and it seems reasonable to expect that the silica plume shown in Fig. 4a of Talley and Joyce (1992) should, like the Mediterranean salt plume, be filled with isolated intense eddies.

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