STRUCTURE OF SPACE-TIME VARIABILITY OF GEOSTROPHIC CURRENTS IN THE SOUTHERN OCEAN

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ABSTRACT

Data collected by the U.S. Navy's Geosat Exact Repeat Mission were used to obtain detailed, statistical descriptions of the variability of the surface geostrophic flow field in a portion of the Antarctic Circumpolar Current in the Southeast Pacific. The descriptions provide, for the first time, the information about the flow long available in other branches of fluid dynamics such as aerodynamics; and are useful for guiding the development of numerical models of the circulation. In particular, we have used Geosat altimetric observations of the variability of the two components of sea-surface slope at crossing points of the Geosat ground track. From the slopes we have calculated time series of two components ($u', v'$) of the surface geostrophic current for a 2.5 yr period at a dense grid of points in the current. We then used the time series at each point to map the spatial distribution of the components of the Reynolds stress $<u'u'>, <v'v'>$, and $<u'v'>$ and their relationship to the mean flow and to bathymetry. The maps show that bathymetry strongly influence the flow, that flow has statistically significant regions of negative viscosity suggesting baroclinic instability, and that the flow has regions of positive viscosity which slows the mean flow. We have calculated the eddy viscosity in the portion of the flow upstream of the Drake Passage. In this region the flow appears to be a free jet; and the eddy viscosity was $8 \times 10^4$ m$^2$/s. We have also calculated the statistical relations among the components of the current, including probability distribution functions and the correlations among the components of the flow lagged in time and space. The correlations show the structure similar to the variability mapped in the North Atlantic by other groups.

INTRODUCTION

Studies of ocean dynamics have been constrained by a lack of a statistical description of the flow that has been available, for example, in such fields as aerodynamics for more than 60 years. Yet such a description of the oceanic mesoscale variability is necessary for understanding the role of eddy variability in ocean dynamics and for developing more accurate numerical models of the oceanic circulation. Among the various statistical functions describing the flow, the turbulent stress and the time and space lagged correlation functions are perhaps the most useful. They are also the most difficult to measure using conventional oceanographic techniques.

Satellite altimeters measure sea level along the subsatellite track; and the level is the sum of the influence of gravity, geoid, tides, and the influence of ocean dynamics, the oceanic topography. The geoid is fixed in time, but the topography has a permanent and a time-variable component. Because the geoid is not well known for wavelengths shorter than 2000 km, the permanent topography cannot be separated from the geoid. The
temporal variability of the oceanic topography can, however, be observed using repeated profiles of the surface. This allows satellite altimeters to measure the time-varying components of the oceanic circulation with wavelengths of 200–2000 km, including the mesoscale variability. We refer to mesoscale variability as eddy variability for simplicity, although it is due to both turbulence and planetary waves.

The study uses data from the U.S. Navy's GEOSAT which carried a precise altimeter that observed the ocean between ±72° latitude for 3 years. Data from the satellite were previously used primarily for describing the geographical distribution of eddies (Fu and Zlotnicki, 1989), their relationship to bathymetry (Sandwell & Zhang, 1989; Chelton, et al., 1990), and the annual and interannual variability of currents (Cheney & Miller, 1988; Shum, et al., 1990). More recent studies have looked at the contribution of eddies to the dynamics of currents (Stammer & Böning, 1991; Le Traon, 1991).

The importance of eddies is well known. The mean-square variability of the horizontal components of the current and the correlations between the components produce stress, the Reynolds stress, which mediate the transfer of momentum in the ocean. Despite the importance of the Reynolds stress for ocean dynamics, it is difficult to measure; and only a few sporadic measurements have been reported (Webster, 1961; Luyten, 1977; and Lukas, 1987). The use of satellite altimeter data for the study of Reynolds stress is even more limited. Tai and White (1990) used Geosat data to map Reynolds stress in the Kuroshio Extension. They found evidence that eddy kinetic energy accelerates the mean flow; but they noted that only ascending tracks of altimeter data were available, and this led to large uncertainty in the accuracy of the calculated stress. Johnson, et al., (1992), investigating the role of the stress in the Antarctic Circumpolar Current, developed an accurate technique for calculating Reynolds stress directly from altimeter data. At the same time, Morrow, et al. (1992) were calculating the distribution of stress in the Antarctic Circumpolar Current using very similar techniques.

The previous work, while limited, is encouraging. Records as short as one year produced useful values of stress having a smooth spatial distribution. This is in marked contrast with laboratory measurements of turbulence which required averages over several hundred to a thousand eddies to obtain statistically stable results. For the ocean, this would require 20–100 year long records for comparable results.

We report here observations of Reynolds stress made using data from the Geosat exact repeat mission. We have mapped the distribution of the stress, and we have used calculated values of stress to study the dynamics of the Antarctic Circumpolar Current upstream of the Drake Passage.

**DATA, FILTERING, AND COMPUTATION OF STRESSES**

Geophysical Data Records from the first 54 cycles of the Geosat Exact Repeat Mission were processed using the techniques described by Zhang (1988), and Sandwell and Zhang (1989) to obtain along-track slopes of the sea surface. The data are from November 7, 1986 to November 17, 1989. No data from the last 14 cycles of the mission nor any data from cycle 51 were used because many observations were missing or anomalous. The processing yielded smoothed values of sea level slope every 32.5 km along the subsatellite track. The use of slopes produced a high-pass filter which removed almost all ephemeris errors, which are predominantly at wavelengths of once per orbital revolution, and long-wavelength errors in the corrections applied to the height data. The smoothing reduces short wavelength noise without reducing the signal due to oceanic mesoscale variability.
The along-track value of slope at cross over points was obtained from a linear interpolation between values closest to the cross-over. The sea-level gradient was calculated from the two slopes at each crossover for each cycle. The geostrophic velocity at the sea surface \( v \) is directly related to the sea-level gradient relative to the geoid \( \zeta \) (Wunsch and Gaposchkin, 1980). Because the local geoid, is not accurately known, only the temporal variability of \( v \) could be determined. Writing the velocity as the sum of a mean \( V \) and a time-varying component \( v' \), we computed \( v' \) from the temporal variability of the slope after subtracting the temporal mean slope at each cross over point from the slope value for each cycle.

The three components of the stress at each cross-over are:

\[
R_{ij} = \rho <u'_i u'_j>
\]

(1)

where \((u'_i, u'_j) = (u'_i, v') = v'\) are the two components of the time-varying surface velocity, \(< >\) denotes the temporal average over all 53 cycles of data, and \( \rho \) is the water density. Note \( R_{11} \) and \( R_{22} \) are the normal stresses, and \( R_{12} \) is the shear stress. The two orthogonal components of velocity and the three stresses were calculated at a \( 32 \times 120 \) grid of points in a large area from \( 51^\circ \text{S} \) to \( 66^\circ \text{S} \) and \( 80^\circ \text{E} \) to \( 170^\circ \text{E} \).

ERROR ANALYSIS OF REYNOLDS STRESS COMPUTATION

The errors in computer values of stress were studied by Johnson, et al. (1992), who considered the two largest sources of error, finding that neither substantially influences the results reported here. The two sources of error are: 1) variability in the flow between observations at a cross over point, and 2) statistical uncertainty in the mean product \(<u'_i u'_j>\).

The first error is proportional to the interval between ascending and ascending passes of the satellite past a cross-over point, which varies between 0.6 and 8.5 days depending on the latitude, and to the decorrelation time of the flow field. We found that the decorrelation time was sufficiently long compared with the time between cross overs that this error is not important.

The second error depends on the statistical significance of the correlation between \( u' \) and \( v' \). We found that \( u' \) and \( v' \) were normally distributed, but that stress was significant at only 20% of the crossover points based on Student’s t-test. This implies that stress is carried by a relatively few eddies, a result consistent with laboratory and field observations of turbulence on a much smaller scale.

LAGGED CORRELATIONS IN TIME AND SPACE

We calculated the averaged, weighted, sample correlations \( \gamma \) lagged in time and space using all observations in the \( 32 \times 120 \) grid of points. The averages were calculated using values of the correlation weighted by the variance of the variable used for calculating the correlation. This gives the most weight to those values with the largest signal-to-noise ratio. The correlations of \( u' \) lagged in time and in the \( x \) direction were averaged over 2500 pairs of cross-over points in time or space.
We found that the correlation function $\gamma$ is nearly exponential in time and space (Figure 1). For the $u'$ component of velocity, $\gamma = 1.0 \exp (-b_x x)$ or $\gamma = 1.0 \exp (-b_t t)$, where $b_x = -0.0144$ km$^{-1}$ and $b_t = 0.0648$ d$^{-1}$. We tested the sensitivity of the result to the time interval between cross overs and obtained the same results for intervals of less than one day and less than three days. The results are also consistent with values for the North Atlantic (Le Traon, 1991) and with the integral time scales observed in the Drake Passage (Inoue, 1985).

Averaged Correlation as A Function of Time and x Lags
Cross-Over Time Less Than 3 Day
Longitudinal Band: 190w - 280w

![Image of 3D plot]

Figure 1. Space-time correlation function for eastward velocity $u'$ for all data for which the interval between ascending and descending passes was less than three days. The correlation is nearly exponential near the origin for lags in time and in the x (east-west) direction.

Because the spectrum of sea-surface slope is the Fourier transform of the slope-correlation function, the exponential shape of the correlation implies that the temporal and spatial spectra $S(\omega)$ or $S(k_x)$ must be of the form:

$$S(\omega) = \left[ \frac{1.0}{b_t^2 + \omega^2} \right]^2 \quad \text{or} \quad S(k_x) = \left[ \frac{1.0}{b_x^2 + k_x^2} \right]^2$$

(1)

where $\omega$ is frequency and $k_x$ is wave number. The result is generally consistent with the spectra of variability calculated by Fu and Zlotnicki (1989) in regions of high variability,
and it provides further support for the idea that the spectrum of eddy variability is well approximated by a power law only for the high frequency or short wave number part of the spectrum. The use of the correlation for calculating spectra has a practical advantage, it allows calculation of variability of each component of velocity using data from a relatively small region. Previous calculations of the spectra required relative long segments of data confined to the satellite's ground track; and the calculations yielded the spectra of only that component of current that is perpendicular to the track.

GEOGRAPHICAL DISTRIBUTION OF STRESS

We mapped the distribution of stress in a broad sector of the southeast Pacific which included the Antarctic Circumpolar Current (Figure 2). All three components of the stress have a spatial distribution similar to that of $<u'u'>$ shown in the figure. The distribution of the normal components agrees qualitatively with plots of the variance of surface elevation published by Sandwell & Zhang (1989) and Chelton, et al. (1990), and with plots of eddy kinetic energy (Shum, et al., 1990).

The plot shows that the distribution of stress is strongly associated with bathymetry. Between 200°E and 260°E the stress is tightly confined to the area between the Eltanin and Udintsev Fracture Zones. East of 230°E, the large stress was almost entirely confined to the region between the subantarctic and polar fronts (Nowlin & Kling, 1986); and by 260°E the axis of high variability centers on 60°S as the current approaches the Drake Passage. West of 220°E the flow is more complex as the current flows between the Eltanin and Udintsev fracture zones and over the Pacific Antarctic Ridge. An area of high variability extends over 4° north of the subantarctic front, and the highest stress occurs between 210°E and 240°E where the flow skirts the Udintsev fracture zone upstream and downstream of the ridge crest, presumably due to turbulence produced as the stream meanders to avoid the shoalest regions immediately above the ridge crest.

In the zone between 238°E and 280°E just upstream of the Drake Passage both the area of high stress and the subantarctic and polar fronts are nearly zonal. Furthermore, the stress becomes weaker in the downstream direction. The flow, therefore, resembles very much a free jet. Such a flow is dynamically interesting, and it can be used for testing theories of the role of eddies in the dynamic. Hence, we chose this region for further study. We are particularly interested in the influence of the stress on the dynamics of the mean currents.

Within this region, we note first that the ratio of $<u'u'>$ to $<v'v'>$. was 1.65±0.10, indicating that the along-stream fluctuations are larger than the cross-stream fluctuation. This compares well with the value of 1.45±0.25 calculated from moored current meter observations in the Drake Passage (Inoue, 1985).

MEAN DYNAMIC TOPOGRAPHY

Calculations of energy transfer between the mean current and eddies and calculations of the work done by stress require knowledge of the distribution of the mean current $U$. Because the geoid is not well known in the South Pacific, the mean current cannot be determined from the altimeter data. We therefore calculated the mean current using observations of density made at deep hydrographic stations. The most useful data were from the R/V Eltanin surveys between 1963 and 1968, and the R/V Discovery survey in 1934 (Figure 3).
Figure 2. Geographical distribution of stress in the Antarctic Circumpolar Current in the Southeast Pacific. The plot includes contours of bathymetry at intervals of 1000 m, the position of the subantarctic and polar fronts, and the distribution of $<u'u'>$ component of stress.
Figure 3. Location of hydrographic stations used for calculating the mean dynamic topography in the region shown in Figure 2. EL-63 denotes data collected by the RV *Eltanin* in 1963 etc., DS are data from the RSS *Discovery*.

To calculate the mean zonal current in the region upstream of the Drake Passage, we calculated the surface dynamic topography relative to 3000 m, then averaged the topography in the zonal direction between 122°W to 78°W using 0.5° bins (Figure 4). We then fitted a smooth curve through the data, from which we calculated the first and second derivatives of the north-south slope. After trying various polynomial and other approximations to the topography $\zeta$, we used a hyperbolic tangent of the form

$$\zeta = a \tanh (by - c) - d$$

where $a = 0.5944$ m, $b = 0.1636$ m/degree, $c = 1.458 \times 10^4$ m, $c = 59^\circ$, and $d = 1.6466$ m are parameters determined by a least-squares fit and $y$ is distance northward measured in degrees of latitude. The curve has the advantage that it defines a zonal jet that closely resembles the observed current. The plot of the normal stress $<u'u'>$ (Figure 5a) corresponds closely with the distribution of mean velocity.
Figure 4. Zonal mean dynamic topography calculated from hydrographic data collected at stations shown in Figure 3. The mean was calculated for data within 0.5° bands of latitude between 122°W and 78°W longitude. The vertical bars give the statistical uncertainty of the mean values.

Using Eq. (2), the mean zonal current is:

$$ U = \frac{g}{f} \frac{\partial z}{\partial y} = \frac{abg}{f} \cosh^{-2}(by - c) $$

(3)

From this, we can calculate the zonal derivative $\partial U/\partial y$ needed in the next section.

**ENERGY TRANSFER**

Eddies can either dissipate or accelerate the mean circulation. The latter is an example of negative viscosity which can result from baroclinic instabilities. Because both processes can occur in the Circumpolar Current, we have investigated the influence of the eddies on the mean flow.

The rate at which the kinetic energy of the mean flow $U$ in a zonal jet is changed by eddy stress is given by (Starr, 1968; see also Webster, 1961):
assuming the flow is bounded so that \( U(y_1) = U(y_2) = 0 \) at latitudes \( y_1, y_2 \) on either side of the axis of the flow, and \( \rho \) is water density.

Zonal Average of Current and Stress

![Graph showing Zonal Average of Current and Stress](image)

Figure 5a. Zonal average of the \(<u' u'>\) component of stress relative to zonal mean current. The stress was calculated from altimetric data; and the mean current was calculated from hydrography using Eqs. (2) and (3). The vertical lines mark the position \( \pm \) the standard deviation of the zonal mean location of the polar front PF and the subantarctic front SAF. The means were calculated using data between 122°W and 78°W longitude. The error bars give the statistical uncertainty of the mean value of each point.

We calculated the integrand in Eq. (4) using \( U \) computed from Eq/(3) and \(<u' v'>\) from the Geosat data. The plot of the zonal averages of \(<u' v'>\) and \( U \) shows that the Reynolds stress and, hence the eddies, tends to decelerate the flow (Figure 5b). We therefore calculated an eddy viscosity \( A_y \) (Pond & Pickard, 1983) using:

\[
- <u' v'> = A_y \frac{\partial U}{\partial y}.
\]  

(5)

This yielded \( A_y = 8 \times 10^3 \text{ m}^2/\text{s} \) with small uncertainty, a value that is consistent with values reported for other oceanic flows.
We also note that the distribution of shear stress (Figure 5b) is not quite symmetric about the maximum value of the mean current. The values between $-66^\circ$ and $-63^\circ$ are positive and statistically different from zero, hence the plot indicates that the stress decelerates the flow near $-58^\circ$ and accelerates the flow near $-63^\circ$. This implies the eddies transfer momentum from the mean flow at the subantarctic front to the mean flow at the polar front. The measured shear stress is sufficiently large that it can transfer most of the momentum from one front to the other in a few thousand kilometers.

**Zonal Average of Current and Stress**

(Longitudinal Band 122°W to 78°W)

![Graph showing zonal average of current and stress](image_url)

Figure 5b. Same as Figure 5a but for the $<u'v'>$ component of the stress. The data were used for computing an eddy viscosity for the flow.

**DISCUSSION**

We have used Geosat data to map the distribution of the variability of two orthogonal components of current in the southeast Pacific and to study the relationship of the variability to mean currents calculated from hydrography.

Maps of variability show that regions of high variability are correlated with bathymetry. The variability is greatest in the region where the flow crosses the Pacific Antarctic Ridge, becoming weaker downstream of the ridge crest. The maps also show that the maximum in the variability corresponds closely with the maximum of the mean current calculated from hydrography, and that it does not have maxima at the the positions of the Polar and Subantarctic Fronts. This is surprising. The conventional view of the dynamics of the Antarctic Circumpolar Current is that it consists of bands of current associated with the two fronts. The latitudinal positions of the fronts are thought to vary in time leading to the smooth slope in dynamic topography in the region. The
Geosat data indicate instead that the variability is greatest in the region between the fronts downstream of the ridge crest.

The observed normal components of the stress are large, that is \( \frac{<u'u'>}{U^2} = \frac{<v'v'>}{U^2} = O(1) \). In fact, the ratio is slightly greater than unity. The large ratio is consistent with variability measured with current meters in various nearby regions of the current (Inoue, 1982; Bryden & Heath, 1985). The shear stress \( <u'v'> \) is considerably smaller, indicating that eddies are not efficient mechanisms for the transport of momentum into or out of the mean flow.

The distribution of shear stress \( <u'v'> \) upstream of the Drake Passage shows that the stress acts as an eddy viscosity. Using the shear of the mean current calculated from hydrography together with the spatial distribution of the zonally averaged shear stress, we calculated an eddy viscosity with small uncertainty. The north-south component of the viscosity \( \nu_x \) was found to be \( \nu_x = 8 \times 10^3 \text{ m}^2/\text{s} \). This is somewhat small, but still in the range of values calculated from other data sets. The low value implies that eddy viscosity is not able to balance the wind stress acting on the current, but it is large enough to influence the dynamics of the observed mean current. There is some indication that eddies transport momentum between fronts in the circumpolar flow.

The maps of the distribution of stress also show that the position of strong variability deviates from the position of the fronts in the region of the ridge crest. This implies the current may have two preferred paths for crossing the ridge crest, the position seen in the Geosat data collected in 1986-1988, and the position observed in the hydrography collected by ships more than twenty years earlier.

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REFERENCES


THE HAWAII OCEAN TIME-SERIES PROGRAM: RESOLVING VARIABILITY IN THE NORTH PACIFIC

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ABSTRACT

Time-series measurements of biological, chemical and physical parameters in the North Pacific Subtropical Gyre have been made on a regular basis since October 1988. These observations now comprise one of the longest sets of time-series measurements in the central Pacific Ocean, and are the only time-series observations available for a variety of parameters in this ocean basin. The time-series observation program has provided a unique view of the extent of variability in the central Pacific, and has begun to improve our knowledge of the dynamics which take place there. The understanding which is emerging from the time-series dataset will be useful for interpreting the historical oceanographic database within the context of future environmental change.

INTRODUCTION

Time-series datasets are rare in the oceanographic literature. This is unfortunate, since it is well-recognized that time series measurements, particularly those made for a decade or more, are powerful tools for understanding and observing the slow process and irregular events that take place in nature (Franklin, 1988). This is now particularly important because of the need to understand the environmental changes that are taking place as a result of anthropogenic influences on the chemistry of the earth’s atmosphere and oceans (Magnuson, 1990). Consequently, time-series measurements in a variety of environments are now especially useful. Time-series datasets are not abundant because long-term observation programs are expensive and are usually difficult to maintain. This is especially true of the ocean sciences where data collection requires costly research vessels and sophisticated sea-going equipment.

The usefulness of time-series observations of the natural world has been amply demonstrated. One of the most prominent examples is the over three decade long record of atmospheric carbon dioxide concentrations compiled by Charles Keeling at Scripps Institution of Oceanography (Keeling, et al., 1982; Moore and Bolin, 1987). This dataset has been invaluable for documenting the anthropogenic increase in atmospheric carbon dioxide concentrations and for developing models of the global carbon budget. Other time-series programs have documented changes in the pH of rainwater in the northern hemisphere (Likens, 1983) and documented other long-term changes in the physical and biological conditions in terrestrial environments (Brock, 1985; Peterson, 1984). Time-series observations in the oceans are more rare. For the most part, time-series datasets in the marine environment have involved measurements at higher trophic levels such as commercial fish and zooplankton (McGowan, 1990). A notable exception includes the long time-series of observations made by the California Cooperative Oceanic Fisheries Investigation (Chelton, et al., 1982). This long-term oceanographic study has included chemical and physical measurements as well as estimates of plankton biomass.
The world's oceans act to modulate global climate and are believed to be the predominant sink for anthropogenic carbon dioxide. However, the magnitude of the oceanic sink in various parts of the world's oceans is uncertain (Tans, et al., 1990; Broecker and Peng, 1992; Sarmiento and Sundquist, 1992). In addition, a detailed understanding of the rates and mechanisms of the cycling of carbon within the interior of the ocean is lacking (Longhurst and Harrison, 1989). In large part, this lack of understanding is due to a paucity of data in both space and time. For this reason time-series measurements of the oceanic carbon system are particularly important for understanding and eventually predicting the magnitude and rates of future climate change.

The need for time-series studies was clearly recognized when in 1987 the International Council of Scientific Unions (ICSU) established the Joint Global Ocean Flux Study (JGOFS) as part of its International Geosphere-Biosphere Program (IGBP) for the study of global change. As a consequence, the JGOFS program called specifically for establishment of oceanic time-series studies at strategic ocean sites which were envisioned as being maintained for at least a decade. As a part of the United States JGOFS study, two time-series were begun in 1988. One of these time-series programs was established at the Bermuda Biological Station where the 35 year time-series established at hydrostation "S" would be continued. A second time-series program was established at a new location in the Central Pacific Ocean near the Hawaiian archipelago (Karl and Winn, 1991).

THE HAWAII OCEAN TIME-SERIES PROGRAM

In November 1988, a long time-series program was begun in the Central Pacific Ocean. This research effort was initiated in an effort to produce a time-series dataset for the central Pacific Ocean. The Pacific time-series was named the Hawaii Ocean Time-Series (HOT) program. The primary purpose of this research effort is to identify annual and interannual variability in the physics, chemistry and biology at a single location in the North Pacific Subtropical Gyre. The Hawaii program is a joint effort between the National Science Foundation's Joint Global Ocean Flux Program (JGOFS) and the World Ocean Circulation Experiment (WOCE). This research effort is coordinated by scientists at the University of Hawaii and was funded initially for a 5 year period beginning in May of 1988. Although the program is presently funded for only a five year period, it is anticipated that the program will continue for at least 10 years.

The time-series permanent station is located 100 km due north of the island of Oahu, Hawaii at 22°45'W and 158°00'N (Figure 1). The location of the time-series station was chosen to minimize the influence of the Hawaiian archipelago (island mass effects) on the biogeochemistry of the water column at the time-series station. The permanent time-series station was therefore located upwind (northeast of the Hawaii Archipelago). Station ALOHA (A Long-Term Oligotrophic Habitat Assessment) is located in 4750 m of water and approximately 50 km (one Rossby wave radius) away from the steep topography associated with the Hawaiian Archipelago. A near-shore equipment test site at Kahe Point is also visited each month.

An extensive suite of measurements are made on each HOT cruise. These measurements include standard hydrographic, and optical measurements using continuous sensors mounted on lowered profiling devices, chemical measurements made at discrete depths collected with standard oceanographic sampling bottles, measurements of the rate of primary production as well as the rate of particle flux, and the measurement of upper ocean currents using an acoustic doppler current profiler (Table 1). The results of these routine measurements are available in tabular form in the
### Table 1. Time-series parameters measured at station ALOHA

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<td>&quot;clean&quot; $^{14}$C incubations</td>
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<tr>
<td>carbon, nitrogen, phosphorus, and mass flux</td>
<td>150, 300, 500</td>
<td>Free-Floating Particle Interceptor Traps</td>
</tr>
<tr>
<td><strong>VI. Currents</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Acoustic Doppler Current Profiler</td>
<td>0–300</td>
<td>hull mounted</td>
</tr>
<tr>
<td>Acoustic Doppler Current Profiler</td>
<td>0–1000</td>
<td>lowered</td>
</tr>
</tbody>
</table>

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annual data reports published by the HOT program (Chiswell, et al., 1990, Winn, et al., 1991). The entire suite of HOT program measurements are also available via the international internet system as described in the annual program data reports. Detailed information about methodology is available in a manual of methods prepared by JGOFS program scientists (Karl, et al., 1990).

![Map of Hawaiian Islands](image)

**Figure 1.** Location of the Hawaiian Ocean Time-series permanent station and the location of the near-shore equipment test site at Kahe Point

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**SELECTED RESULTS FROM THE HAWAII OCEAN TIME-SERIES PROGRAM**

**Hydrography**

Figures 2 and 3 show contour plots of representative results of hydrographic measurements made over a three-year period at Station ALOHA. Figure 2 shows a contour plot of potential temperature to 1000 decibars. A contour plot of nitrate plus nitrite to 1000 decibars is shown in Figure 3. As one would expect, surface water temperature has varied seasonally from approximately 23 to 26 degrees Celsius at the time-series permanent station. The isotherms below the mixed layer have remained relatively constant with only minor variations. Nitrate plus nitrite has remained low in the surface waters, and there is no evidence of substantial injection of nutrients into the upper ocean as a result of seasonal mixing.
Figure 2. Contour plot of potential temperature (°C) to 1000 decibar versus time at Station ALOHA

Figure 3. Contour plot of nitrate plus nitrite to 1000 decibar versus time at Station ALOHA
Primary Production and Particle Flux

The results of our time-series measurements of integrated primary production and carbon flux at 150 m are shown in Figure 4. These measurements have been made with carbon-14 using "clean" techniques (Fitzwater, et al., 1982) and 12 hour on-deck and in situ incubations (Karl, et al., 1990; Winn, et al., 1991). These data comprise the longest time-series of primary production and particle flux in the North Pacific Subtropical Gyre. The rate of primary production measured over a three year period ranges from a high of approximately 1100 and low of approximately 250 mg C m⁻²d⁻¹. The highest rate of production was measured on a cruise during August of 1989. A surface accumulation of Trichodesmium sp. was observed near Station ALOHA on this cruise (Karl, et al. 1992), and may have been responsible for this anomalously high value. Excluding this value, the rate of primary production measured at the HOT permanent station varies by approximately a factor of 3. The temporal variability observed in the rate of primary production at Station ALOHA appears to be stochastic with little evidence of a seasonal cycle.

The mean rate of primary production is approximately 450 mg C m⁻²d⁻¹. This rate of production is higher than historical data would suggest (Ryther, 1969), but is remarkably consistent with more recent measurements in the North Pacific Subtropical Gyre using newer "clean" C-14 primary production techniques (Martin, et al., 1987).

The rate of carbon flux at 150 m shows substantial annual variability and ranges over a factor of approximately 3 (Figure 4b). In both 1988 and 1989 there is a peak in carbon flux at 150 m in the spring and a second peak in summer. Although a peak in carbon flux in both 1989 and 1990 is observed in both the spring and the summer, only a single peak is observed in early summer in 1991.

In general, the rate of particle flux in marine environments is believed to be controlled primarily by the rate of primary production in the overlying euphotic zone. Although temporal variability in the flux of particulate material has been measured in several marine environments, most of the direct evidence for a link between primary production and particle flux at depth comes from observations in the Sargasso Sea at or near the Bermuda time-series site. Temporal variability in particle flux at 3200 m in the Sargasso Sea has been shown to vary in a pattern consistent with expected changes in surface water primary production (Deuser and Ross, 1980; Deuser, et al., 1981; Deuser, 1986). More recently, Deuser, et al. (1990) have shown a direct relationship between carbon flux at this depth and satellite derived estimates of surface ocean pigment concentration near the Bermuda time-series site. In addition, Asper, et al. (1992) have recently demonstrated a relationship between primary production measured with carbon-14 incubations and the flux of total particulate material throughout much of the water column at this same location. These results are not surprising, since one would expect the flux of particulate material in the central ocean basins away from the direct influence on terrigenous runoff to be dependent upon the production of biogenic particulate material in the euphotic zone.

It is surprising that the first three years of data collected by the HOT program show little evidence of a relationship between carbon flux at 150 m and euphotic zone primary production. Whereas a clear oscillation in the rate of particle flux is easily resolved by our monthly sampling, a similar trend in the rate of primary production is not evident (Figure 3). Although the poor correlation between these measurements has not yet been explained, several hypotheses can be advanced to account for this observation. These explanations fall into two broad categories. First, it is possible that the
Figure 4. Primary production and particle flux measured over a three-year period at Station ALOHA. Panel A shows primary productivity integrated over the upper 200 m of the water column from eight incubation depths. Panel B shows carbon flux measured at 150 m over a 72-hour period with free-floating sediment traps.
expected relationship between particle flux and primary production exists at Station ALOHA, but our primary production measurements are unable to resolve the annual cycle in primary production that the flux data suggest exists in this process. Second, it is also possible that a tight relationship between primary production and particle flux does not exist at this location, and that the some other process or combination of processes are controlling the measured rate of particle flux.

A correlation between particle flux and primary production may exist at Station ALOHA but, especially in the central Pacific Ocean where temporal variability in the physical and biological conditions are known to be small, our C-14 measurements may not be able to resolve the annual signal in primary production. Support for this scenario can be found in an experiment conducted in March of 1990. On this cruise, primary production was measured in situ on three consecutive days (Figure 3b). The production rates measured over this three day period varied by more than 100 mg C m\(^{-2}\) day\(^{-1}\). Excluding the one high value measured in August of 1989, this is equivalent to approximately 25% of the range of values measured over the duration of the program. The high degree of variability observed over this 72 hour period could be due to a highly variable rate of production which varies widely from day-to-day. The average rate experienced over the 72 hour trap deployment may therefore not be representative of that measured on a single day. Imprecision in the measurement of primary production may also play a role in this. Since we are calculating the integrated rate of production from a total of only eight incubation depths in the upper 200 m, and since primary production rates measured in incubation bottle maintained at fixed depths are known to be aliased by internal waves, it is possible that a large portion of the variability observed over this 3-day period is also due to "noise" in our primary production measurements.

As an alternative, it is also possible that factors other than primary production may greatly influence particle flux at the time-series site. Variability in the number of grazers and/or vertical migrants, for example, could be responsible for changes in the rate of particle flux independent of a change in the rate of primary production. In addition, atmospheric inputs of dust (Ditulio and Laws, 1991) could also influence the rate of particle flux on time and space scales which are not reflected in rates of primary production.

Whatever the underlying cause in the annual variation in the rate of particle flux at Station ALOHA, it is interesting that the sediment trap particle flux measurements are capable of resolving an annual cycle in an environment that is known for its temporal stability. This observation is especially noteworthy in light of the recent claims that particle traps do not provide accurate measurements of particle flux (Buesseler, 1991). Contrary to Buesseler’s finding the results of the time-series program indicate that particle traps do, in fact, provide a excellent record of annual variability in an environment known for its low seasonal variability. Although we cannot yet definitively identify the mechanism which produces the observed annual variability in particle flux it is almost impossible that the regular pattern of variability observed over this three year period is due to random errors in the measured rate of particle flux, and therefore must be indicative of an annual cycle in the physical, chemical or biological conditions in the central Pacific Ocean.

**Dissolved Inorganic Carbon**

The first time-series measurements of dissolved inorganic carbon and titration alkalinity obtained in the surface waters of the central Pacific Ocean are shown in Figure 5. Specific alkalinity averages approximately 2315 ueq/kg and is very consistent with
Figure 5. Titration alkalinity and dissolved inorganic carbon measured over a three-year period at station ALOHA. Panel A shows titration alkalinity normalized to 35 ppt salinity. Error bars show the standard deviation of replicate measurements in the upper 50 m. Panel B shows dissolved inorganic carbon normalized to 35 ppt salinity. Error bars represent the standard deviation of replicates in the upper 50 m.
other measurements made in the North Pacific Subtropical Gyre (e.g., GEOSECS Program). Although some real variability in the specific alkalinity is evident, variability in titration alkalinity normalized to 35 parts per thousand salinity appears to be stochastic and shows no evidence of a regular annual cycle. In contrast, dissolved inorganic carbon (DIC) averaged over the upper 50 decibars and normalized to 35 parts per thousand salinity, displays a very regular annual cycle with a maximum in surface water in winter and a minimum in summer. Because the partial pressure of carbon dioxide changes by approximately 4% per degree Celsius, this pattern is consistent with a flux of carbon dioxide across the air-sea interface driven by changes in surface water temperature. In winter, when surface water temperature falls, pCO2 also decreases and the net flux of carbon dioxide is from the atmosphere to the ocean. In summer, when surface water temperature rises, the net flux of carbon dioxide is from the ocean to the atmosphere. Using the time-series dataset and assuming that the mixed layer averages 60 m, the seasonal exchange of carbon dioxide between the ocean and the atmosphere at Station ALOHA is approximately 1 mole of carbon per m-2. Therefore approximately 1 mole of carbon is transported across the air-sea interface from the atmosphere to the ocean during fall and approximately this same quantity of carbon dioxide is transported in the opposite direction during spring. This flux is approximately 20 times the projected rate of anthropogenic carbon dioxide accumulation rate (Tans, et al., 1990, and references therein).

CONCLUSIONS

In contrast to the classical view of the central Pacific as a extremely stable environment, the time-series dataset has shown that considerable variability does exist in this region. Particle flux measurements in the upper ocean show an annual cycle in the rate of carbon flux at the time-series site. In both 1989 and 1990, two peaks in particle flux, one in spring and one in summer are observed. In 1991, only a single statistically significant peak in mid-summer is apparent. A similar pattern in primary production is not obvious in any of the three years of time-series observations. Although the cause of the annual and interannual variability in carbon flux has not yet been identified, one interpretation of these data is that carbon flux, measured by free-floating sediment traps, provides a more accurate estimate of biological activity in the overlying euphotic zone than primary production measured with carbon-14.

The time-series observations of upper ocean dissolved inorganic carbon and alkalinity indicate a substantial flux of carbon dioxide across the air-sea interface at Station ALOHA. These data suggest that approximately one mole of carbon dioxide moves in and out of the surface ocean each year. This flux appears to be driven by the annual oscillation in upper ocean temperature. The magnitude of this annual exchange is approximately 15 to 20 times the projected rate of carbon dioxide increase in the surface ocean due to the rise in the increase in atmospheric carbon dioxide, and indicates that several years of regular measurements will be needed to accurately quantify the upper ocean anthropogenic carbon dioxide accumulation rate.

The data collected by the HOT program, comprise the longest and most extensive set of time-series measurements available for the central Pacific Ocean. These data are the first and only time-series available for a variety of parameters in the North Pacific Subtropical Gyre. These time-series data have already provided a unique view of the extent of variability in this ocean basin, and have begun to improve our understanding of the dynamics which take place there. The understanding which is emerging from the time-series dataset will be useful for interpreting the results of the spatial scale studies planned for the central Pacific, and for interpreting the historical database within the context of future environmental and climate change.
REFERENCES


SELF-GENERATION OF CONTROLLER OF UNDERWATER VEHICLE FOR CONSTANT ALTITUDE OVER COMPLICATED TOPOGRAPHY

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ABSTRACT

A guidance system to keep an autonomous underwater vehicle at a sufficiently low altitude is constructed with the "Self-Organizing Neural-net-Controller System," which has been developed at the Institute of Industrial Science of the University of Tokyo. This system includes an adaptive controller made of an artificial neural network. The forward model network which estimates the dynamics of a vehicle and topography around it is also made of an artificial neural network. In order to construct a precise forward model network, its structure is modularized and the difference type network is introduced. The system is demonstrated in computer simulations by generating a controller which operates the PTEROA150 vehicle over a complicated topography of the seabed.

INTRODUCTION

When an autonomous underwater vehicle (AUV) swims to get scientific data in the vicinity of the seabed, a guidance system to keep a sufficiently low altitude from the seabed is necessary to accomplish the mission. This kind of guidance system may be made based on if-then algorithms. But it seems considerably difficult to cover all the cases that might happen, because the underwater environment is substantially hostile. Therefore, the guidance system for AUVs is required to have high autonomy including adaptability. Here, the "Self-organizing neural-net-controller system" (SONCS) developed by Fujii (1990, 1991) and Ura (1989) is introduced to generate an adaptive controller for constant altitude swimming on the basis of range data obtained from echo sounders.

SONCS

Figure 1 shows the overall structure of the SONCS. The major parts of the SONCS are a controller network (CN) and a forward model network (FWDN). Each network is a connectionist model. The CN generates a control signal from the values related to the control, such as state variables and environmental data. The FWDN generates the values which are used to evaluate the result of the control, and transmits back-propagation signals for adjustment of the synaptic weights of the CN.

The adjustment of the CN is carried out so as to reduce the evaluation function which is calculated with the outputs of the FWDN. It should be noted that the synaptic weights of the FWDN should not be changed through this modification because the FWDN represents the real world.
The procedure of the SONCS is divided into 5 stages.

1. The CN is initialized using the back-propagation method on the basis of navigation data. These data are obtained controlled by an appropriate controller, which is called a "rudimentary controller" because it is not necessary to be highly tuned up. After initialization, the CN controls the vehicle as the same way as the rudimentary controller did.

2. The FWDFN is constructed on the basis of the previous navigation data. Then, the FWDFN makes correspondence between the control signals of this time step and the values to be evaluated of the next time step.

3. The vehicle is navigated by the CN for a certain range.

4. The CN is modified so as to reduce the evaluation function on the basis of the navigation data obtained in stage (3).

5. Stages (3) and (4) are repeated till the value of the evaluation function becomes sufficiently small. This repetition corresponds to training of swimming. If the FWDFN doesn't make accurate outputs comparing to the corresponding value of the actual vehicle, it should be modified independently on the basis of the vehicle's data through training.

VEHICLE

The equations of motion used in the following simulation (Ura and Otsubo, 1988) is derived from the PTEROA150 in Figure 2 which was developed at the Institute of Industrial Science of the University of Tokyo. The PTEROA150 is 150 cm in length and 220 kg in dry weight. It is equipped with 4 channels of active echo sounders to detect the
Figure 2. PTEROA150

Figure 3. Configuration of robot and direction of echo sounders
topography of the seabed. Logitudinal motion is controlled only by changing the trimming angle of a pair of elevators. The SONCS is adopted to the longitudinal guidance system for the PTEROA150 over a triangular mountains.

Sounders are arranged in the longitudinal plane as illustrated in Figure 3. They are directed from directly downward to straight ahead at angle of 0, 30, 60, and 90 degrees. \(k_0 \sim k_3\) are distances to the bed which are measured by the corresponding sonic beam. \(alt\) represents an altitude defined by the distance from the vehicle to the line that is determined by the reflecting points of sonic beams for \(k_0\) and \(k_1\). The inertial coordinate system is denoted by \(Xe\) and \(Ze\). \(\theta\) is the pitching angle and \(\delta_e\) is the elevator trimming angle which is a controllable variable. In the following simulations the thrusting force is fixed at 50 N so that the average speed of the vehicle is approximately 5 knots.

The objective of the control is to swim keeping a constant altitude \(alt_0\) from a seabed which consists of series of triangles as shown in Figure 3.

**ENHANCEMENT OF THE FORWARD MODEL**

Considering the scheme of the back-propagation method for modification of the CN, the differentiations of the outputs of the FWDN by the control signals should be also accurate in order to adjust the CN quickly and appropriately. For the guidance system, however, it is not easy to construct a simple FWDN with sufficient precision in both the outputs and their differentiations, because the I/O relation is extremely complex to represent. Here, two methods are introduced to realize higher precision of the FWDN.

1) Modularization of structure of the FWDN

For constant altitude swimming, the evaluation function \(E\) is defined as:

\[
\Sigma E = (alt - alt_0)^2
\]  

(1)

where \(alt_0\) is the target altitude. So that the output of the FWDN is altitude, and its inputs are the values that are related to the calculation of the altitude of the next time step. Thus, let \(S\), \(R\), \(u\) and \(O\) be a state vector, a range vector, a control vector and an output vector, respectively. Here,

\[
\begin{align*}
S(t) &= \begin{pmatrix} \Delta Xe(t) \\ \Delta Ze(t) \\ \Delta \theta(t) \end{pmatrix}, \\
R(t) &= \begin{pmatrix} k_0(t) \\ k_1(t) \\ k_2(t) \\ k_3(t) \end{pmatrix}, \\
u(t) &= (\delta_e(t)), \\
O(t) &= (alt(t)).
\end{align*}
\]  

(2)

The mapping \(f\) by the FWDN is given by:

\[
O(t + \Delta t) = f(S(t), S(t - \Delta t), S(t - 2\Delta t), ..., \\
R(t), R(t - \Delta t), R(t - 2\Delta t), ..., \\
u(t), u(t - \Delta t), u(t - 2\Delta t), ...)
= f(S^*(t), R^*(t), u^*(t)),
\]  

(3)
where $t$ represents the present time step and $\Delta t$ is interval time of control. An assembly of present and past vectors is denoted by $\ast$.

Equation (3) can be divided into three mappings as:

$$S(t + \Delta t) = f_1(S \ast (t), u \ast (t)), \quad (4)$$

$$R(t + \Delta t) = f_2(S(t + \Delta t), R \ast (t)), \quad (5)$$

$$O(t + \Delta t) = f_3(R(t + \Delta t)). \quad (6)$$

Here, $f_1$ represents the dynamics of the vehicle and provides the state vector at the next time step $t + \Delta t$. The range data at $t + \Delta t$ are estimated by $f_2$. Then, the altitude at $t + \Delta t$ is calculated on $R(t + \Delta t)$. In the same way, the FWDN can be divided into three sub-networks which correspond to Eq. (4) to (6), respectively. Since each network represents a simple mapping, it is expected that high precision is accomplished using a general learning scheme. As a result the FWDN yields high accuracy.

In order to involve past data of input vectors, the sub-networks for $f_1$ and $f_2$ include recurrent loops, where the signals from the hidden layer are propagated to the input layer one time step delayed. The past values of input vectors are, therefore, not necessary to be treated explicitly in inputs. To represent the inclusion of the recurrent loops, Eq. (4) and (5) are written as:

$$S(t + \Delta t) = f_1 \ast (S(t), u(t)), \quad (7)$$

$$R(t + \Delta t) = f_2 \ast (S(t + \Delta t), R(t)). \quad (8)$$

2) Taking differences between I/O data

Equation (7) can be expressed as:

$$S(t + \Delta t) = S(t) + \Delta S(t)$$
$$= S(t) + \Delta f_1 \ast (S(t), u(t)). \quad (9)$$

When each component of $\Delta S(t)$ is small comparing to that of $S(t)$, it is not easy to construct an accurate network for $f_1 \ast$. Because learning by the back-propagation method makes the outputs of a network imitate the teaching data. Since $\Delta S$ in Eq. (9) can be calculated as precise as $S$ in Eq. (7), the network should be constructed to express $\Delta f_1 \ast$. Hereafter the network with this structure is called the "difference type network." Eq. (8) could be also expressed by a difference type network.

The overall network for the constant altitude swimming for the PTEROA150 is illustrated in Figure 4. A small circle represents a distributor of which output value is same as its input value. A large circle represents a neuron. At the neuron marked off by black, the corresponding input value is added to the output because the sub-network is a difference type. The leftmost network is the CN, inputs of which are the state vector, the range data and the differences of range data. The output of the CN is the elevator trimming angle. It can be seen that the FWDN consists of three sub-networks, which represent $\Delta f_1 \ast$, $\Delta f_2 \ast$, and $f_3$, and are called the dynamics, the geometric and the altitude networks, respectively.
INITIALIZATION

The data for the initialization of the CN and the construction of the FW DN are shown in Figure 5. The topography of the seabed consists of a series of triangles 50 m in span and 10 m in height. Since the FW DN is valid inside domain of the teaching data, it is desirable that these data are widely scattered. Therefore the data in Figure 5 contain the frequency from 0.01 Hz to 1.0 Hz in elevator trimming angle. The total sample points are approximately 1,000 in sampling rate of 10 Hz. When range data exceed 200 m, they are regarded as 200 m. The target altitude is 10 m in the following simulations.

Figures 6, 7 and 8 show the outputs of the dynamics, the geometric and the altitude networks, respectively. Here, the outputs of the network denoted by broken lines show good fitting to those of teaching data denoted by solid lines. Despite the range data are not continuous, the learning of the network progresses very well as shown in Figure 7. The altitude network deals with only static relation between the range data and the altitude so that its learning converges quickly.

TRAINING

Figure 9 shows a training process applied to the seabed which is the same topography used in the initialization. The vehicle collides onto the seabed at 45 m in the horizontal range without training. Once trained the vehicle gets the ability to swim avoiding collision. It can be seen that a guidance system to lead the vehicle approximately at the target altitude is constructed after 2 times of training. Thus, the vehicle succeeds in pursuing its mission within ± 6 m discrepancy.

Figure 10 is an example applied to the seabed which consists of triangles 100 m in span and 10 m in height. The top figure shows the trajectory controlled by the resulted controller in Figure 9. Although the maximum discrepancy in trajectory is 6 m at the first trial, it decreases to 3.5 m after 5 times of training.

Figure 11 is another example applied to the seabed of high triangles. Similarly the top figure is the trajectory controlled by the resulted controller in Figure 9. At first, the vehicle comes close to upward slopes. After 9 times of training, the vehicle gets the ability to swim in a good manner over long and steep slopes. In this case, the maximum discrepancy is reduced from 6.5 m to 4.5 m.

In cases of Figures 10 and 11, the geometric network is not exact because the FW DN is generated in the initialization stage and used disregarding difference of topography. But it is interesting that the results are fairly satisfactory. Speed of acquiring the ability of swimming which depends on the accuracy of the FW DN becomes higher than the example in Ura and Suto (1991).

DISCUSSION

The vehicle swims up and down in a short period in every case. This vibration does not decrease mainly because the evaluation function does not include terms related to the pitching angle.

The trajectories in general seem to be lower before the top and higher before the bottom of trough. This tendency mainly depends on the definition of the altitude in Figure 3.
Figure 5. Data for network initialization

Figure 6. Outputs of the dynamics network after learning
Figure 7. Outputs of the geometric network after learning

Figure 8. Output of the altitude network after learning
Figure 9. Training process over triangular ridges 50 m in span and 10 m in height.

Figure 10. Training process over high triangular ridges 100 m in span and 10 m in height.
Although no system for collision avoidance is implemented explicitly, the resulted controller operates the vehicle avoiding collision, because slopes are moderate. However, when there is an extremely steep wall in front of the vehicle, pull-up manoeuvre should be done far before the wall. Since the evaluation function Eq. (1) made of $k_0$ and $k_1$, cannot distinguish this case, it is necessary that the evaluation function includes terms related to $k_2$ and $k_3$.

**CONCLUSIONS**

It is demonstrated that the Self-Organizing Neural-net-Controller System (SONCS) is applicable to the guidance system for an autonomous underwater vehicle, which is able to take into account of not only the dynamics of the vehicle but also the environmental information. The generation of the controller is accomplished only by giving an evaluation function and repeating training.

Modularization of the FWDN and introduction of the difference type network proposed here to construct a precise FWDN yield high speed acquirement of the ability of swimming.
REFERENCES


