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Mean Zonal Currents Below 1500 m Near the Equator, 159°W

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Recent current measurements have shown that the subthermocline equatorial circulation in the central Pacific is considerably more complicated than had been thought. Below 1500 m depth on 159°W there are three currents south of the equator that appear to be permanent. Eastward currents are found at 1700 m and 3100 m, and a westward current is found at 4000 m. The 1700-m current is at the southern edge of the 3°S to 3°N range of current measurements, so its meridional extent and its transport are unknown. The 3100-m and 4000-m currents have 16-month mean transports of about 6–7 $\times 10^6$ m^3 s^{-1}. All three of these currents may be important components of the general circulation.

INTRODUCTION

Present knowledge of ocean circulation below the thermocline is sketchy; it is based on deep hydrographic sections and a few direct measurements of currents [*Warren*, 1981]. These measurements are usually interpreted in the framework of the *Stommel and Arons* [1960] model. Stommel and Arons constructed an elegant schematic picture of the thermohaline circulation based on linear, inviscid dynamics in the ocean interior, uniform upwelling at the top of the thermocline to balance sinking at the localized polar regions of dense water formation, and western boundary currents as required to close the circulation. It is remarkable that this model has endured for nearly 30 years and that wherever deep western boundary currents predicted by the model have been sought, they have been found [*Warren*, 1981].

Recently, *Kawase* [1987] has shown that the Stommel and Arons model is the low-dissipation limit of a more general model. While the Stommel and Arons model predicts no flow on the equator except in the western boundary current, the Kawase model predicts that a western boundary current approaching the equator can split, with some of the transport continuing across the equator and some forming a zonal jet on the equator. The relevance of this model to the deep currents that have been observed on and near the equator is far from clear, but it at least shows one way in which deep equatorial currents might be integral parts of the global thermohaline circulation.

Firing [1987] presented a 16-month time series of full-depth absolute current profiles from 3°S to 3°N on 159°W (Figures 1 and 2). The subthermocline currents above 2000 m on the equator (the deep jets) and the main extraequatorial subthermocline currents above 1500 m (the westward south equatorial intermediate current and the eastward north and south intermediate countercurrents) were discussed by *Firing* [1987], but little attention was given to the mean currents below 1500 m south of the equator. Here we will show that the currents below 1500 m are steady, carry substantial transports, and may well be important parts of the general circulation.

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OBSERVATIONS

The primary source of data is a set of current profiles obtained with a Pegasus acoustically tracked dropsonde [*Spain et al.*, 1981] during the Line Islands Profiling Project, a component of the Pacific Equatorial Ocean Dynamics (PEQUOD) project. Measurements were made every half degree of latitude between 3°N and 3°S on both the southbound and the northbound leg of each of 21 cruises, starting in March 1982 and ending in June 1983. Details of the data set are given by *Firing* [1987].

Three currents will be considered here: an eastward current with its maximum speed (in the mean) of 4.5 cm s^{-1} at 3°S, 1720 m; a second eastward current centered at 1.5°S, 3100 m (3.9 cm s^{-1}); and a westward current with its core at 2.5°S, 4000 m (4.3 cm s^{-1}). For convenience these will be designated according to their approximate core depths as the 1700-m, 3100-m, and 4000-m currents.

The standard errors of the mean currents were calculated taking into account the autocorrelations of the time series (see appendix). In the core of the 1700-m current the estimated standard error varied from 0.8 to 1.3 cm s^{-1}, depending on position and vertical averaging interval (the highest value occurring with no vertical averaging). At 1700 m, for example, the mean eastward current at 3°S was 4.5 cm s^{-1} with a standard error of 1 cm s^{-1}. At 2.5°S, the mean and standard error were 3.2 cm s^{-1} and 1.3 cm s^{-1}, respectively. Both of these means are significant beyond the 98% confidence level. Mean zonal components in the 3100-m and 4000-m currents are also significant with a high degree of confidence. The zonal velocity averaged from 2800 to 3280 m ranged from 3.1 to 4.7 times its standard error at latitudes from 0.5°S to 2.0°S. Averaging from 3800 to 4280 m, the means were 2.6 to 4.4 times the standard errors from 2.0°S to 3.0°S. All of these mean-to-standard error ratios imply confidence levels above 99%.

To test for the presence of a trend, or a slow variation connected with the 1982–1983 El Niño, the data set was divided into three nearly equal time segments and the currents were averaged for each segment (Figure 3). The upper ocean currents varied greatly from segment to segment; the middle segment (August through December 1982) includes the period when the undercurrent was absent and there was a strong eastward jet at the surface [*Firing et al.*, 1983]. Below the thermocline, all major currents, including the 1700-m, 3100-m, and 4000-m currents, were evident during each

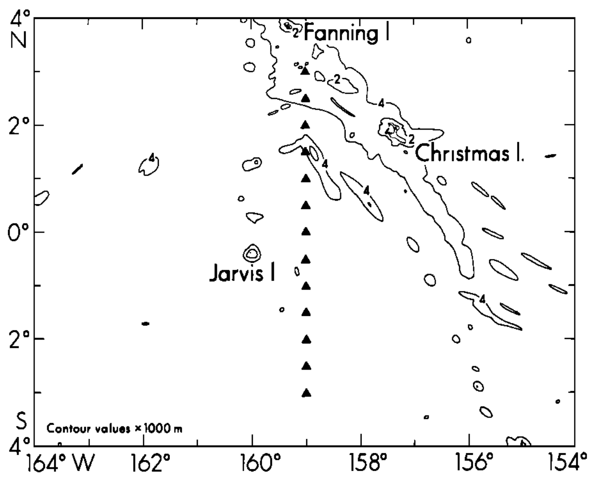

Fig. 1. Line Islands Profiling Project station locations. The Christmas Island Ridge blocks deep zonal flow north of 1°S.

segment. Mean core speeds always exceeded 3.0 cm s^{-1}. We may conclude that each of the currents under discussion was a genuine feature of the mean circulation during the 16-month period of observation.

Although the currents were highly significant in the 16-month mean, they fluctuated in strength and position, perhaps because of superimposed equatorially trapped waves [e.g., *Moore and Philander*, 1977]. The eastward 1700-m current was briefly interrupted in June 1982 and in January 1983 (Figure 4). It appears that its core was always at or south of 3°S, so we have no way of knowing its true position or strength. The northern boundary of the current varied from 2°S to near 3°S. Eastward flow in the 3100-m current was interrupted in May and November 1982 and in June 1983 (Figure 5). At other times the current at this level was eastward from about 2.5°S to 1°N and sometimes all the way from 3°S to 2.5°N. At 4000 m the westward current was continuous in most of the latitude range from 2°S to 3°S except during August 1982 (Figure 6). Since the section extends only to 3°S, we do not know the maximum southern boundary of the 4000-m current. Westward flow was sometimes present north of the equator as far as 1°N, the northern limit of the section at this depth, but the time-mean current at the equator was 1.5 cm s^{-1} to the east. Therefore the 4000-m current is a southern hemisphere feature at 159°W. This is not surprising, since the Christmas Island Ridge blocks deep zonal flow north of the equator (Figure 1).

The 16-month mean transports were calculated by vertically integrating the mean zonal velocity components (Figure 2) with absolute value greater than or equal to 1 cm s^{-1} over appropriate depth and latitude ranges (Tables 1–3). The transport of each current was taken as 55 km times the sum of the transports per unit width for each half degree of latitude from 3°S to 0.5°N. Since the 1700-m current was centered at or south of 3°S, its total transport may be considerably greater than the estimate of 1.5×10^6 m^3 s^{-1} from this section. The transports of the 3100-m and 4000-m currents are nearly equal and opposite, at 7.4×10^6 m^3 s^{-1} to the east, and 6.2×10^6 m^3 s^{-1} to the west, respectively. The southern boundary of the 4000-m current was usually south of 3°S, so its total transport is also underestimated.

Unlike the zonal velocity component, the meridional component below 1500 m was not significantly different from

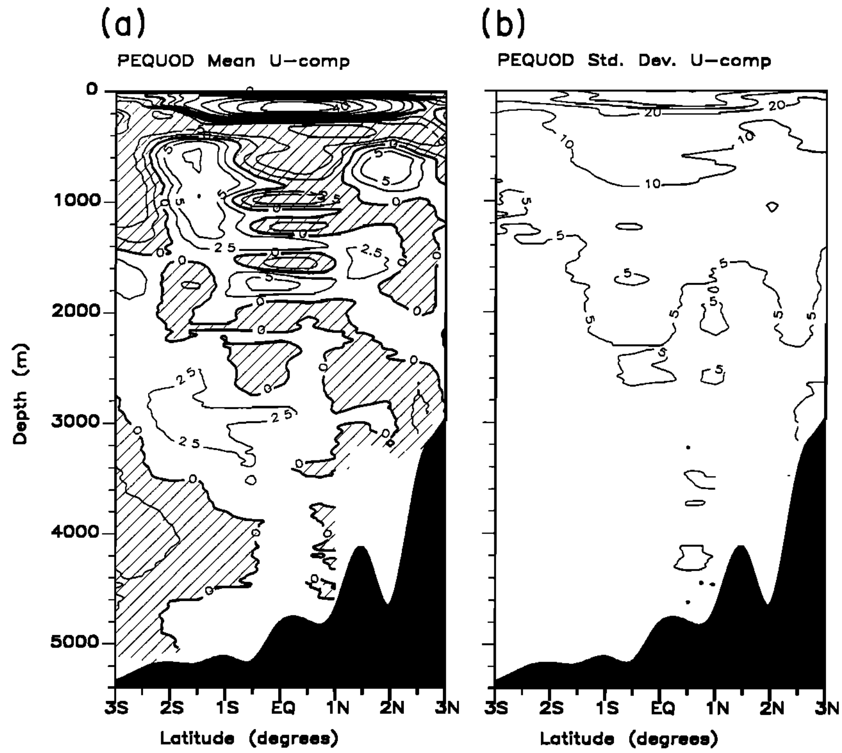

Fig. 2. Zonal component of current at 159°W, averaged from March 1982 through June 1983. (a) Mean (cm s^{-1}). (b) Standard deviation (cm s^{-1}). Regions of westward current are shaded. Note the eastward currents centered at 1720 m, 3°S, and 3100 m, 1.5°S, and the westward current at 4000 m, 2.5°S.

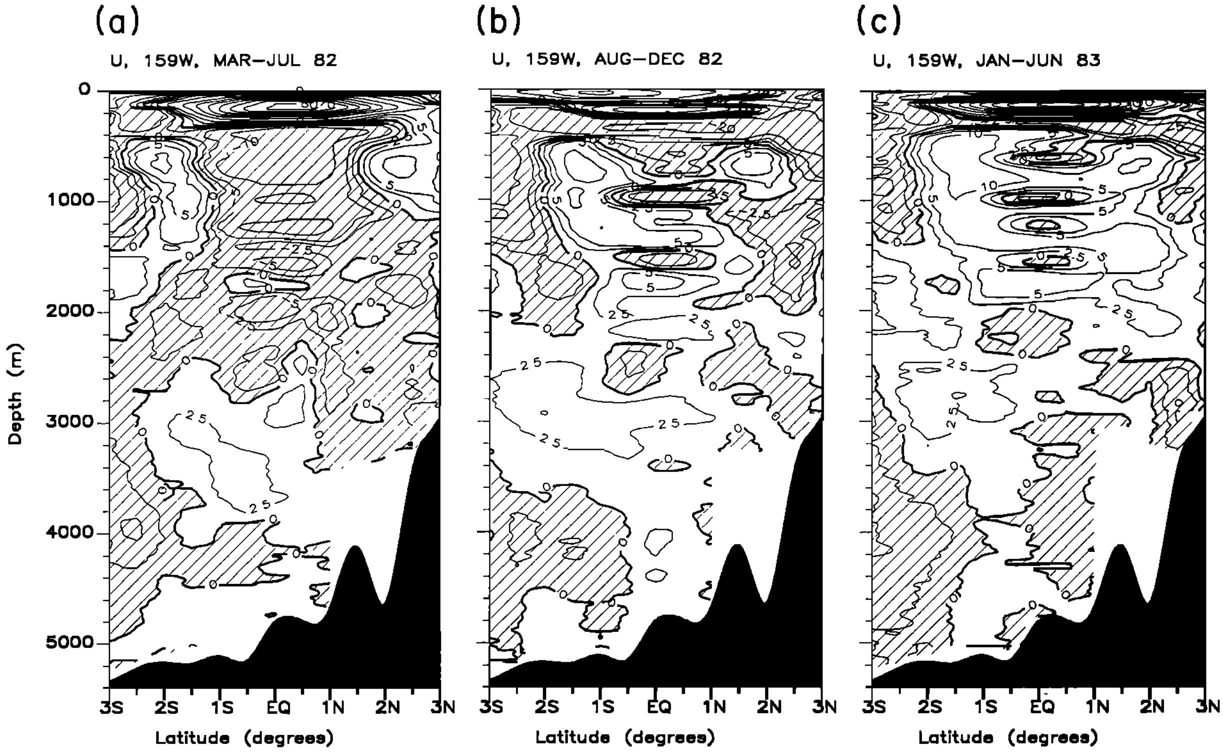

Fig. 3. Zonal component of current (cm s^{-1}) at 159°W, averaged over three periods. (a) March–July, 1982. (b) August–December, 1982. (c) January–June, 1983. The currents noted in Figure 2 can be identified in each of the three averaging intervals, in spite of variations in speed and position.

zero. Mean values rarely exceeded 1 cm s^{-1} and were less than 0.5 cm s^{-1} over most of the region. Standard errors are typically between 0.4 and 0.9 cm s^{-1}.

Discussion

The deep equatorial currents must be considered as significant parts of the general circulation if (1) they are permanent, (2) their horizontal scale is large in the sense that they span a large fraction of a basin width or are connected to western boundary currents or other basin-scale currents, and (3) their transports are the same order of magnitude as deep western boundary current transports. There is reason to believe that the first two conditions are satisfied for each of the three currents discussed here and that the third condition is satisfied for at least the 3100-m and the 4000-m currents.

Present evidence that the 1700-m and the 4000-m currents are permanent consists entirely of the measurements reported here. However, there is additional evidence that the 3100-m current is permanent. *Taft et al.* [1974] found that the current at 1°S, 150°W, 3100 m was consistently to the east during the 4.5 months when a current meter was moored there. The mean eastward component was 3.9 cm s^{-1}, identical to the core speed of the 3100-m current during PEQUOD (but larger than the 2.8 cm s^{-1} mean at 1°S). Since the *Taft et al.* [1974] mooring was 9° east of, and

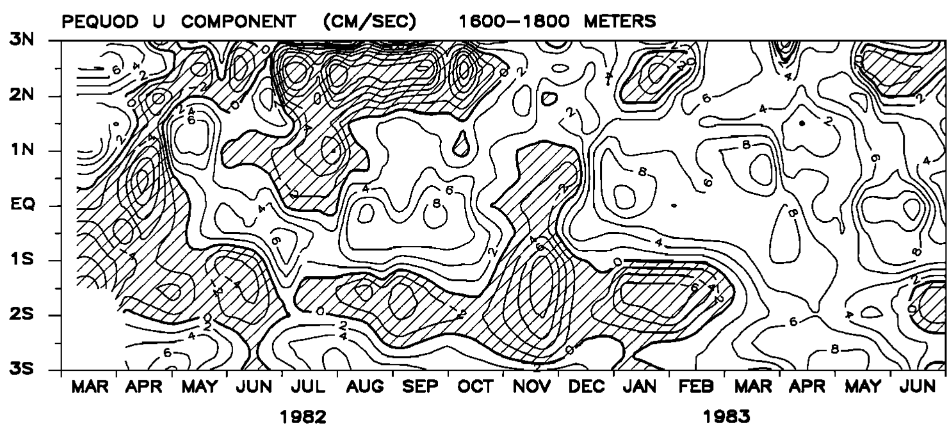

Fig. 4. Zonal component of current averaged from 1600–1800 m. The eastward current at 2.5°S and 3°S is the northern edge of the "1700-m current."

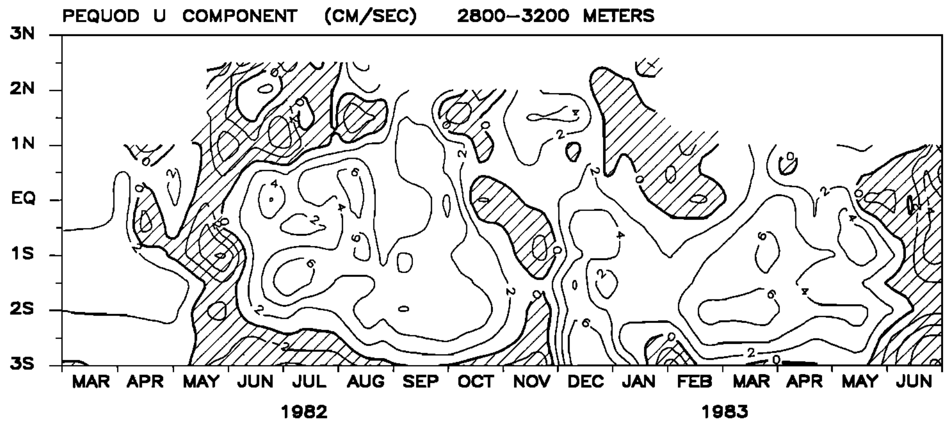

Fig. 5. Zonal component of current averaged from 2800–3200 m. The eastward current south of 1°N is the "3100-m current."

12 years earlier than, the PEQUOD section, we are encouraged to think of the 3100-m current as both permanent and of considerable zonal extent.

Beyond this indication that the 3100-m current has a zonal scale of at least 10°, the argument for large zonal scale of the deep currents rests on dynamics and scaling. Except where constrained by topography, as at the western boundary, currents near the equator tend to have large zonal scales and small vertical and meridional scales. The only obvious source of a small zonal scale in mid-ocean is topography. Since the deep currents south of the equator on 159°W are several hundred kilometers from the Christmas Island Ridge (Figure 1), it seems unlikely that they are local topographic features. Apart from isolated seamounts, there are only two barriers to continuous zonal flows within 4°of the equator all the way from the Solomon Rise near 160°E to the East Pacific Rise near 100°W. The Gilbert Islands form a porous barrier from about 3°S to 3°N near 175°E. Major sill latitudes and approximate depths [*U.S. Naval Oceanographic Office*, 1973], are 3100 m at 2.2°S, 3500 m at 1.7°S, 2700 m at 0.9°S, 3800 m at 0.2°S, 3100 m at 0.5°N, 2000 m at 1.2°N, and 3800 m at 2.4°N. North of 1°S near 157°W, the Christmas Island Ridge is a major barrier. There are sills near 3500 m from 1°S to the equator, near 3100 m at 0.5°N and 1.2°N, and near 2800 m at 2°N. Hence it is not surprising that we found no northern hemisphere conterpart at 159°W to the southern hemisphere 4000-m current.

It is interesting to compare the present observations with the current profiles made by *Eriksen* [1981] near the Gilbert Islands. Two meridional sections were occupied: 168°E, west of the island chain, and 179°E, east of the chain. Meridional resolution is coarse and there is only one profile at each station, so no strong conclusions can be drawn regarding the presence at these latitudes of the mean currents seen at 159°W; we can only note the degree of similarity among the observations. On 179°E, westward current was measured at 1.5°S and 3°S in most of the depth range below 3700 m. This is consistent with the presence of the 4000-m current. Eastward flow between 2900 and 3600 m at 1.5°S and between 3100 and 4200 m at 0.45°S may indicate the presence of the 3100-m current. Flow was also to the east between 600 and 1900 m at 3°S, consistent with the presence of the 1700-m current, but not with the westward current usually found above 1500 m at 3°S on 159°W (the south equatorial intermediate current) [*Firing*, 1987]. On 168°E there was no profile at 3°S, and all flow below 1500 m at 1.5°S was to the west; there was no indication of the 3100-m current on this section.

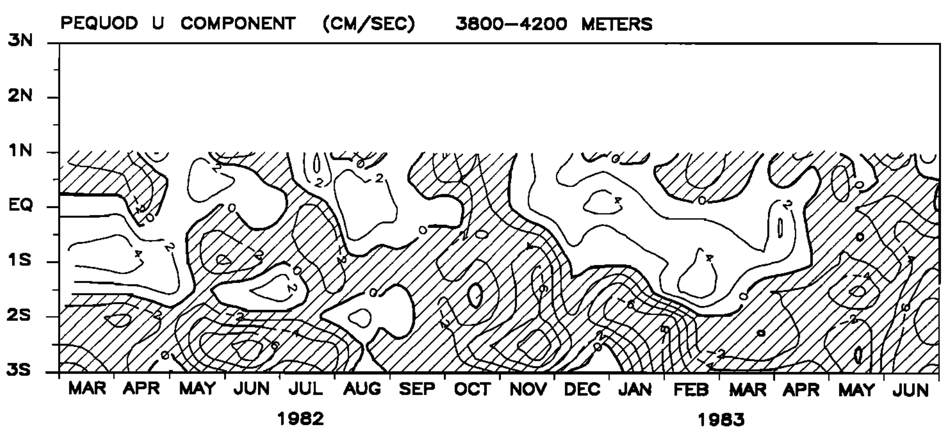

Fig. 6. Zonal component of current averaged from 3800–4200 m. The westward current south of the equator is the "4000-m current."

TABLE 1. The 1700-m Current

Latitude	Depth Range m	Transport 10^6 m^3 s^{-1}
3.0°S	1460–2020	0.85
2.5°S	1540–2100	0.69
	total	1.54

The 16-month mean (± standard deviation) zonal velocity component is 4.5 (± 3.7) cm s^{-1} (positive eastward) at 1720 m, 3°S. The core may be south of this, beyond the range of these measurements.

TABLE 3. The 4000-m Current

Latitude	Depth Range m	Transport 10^6 m^3 s^{-1}
3.0°S	2940–4940	−2.82
2.5°S	3540–4860	−1.95
2.0°S	3740–4380	−0.81
1.5°S	3620–4360	−0.46
1.0°S	4020–4260	−0.14
	total	−6.18

The 16-month mean (± standard deviation) zonal velocity component is 4.3 (±4.3) cm s^{-1} (positive eastward) in the core at 3920–4000 m, 2.5°S.

The estimated transports of the 3100-m and 4000-m currents may be compared with estimates of western boundary current transports in the Pacific. *Warren and Voorhis* [1970] used a combination of hydrography and neutrally buoyant floats to estimate the northward flow of water across 22°S along the effective western boundary of the central South Pacific, the Kermadec Ridge. The upper boundary of the flow ranged from 3100 to 4300 m. The transport was estimated as 13×10^6 m^3 s^{-1}, about half of the *Stommel and Arons* [1960] model prediction for that latitude. At the equator, the Stommel and Arons model predicts a smaller northward transport, 10×10^6 m^3 s^{-1}. Brief direct current measurements in the Samoan Passage (10°S, 170°W) *Reid and Lonsdale* [1974] showed northward currents of a few centimeters per second below about 3800 m, but there were not enough observations for a reliable estimate of transport.

Given that the 3100-m and 4000-m currents have transports of a quarter to half the size of the deep western boundary currents in the tropical Pacific, that they were present during most of a 16-month period, and that they do not appear to be small-scale topographic features, they must be viewed as likely parts of the deep general circulation. Hence their study should logically be included in the World Ocean Circulation Experiment (WOCE) [*Nowlin*, 1987]. Observations are required to determine the zonal structure, sources, and destinations of the currents observed at 159°W, and to search for similar currents in the Indian and Atlantic oceans. No set of deep currents similar to that in Figure 2 has been reported yet from a high-resolution numerical model including the equator [e.g., *Philander and Pacanowski*, 1980; *Cox and Bryan*, 1984; *Cox*, 1985]. There are several possible reasons, including inadequate vertical resolution below the thermocline, inadequate horizontal resolution, and lack of bottom topography. Numerical experiments that are carefully designed to focus on deep circulation in the tropics are badly needed.

TABLE 2. The 3100-m Current

Latitude	Depth Range m	Transport 10^6 m^3 s^{-1}
2.5°S	2400–3060	0.66
2.0°S	2480–3440	1.35
1.5°S	2320–3360	1.46
1.0°S	2500–3720	1.37
0.5°S	2800–3820	1.28
0°	2720–3060	0.41
0.5°N	2360–3040	0.86
	total	7.39

The 16-month mean (± standard deviation) zonal velocity component is 3.9 (±3.5) cm s^{-1} (positive eastward) in the core at 3080–3100 m, 1.5°S.

APPENDIX ESTIMATION OF THE STANDARD ERROR OF THE MEAN

Consider a sequentially correlated time series (x_i; $i = 1 \ldots n$) and let μ be the ensemble mean of x_i. Then the standard error, σ, of the sample mean, $\overline{x} = \sum_{i=1}^{n} x_i$, is given by

$$\begin{aligned} \sigma^2 &= \mathcal{E}\left[(\overline{x} - \mu)^2\right] \\ &= \frac{1}{n} \sum_{k=1-n}^{n-1} (1 - \frac{|k|}{n})\, \gamma_x(k) \end{aligned} \quad (1)$$

where $\gamma = \mathcal{E}[(x_i - \mu)(x_{i+k} - \mu)]$ is the autocovariance of x at lag k. This formula was presented by *Bayley and Hammersley* [1946] and *Anderson* [1971], and its analog for a continuous function of time was used by *Leith* [1973]. Equation (1) leads to an expression for the number of equivalent degrees of freedom, n^*:

$$\frac{1}{n^*} = \frac{1}{n} \sum_{k=1-n}^{n-1} (1 - \frac{|k|}{n})\, \rho_x(k) \quad (2)$$

where $\rho = \gamma_x(k)/\gamma_x(0)$ is the autocorrelation function [*Bayley and Hammersley*, 1946].

To apply these expressions, one must estimate the the autocovariance or autocorrelation function, usually from the same time series from which one is calculating the mean and standard error. Conventional estimators are

$$\tilde{\gamma}_x(k) = \frac{1}{n} \sum_{i=1}^{n-|k|} (x_i - \overline{x})(x_{i+k} - \overline{x}) \quad (3)$$

$$\gamma_x^*(k) = \frac{1}{n - |k|} \sum_{i=1}^{n-|k|} (x_i - \overline{x})(x_{i+k} - \overline{x})$$

The estimator $\tilde{\gamma}_x$ has lower mean squared error than γ_x^* [*Parzen*, 1964; *Jenkins and Watts*, 1968] and is generally preferred. However, both estimators are biased in such a way that neither is suitable for use in (1). Combining several expressions from *Anderson* [1971], we find the bias of $\tilde{\gamma}_x$:

$$\begin{aligned} \mathcal{E}\left[\gamma_x(k) - \tilde{\gamma}_x(k)\right] \approx &- \frac{|k|}{n} \gamma_x(k) \\ &- \frac{1}{n} \left(1 - \frac{|k|}{n}\right) \sum_{l=1-n}^{n-1} \left(1 - \frac{|l|}{n}\right) \gamma_x(l) \end{aligned} \quad (4)$$

The first term on the right gives the tendency of $\tilde{\gamma}_x$ to underestimate γ_x by a factor proportional to the lag, $|k|$. The second term is a negative bias at all lags. It arises from the difference between the sample mean $\bar{x}$ and the ensemble mean μ. Although the magnitude of the bias is inversely proportional to n, its cumulative effect in the calculation of σ^2 is $O(1)$ because of the summation in (1). For example, if x_t is an uncorrelated normal random variable, then σ^2 will be systematically underestimated by a factor of 1/3, independent of the number of data points. This has been verified numerically by Monte Carlo experimentation.

The tendency of the negative bias term in (4) to cause an underestimate of σ^2 can be counteracted by setting the estimated autocovariance to zero for lags beyond some threshold. I have chosen to retain only the central peak of the autocovariance function:

$$\begin{aligned} \tilde{\gamma}_x'(k) &= \tilde{\gamma}_x(k) \qquad && |k| < k_0 \\ \tilde{\gamma}_x'(k) &= 0 && \text{otherwise} \end{aligned} \tag{5}$$

where k_0 is the smallest lag for which $\tilde{\gamma}$ is negative. Using (5) in (2) ensures that $n^* \leq n$. It gives a conservative estimator in the sense that it is biased toward low n^* and large σ. For example, in a Monte Carlo experiment with 100 sequences of 50 uncorrelated normal random variables, the variance of the sample means was estimated using (1) and (5) to be 0.023, compared with the theoretical value of 0.020. In other experiments in which the time series contained both a normal random variable and a harmonic component with unit amplitude, the estimated variance of the sample mean was always greater than the actual variance so long as the period of the harmonic component was less than or equal to the series length. The estimated variance typically exceeded the actual variance by a factor of 2 or more.

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REFERENCES

Anderson, T. W., *The Statistical Analysis of Time Series*, 704 pp., John Wiley, New York, 1971.

Bayley, G. V., and J. M. Hammersley, The "effective" number of independent observations in an autocorrelated time series, *J. R. Stat. Soc., B8*, 184–197, 1946.

Cox, M. D., An eddy resolving numerical model of the ventilated thermocline, *J. Phys. Oceanogr., 15*, 1312–1324, 1985.

Cox, M. D., and K. Bryan, A numerical model of the ventilated thermocline, *J. Phys. Oceanogr., 14*, 674–687, 1984.

Eriksen, C. C., Deep currents and their interpretation as equatorial waves in the western Pacific Ocean, *J. Phys. Oceanogr., 11*, 48–70, 1981.

Firing, E., Deep zonal currents in the central equatorial Pacific, *J. Mar. Res., 45*, 791–812, 1987.

Firing, E., R. Lukas, J. Sadler, and K. Wyrtki, Equatorial undercurrent disappears during 1982–1983 El Niño, *Science, 222*, 1121–1123, 1983.

Jenkins, G. M., and D. B. Watts, *Spectral Analysis and its Applications*, 525 pp., Holden-Day, San Francisco, Calif., 1968.

Kawase, M., Establishment of deep ocean circulation driven by deep-water production, *J. Phys. Oceanogr., 17*, 2294–2317, 1987.

Leith, C. E., The standard error of time-average estimates of climatic means, *J. Appl. Meteor., 12*, 1066–1069, 1973.

Moore, D. W., and S. G. H. Philander, Modeling of the tropical oceanic circulation, in *The Sea: Ideas and Observations on Progress in the Study of the Seas, vol. 6, Marine Modeling*, edited by E. D. Goldberg, I. N. McCave, J. J. O'Brien, and J. H. Steele, pp. 319–361, Wiley-Interscience, New York, 1977.

Nowlin, W. D., Status of US planning for WOCE, *Bull. Am. Meteorol. Soc., 68*, 1559–1565, 1987.

Parzen, E., An approach to empirical time series analysis, *J. Res. Natl. Bur. Stand., Sect. D., Radio Sci., 68*, 551–565, 1964.

Philander, S. G. H. and R. C. Pacanowski, The generation of equatorial currents, *J. Geophys. Res., 85*, 1123–1136, 1980.

Reid, J. L., and P. F. Lonsdale, On the flow of water through the Samoan Passage, *J. Phys. Oceanogr., 4*, 58–73, 1974.

Spain, P. F., D. L. Dorson, and H. T. Rossby, Pegasus: A simple, acoustically tracked velocity profiler, *Deep Sea Res., Part A, 28*, 1553–1567, 1981.

Stommel, H., and A. B. Arons, On the abyssal circulation of the world ocean II, An idealized model of the circulation pattern and amplitude in oceanic basins, *Deep Sea Res., 6*, 217–233, 1960.

Taft, B. A., B. M. Hickey, C.Wunsch, and D. J. Baker, Jr., Equatorial Undercurrent and deeper flows in the central Pacific, *Deep Sea Res., 21*, 403–430, 1974.

U.S. Naval Oceanographic Office, Bathymetric Atlas of the North Pacific Ocean, *N.O. Publ. 1301-2-3*, Suitland, Md., 1973.

Warren, B. A., Deep circulation of the world ocean, in *Evolution of Physical Oceanography*, edited by B. A. Warren and C. Wunsch, pp. 6–41, MIT Press, Cambridge, Mass., 1981.

Warren, B. A., and A. D. Voorhis, Velocity measurements in the deep western boundary current of the South Pacific, *Nature, 228*, 849–850, 1970.

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