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Equatorial Currents: Observations and Theory

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6.1 Introduction

Historically, our knowledge of the circulation patterns in the tropics was derived from compilations of ship-drift data and so was restricted to a description of the surface currents. Although information of this type is crude, a picture of the spatial and temporal structure of the surface flow field was deduced over the years, and it has been little improved upon in the modern era of instrumentation. By contrast, almost all of the information about subsurface equatorial flows has been acquired recently. It is remarkable that one of the major ocean currents, the Pacific Equatorial Undercurrent, was not discovered until 1952.

For many reasons, progress in understanding equatorial circulations has been slow. The equatorial regions are vast and remote. The swift currents and high vertical shears put special demands on instrumentation. The geostrophic approximation, which is so useful at mid-latitudes, breaks down close to the equator and cannot be relied on to give accurate information about the currents. Finally, the flows seem much more time dependent than at mid-latitudes. Hence, on the basis of individual cruises, haphazardly taken in time and space, it is difficult to develop a consistent picture of the circulation patterns.

The variability of equatorial circulations has only recently been appreciated. This is partly because a great deal of the information about equatorial circulations comes from the central Pacific, where historically the mean appears to dominate the transient circulation; recent NORPAX (North Pacific Experiment) observations (Wyrski, McLain, and Patzert, 1977; Patzert, Barnett, Sessions, and Kilonsky, 1978), however, suggest that the variability can be quite large even there. In other regions such as the western or eastern Pacific or the Indian Ocean, the fluctuating components are as large as or larger than the means.

The goal of this chapter is to give a short overview of the outstanding features of the equatorial ocean circulation patterns, the dominant spatial and temporal structures of the Pacific equatorial wind field (as an example of the kinds of driving mechanisms that need to be considered), and a summary of some of the theoretical ideas that have been developed to explain the ocean circulation and its relation to the wind field. No attempt is made to be comprehensive because in recent years there have been numerous excellent reviews of equatorial phenomena and theories for them. These include articles by Knauss (1963), Tsuchiya (1970), Rot-schi (1970), Philander (1973), Gill (1975a), and Moore and Philander (1977). A collection of papers discussing various topics of equatorial oceanography is contained in the proceedings of the FINE (1978) workshop, held

at Scripps Institution of Oceanography during the summer of 1977. A comprehensive discussion of analytic techniques for studying forced baroclinic ocean motions in the equatorial regions is presented in a three-part paper by Cane and Sarachik (1976, 1977, 1979). In discussing the theories, we shall stress the important physical ideas of each model (avoiding whenever possible the use of mathematics), put them in historical perspective, and relate them to the observations. The objective here is to identify the observations for which we have physical theories, and thereby indicate where further work is needed.

The importance of knowing the detailed time and spatial structure of the wind field is emphasized throughout this chapter. The reason is that in the tropics, the characteristic response times for baroclinic oceanic processes are much shorter than they are at mid-latitude and are much closer to the time scales characterizing the wind variations. Therefore the baroclinic response to atmospheric forcing is expected to be much stronger than at mid-latitude. The implication is that in order to arrive at a satisfactory explanation of the oceanic features, the temporal and spatial structure of the atmospheric forcing must be known accurately.

6.2 Observations

6.2.1 The Ocean

The surface currents are characterized by zonal bands in which the flow is alternately eastward or westward (Knauss, 1963). The eastward flows are referred to as *countercurrents* because they flow counter to the direction of the easterly trade winds. The westward flows are referred to as *North and South Equatorial Currents*. In the Atlantic and the Pacific, the North Equatorial Countercurrent (NECC) is approximately located between 5° and 10°N with westward flow to the north of this region in the North Equatorial Current (NEC) and westward flow to the south of it in the South Equatorial Current (SEC). There is also evidence for a South Equatorial Countercurrent (SECC) in both oceans between 5 and 10°S (Reid, 1964b; Tsuchiya, 1970; Merle, 1977); these flows are not as well developed, however, as the NECC. Both the intensity and location of the various currents vary seasonally (Knauss, 1963; Merle, 1977). The SEC and NECC are strongest during July and August. In the northern winter and spring the SEC generally vanishes and the NECC is weak; in the eastern Pacific there is some evidence that during this time the NECC is discontinuous at some longitudes or is entirely absent (Tsuchiya, 1974). In the northern summer the Pacific NECC assumes its northernmost position, whereas in the northern winter the current lies closest to the equator. The data base is insufficient to show an analogous migration of the Atlantic NECC.

The structure of the surface currents in the Indian Ocean differs markedly from those in the other two oceans (*African Pilot*, 1967). In the Indian Ocean the SEC usually lies totally south of 4°S. The predominantly eastward flow in the Indian Ocean is almost totally confined between the equator and the SEC. North of the equator the flow direction varies seasonally. During the northeast monsoon it is to the west.

The different circulation pattern in the Indian Ocean is no doubt related to the nature of the wind field. In the Atlantic and the Pacific, the southeast and the northeast trades are well developed over most of the ocean throughout the year. The mean stress is generally greater than the annual or semiannual components. In the Indian Ocean south of about 10°S, the southeast trades are reasonably steady. North of 10°S, the mean winds are weak and the stress field is dominated by the strong, regular forcing of the southwest and northeast monsoons.

The earliest measurements that indicated that the currents at depth might behave differently from those at the surface were made in 1886 by Buchanan (1888; see discussion by Montgomery and Stroup, 1962). Using drogues, he found that the subsurface flow on the equator in the Atlantic at 14°W was toward the southeast, while the surface flow had a slight westward set. These unusual measurements were thought not to be representative until after the discovery of the Equatorial Undercurrent in the Pacific in 1952 (Cromwell, Montgomery, and Stroup, 1954). The Equatorial Undercurrent has been found to be a subsurface eastward flow that is about 100–200 m thick and 200–300 km wide. It is centered approximately on the equator. Its core lies just beneath the base of the mixed layer in the top of the equatorial thermocline. Undercurrents are found in all three oceans. The one in the Indian Ocean, however, appears to be present primarily during the northeast monsoon, and most observations of it have been made during the spring. A thorough discussion of early evidence for the Equatorial Undercurrent in all three oceans is given in Montgomery (1962); see also Philander (1973).

Evidence for additional subsurface countercurrents in the Pacific is discussed in detail by Tsuchiya (1975). He finds narrow subsurface countercurrents in the Pacific Ocean symmetrically located about the equator. The jets occur at a depth of 200–300 m, and are associated with the poleward limits of the thermocline (a subsurface, relatively homogeneous layer of equatorial 14°C water). In the eastern Pacific, they are located roughly at 5°N and 5°S and are distinct from the surface countercurrents as well as from the Equatorial Undercurrent itself. Further to the west, the jets move closer to the equator (Taft and Kovalala, 1979) and have been observed to merge with the Equatorial Undercurrent (Hisard, Merle, and Voituriez, 1970). The jets appear in

virtually all central and eastern Pacific Ocean hydrographic sections, and thus appear to be permanent features of the equatorial current system there. There is also evidence that similar currents exist in the Atlantic Ocean as well (Tsuchiya, 1975; Cochrane, Kelly, and Olling, 1979).

Although something is known about the temporal fluctuations of the surface currents from ship-drift observations, little systematic information exists about the time and space variations at any frequency. The recent survey article by Wunsch (1978b) discusses the observational evidence for equatorial motions with periods from a few days to several months. Fluctuations have been observed with periods of 4 to 5.5 days (Wunsch and Gill, 1976; Weisberg, Miller, Horigan, and Knauss, 1980; Weisberg, Horigan, and Colin, 1979) Harvey and Patzert (1976) present evidence that suggests waves with a period of about 25 days. Features with a similar time scale and wavelength of about 1000 km are observed in the satellite infrared images of the equatorial front by Legeckis (1977b). In the Atlantic, Düing et al. (1975) found evidence for a meandering of the Equatorial Undercurrent with a "period" of 2-3 weeks and a "wavelength" of 3200 km during GATE (GARP Atlantic Tropical Experiment).

Neumann, Beatty, and Escowitz (1975) and Katz et al. (1977) have shown that the east-west slope of the thermocline in the Atlantic varies on a seasonal time scale. This variation appears to be in phase with changes in the zonal component of the wind stress. When the trades are strongest the tilt of the thermocline is the greatest, and when the stress is at a minimum so is the slope. Meyers (1979), in a similar analysis for the Pacific, finds strong annual and interannual variations in the depth of the 14° isotherm. This isotherm was chosen because it is representative of movements of the thermocline as a whole. The east-west slope was a minimum in May and June and a maximum during October and November. The easterly wind stress is weakest in March-May; there is some phase change, however, in the annual component with longitude. Hence there appears to be a time lag between the minimum east-west slope and the minimum wind. The exact relation between the winds and the variations in the slope was difficult to deduce because the variations about the mean slope are small and because smaller spatial scales than the basin width were present in the slope of the 14° isotherm. These smaller scales appear to propagate westward. In the Indian Ocean the mean east-west stress is small. In the transition between the monsoons, however, westerly winds appear along most of the equator. Associated with these is an eastward oceanic jet and simultaneously a rise in the sea level off Sumatra and a rise of the thermocline off East Africa (Wyrtki, 1973a). The oceanic response appears to occur coincidentally with the winds.

6.2.2 The Wind Field

To date, most theoretical treatments of equatorial flows consider the applied wind stress to be some simple functions of time and latitude, whereas even the most general meteorological description of the trades indicates that there is considerable spatial and temporal structure (Riehl, 1954). In recent years, there have been several thorough analyses of the stress field over the ocean as derived from historical merchant-vessel reports. In the Pacific, this analysis was done for the whole ocean by Wyrtki and Meyers (1975) and for the eastern part by Hastenrath and Lamb (1977). The latter authors and Bunker (1976) also analyzed the winds over the tropical Atlantic. From these studies, it is possible to look in detail at the long-period temporal fluctuations and the large-scale spatial structure of the stress field. For the sake of brevity, we shall limit our discussion to the Pacific data. Some similar features are evident in the Atlantic, while the Indian Ocean wind field is dominated by the monsoon circulation.

Values of the zonal stress within 4° of the Pacific equator are shown in figure 6.1. These curves are derived from the mean monthly values of Wyrtki and

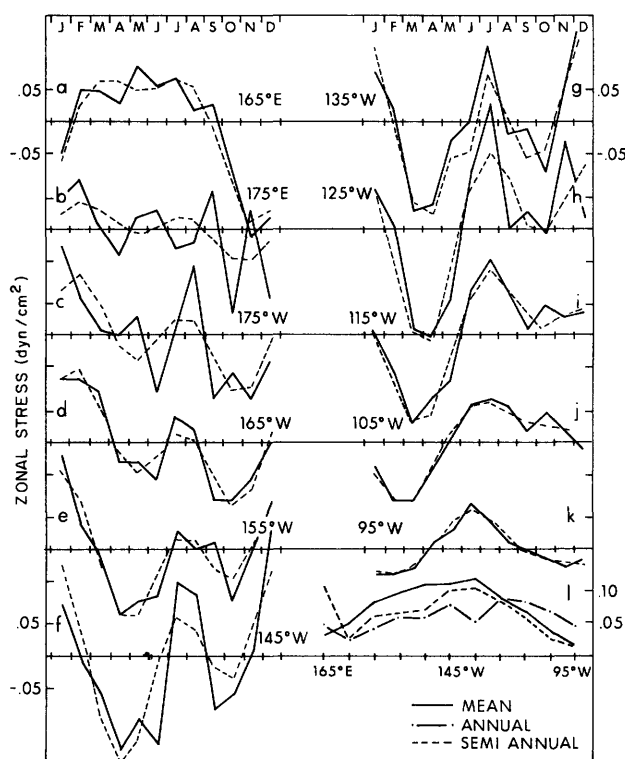


Figure 6.1 The monthly variation in the mean zonal stress between 4°N and 4°S. In figures 6.1a-6.1k, the solid curves display the average stress for blocks of 10° of longitude centered at the indicated values. The dashed curves show the proportion that is fitted by the annual and semiannual components. Figure 6.1l shows the amplitude of the mean, annual, and semiannual components as a function of longitude. The amplitude of the mean component has been reduced by a factor of five, i.e., at 145°W it is about 0.5 dyn cm⁻².

Meyers (1975), which are tabulated for areas of 2° of latitude by 10° of longitude. The monthly values of the equatorial zonal stress with the yearly mean removed are shown in figures 6.1a–6.1k. As can be seen, the annual and semiannual components account for almost all of the annual variation. The amplitudes and phases of these components are tabulated in table 6.1. The phase of the annual component decreases rapidly, although irregularly, to the east. The phase of the semiannual component also decreases to the east, but more slowly. The amplitudes of the mean, annual, and semiannual components as a function of longitude are shown in figure 6.11. The mean stress over most of the Pacific is about a factor of five larger than the annual and semiannual components. The amplitude of each component varies strongly in the zonal direction. Thus we expect that models driven by zonally uniform stress distributions may not be adequate for describing the oceanic response to the wind.

Meyers (1979) presents evidence that the longitudinal structure of the wind field must be properly accounted for. He finds significant energy at the semiannual period in the eastern Pacific Ocean even though there is little energy in the wind stress there at that period. A study of the EASTROPAC data supports his findings. Figure 6.2 shows the average rate of change of the depth of the 20° isotherm (m/60 days) between 1°N and 1°S at four different longitudes. As can be seen, there is a pronounced semiannual variation, and the changes occur almost simultaneously at each longitude. The large amplitude of the changes (~50 m/60 days) is surprising considering the small amplitude of the semiannual component of the stress at these longitudes (0.02–0.03 dyn cm⁻²). We suggest that these fluctuations are caused by baroclinic waves that have propagated eastward from a region where the semiannual component of the stress is much larger. Meyers reached the same conclusion.

So far we have considered only winds in the vicinity of the equator and have ignored the transient changes in the stress, and the curl of the stress, that are related to seasonal movements of the Intertropical Convergence Zone (ITCZ). To investigate these effects, the monthly values of the stress averaged between 100 and 120°W were examined. To emphasize the meridional structure, the weak monthly mean stress at the equator (~0.1 dyn cm⁻²) was subtracted from the monthly mean value of the stress at each latitude. Resulting monthly values are shown in figure 6.3. The annual movement of the ITCZ has little effect close to the equator, with the largest amplitude of the variation in the stress occurring at about 9°N. There the amplitude is about 0.5 dyn cm⁻². The variation south of the equator is considerably less. But it is evident that the annual variation in wind-stress curl can be very large and must be taken into account in the models.

Table 6.1 Amplitude and Phase of Annual and Semiannual Components of Zonal Stress^a

	Annual		Semiannual	
	Amplitude (dyn cm ⁻²)	Phase	Amplitude (dyn cm ⁻²)	Phase
165°E	0.11	308°	0.05	77°
175°E	0.03	263°	0.03	46°
175°W	0.04	239°	0.06	39°
165°W	0.06	224°	0.07	34°
155°W	0.06	153°	0.07	13°
145°W	0.08	122°	0.10	356°
135°W	0.05	113°	0.11	331°
125°W	0.09	72°	0.10	337°
115°W	0.08	55°	0.06	333°
105°W	0.07	34°	0.03	287°
95°W	0.04	342°	0.02	291°

a. Amplitude = $a^2 + b^2$, phase = $\tan^{-1}(b/a)$, and

$$\tau = \tau_0 + a_1 \cos \frac{2\pi}{12} t + b_1 \sin \frac{2\pi}{12} t + a_2 \cos \frac{2\pi}{6} t + b_2 \sin \frac{2\pi}{6} t.$$

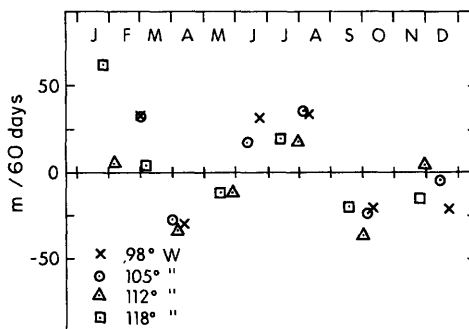


Figure 6.2 The rate at which the depth of the 20°C isotherm was observed to change during the EASTROPAC expedition in 1967 and early 1968.