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On the Obscurantist Physics of "Form Drag" in Theorizing about the Circumpolar Current*

BRUCE A. WARREN

Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

JOSEPH H. LACASCE

MIT–WHOI Joint Program in Oceanography, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

PAUL E. ROBBINS

Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

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ABSTRACT

The authors point out that, since the "form-drag" force balance commonly advanced for the Antarctic Circumpolar Current is really just a statement that northward Ekman transport in the circumpolar Drake Passage zone is compensated by deep southward geostrophic flow, the balance is actually irrelevant to the magnitude of the current itself. It is thus misleading to ascribe a role to form drag in its physics. Sverdrup dynamics seems to offer a more promising analysis of the real Circumpolar Current—as proposed long ago.

Theories of wind-driven circulation are usually based on the idea of interior geostrophic flows forced by divergence of Ekman transport due to curl of wind stress. The curious exception is the Antarctic Circumpolar Current, for which some contemporary theorizing (e.g., McWilliams et al. 1978; Johnson and Bryden 1989; Treguier and McWilliams 1990; Wolff et al. 1991; Marshall et al. 1993) asserts instead that the stress is transmitted by some hypothetical agency (e.g., "interfacial form drag") directly to the bottom, where it is opposed by pressure differences across submarine ridges like the Scotia Arc, Kerguelen Plateau, Macquarie Ridge, and Pacific–Antarctic Ridge. This description proceeds, essentially, from the time-averaged zonal momentum equation (usual notation),

$$fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} + \frac{\partial \tau_x}{\partial z}, \tag{1}$$

whose vertical integral from the ocean bottom at $z = -D(x)$ to the sea surface at $z = 0$ is

$$f\int_{-D}^{0} v\,dz = -\frac{\partial}{\partial x}\int_{-D}^{0}\frac{1}{\rho}p\,dz - \frac{1}{\rho}p(-D)\frac{\partial D}{\partial x} + \tau_x(0) - \tau_x(-D). \tag{2}$$

[Meridional Reynolds-stress contributions to (1) and (2) are demonstrably slight (Johnson and Bryden 1989).] If the bottom stress is negligibly small, the circumpolar integral C of (2) along a latitude parallel around the Southern Ocean (Drake Passage zone) is

$$\int_C \tau_x(0)\,dx = \int_C \frac{1}{\rho}p(-D)\frac{\partial D}{\partial x}\,dx \tag{3}$$

since the integral of v must vanish, owing to mass conservation. The right-hand term has been called the "mountain drag", "bottom form drag", or "topographic form stress"; it is nonzero when there are pressure differences across ridges (varying D), and it has been invoked to balance directly the imposed wind stress in the left-hand term.

The mathematics is certainly pertinent, but the language obscures the more conventional physics that (3) represents. If $z = -H$ is the sill depth of the highest ridge in the open circumpolar zone, then the vertical, circumpolar integral of (1) for $-H \leq z \leq 0$ is

$$f\int_C dx\left(\int_{-H}^{0} v\,dz\right) = \int_C \tau_x(0)\,dx \tag{4}$$

* Woods Hole Oceanographic Institution Contribution Number 8974.

Corresponding author address: Dr. Bruce A. Warren, Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, MA 02543.
E-mail: bwarren@cliff.whoi.edu

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on the assumption, contrary to that of Johnson and Bryden (1989), that $\tau_x(z)$ falls to zero at $z > -H$; and the integral for $-D < z < -H$ is

$$f \int_C dx \int_{-D}^{-H} v dz = -\int_C dx \int_{-D}^{-H} \frac{1}{\rho} \frac{\partial p}{\partial x} dz. \quad (5)$$

Here (4) specifies the meridional Ekman flux, (5) identifies the net meridional geostrophic flow that can exist below the ridge crests, and (3) merely states that the one must balance the other if the total meridional flow is zero. Like Saunders and Thompson (1993), we take it for granted that this recasting of the form-drag mathematics was recognized long ago, and we only recall it again to attention here.

These relations simply express the familiar idea (e.g., Warren 1990; Toggweiler and Samuels 1993; Tomczak and Godfrey 1994) that net northward Ekman transport across the open circumpolar zone is provided by upwelling south of the zone, which in turn is supplied by a net southward geostrophic flow of deep water below the crests of the major ridges in the zone. Indeed, it was recognized nearly sixty years ago (Deacon 1937; Sverdrup et al. 1942) that the warm deep water of the Southern Ocean derives from north of the Circumpolar Current, and that it upwells into the cooler, fresher surface layer, in which it is returned northward across the current. In all of descriptive oceanography, this is one of the most enduring and firmly based interpretations of tracer properties.

Nevertheless, Johnson and Bryden (1989) doubted that there could be such an upwelling of dense deep water to replenish the lighter surface water because, they said, there was large buoyancy loss over the surface of the Southern Ocean, which would make rising water denser yet. That is not true: the mean buoyancy flux in lats 60°–70°S is *positive.* Buoyancy flux is measured approximately as $\alpha Q/C_p + \rho\beta SF$, where α is the thermal expansion coefficient, Q the heat flux into the ocean (negative Q means cooling of surface water), C_p the specific heat at constant pressure, ρ the surface density, β the haline expansion coefficient ($\approx 0.8 \times 10^{-3}$), S the surface salinity, and F the freshwater flux into the ocean. A wide range of reputable estimates has been made for the heat flux over the Southern Ocean. A recent, careful inference of oceanic energy transports from the global radiation budget and atmospheric transports in 1988, constrained to satisfy zero net heat flux over the land and ocean areas separately, yielded a southward transport across lat 60°S of about 1.5×10^{14} W (Trenberth and Solomon 1994). Divided by the area of ocean between lats 60° and 70°S (1.7×10^{13} m^2), this value implies a mean-annual surface heat flux in that zone of about -8.8 W m^{-2}. However, the mean surface temperature is so low, around -1°C (Gordon 1981), that the mean α is very small, about 0.4×10^{-4} per °C, and the negative contribution to the mean-annual buoyancy flux by the heat flux, -0.9×10^{-7} kg m^{-2} s^{-1}, is much smaller in magnitude than it would have been for an equivalent heat flux into warmer, subtropical water. (A more accurate estimate formed from monthly products of Q and α would have been smaller yet.)

The freshwater flux—precipitation plus prorated glacial breakoff minus evaporation—is not large, perhaps 40 cm yr^{-1} (Gordon 1981), but it is still big enough to make a positive contribution to the mean-annual buoyancy flux, 3.5×10^{-7} kg m^{-2} s^{-1} (1027 kg m^{-3} $\times$ 0.8×10^{-3} $\times$ 34 $\times$ 40 cm yr^{-1}) that substantially exceeds the negative contribution from heat flux. The net buoyancy flux, 2.6×10^{-7} kg m^{-2} s^{-1}, therefore is significantly positive: it is wholly compatible with upwelled deep water becoming lighter as it is transformed into surface water, and it does not imply any idiosyncratic abolition of Ekman-layer transport in the Southern Ocean.

These heat and freshwater fluxes individually buttress the Deacon circulation scheme as well. Hellerman and Rosenstein (1983) estimated the average eastward wind stress at lat 60°S as 0.7 dyn cm^{-2}, which dictates a zonally integrated northward Ekman transport across lat 60°S of 11×10^6 m^3 s^{-1}. The warm deep water is cooled by about 3°C as it rises to the surface (Gordon 1981), and if the upwelling rate is approximated as equal to the Ekman transport (the contribution to the heat budget from formation of Antarctic Bottom Water is neglected here), then this cooling requires a surface heat loss south of lat 60°S of 1.4×10^{14} W (1027 kg m^{-3} $\times$ 4×10^3 J kg^{-1} per °C $\times$ 3°C $\times$ 11×10^6 m^3 s^{-1}), which is very close to the estimate by Trenberth and Solomon (1994) of 1.5×10^{14} W for the energy transport available to supply it in 1988.

Similarly, the salinity in the warm deep water is about 34.7, some 0.7 greater than the mean-annual surface salinity south of lat 60°S of around 34.0 (Gordon et al. 1982). Salinity reduction of that amount requires a surface freshwater flux in the Deacon scheme of 0.22×10^6 m^3 s^{-1} (11×10^6 m^3 s^{-1} $\times$ 0.7/34.7), which agrees exactly with the integrated transport in lats 60°–70°S obtained from Gordon's (1981) rough estimate of flux per unit surface area (40 cm yr^{-1} $\times$ 1.7×10^{13} m^2).

Of course the closeness of these agreements must be partly coincidental. Inasmuch as the Hellerman–Rosenstein (1983) wind stress varies from 0.3 dyn cm^{-2} at 65°S to 1.0 dyn cm^{-2} at 55°S, the pertinence of its value at exactly 60°S for calculating surface-water transport is not so clear. Moreover, on the basis of globally analyzed wind fields, Trenberth et al. (1990) believe that Southern Ocean wind stresses are something like twice as great as estimated by Hellerman and Rosenstein (1983), while at the other extreme Harrison (1989) used a different formulation of drag coefficient from theirs to generate an Ekman transport at lat 59°S of 7×10^6 m^3 s^{-1} *southward* (in contrast to the Hellerman–Rosenstein value there of 12×10^6 m^3 s^{-1} northward). As to the heat flux over the Southern Ocean, Hastenrath (1980), Emig (1967), and Gordon (1981) have variously used bulk-formula calculations of the surface flux to estimate southward energy transports across lat 60°S of 1.3 $\times 10^{14}$ W, 3.9 $\times 10^{14}$ W, and 5.4 $\times 10^{14}$ W, respectively. (But even with the largest of these values the surface buoyancy flux in lats 60°–70°S remains positive.) Thus, the Deacon scheme of meridional circulation is easily consistent with respectable constructions of heat and freshwater budgets, but there is still enough spread in estimates of the observational ingredients that its proof continues to rest mainly on the tracer-property fields and simple dynamics.

Since it is this meridional circulation to which the "form-drag" equation (3) actually pertains, it follows that (3) includes no attribute (e.g., volume transport) of the Circumpolar Current. It does not even demand that the current should flow eastward rather than westward. In fact, according to the resolution (4) and (5), the Circumpolar Current is irrelevant to the integrated zonal momentum balance and could actually be absent without affecting (3), (4), and (5).

The notion of form drag was introduced to discussion of the Circumpolar Current by Munk and Palmén (1951), who sought an agency to oppose the eastward wind stress acting upon the meridionally unbounded Drake Passage zone. A more conventional, straightforward answer to their quest is the Coriolis acceleration (not discussed by Munk and Palmén) associated with the northward Ekman transport. If Ekman-layer dynamics applies to the Drake Passage zone as much as is thought to the rest of the world ocean, then the surface stress should be absorbed in a thin boundary layer in which the stress divergence drives Coriolis-balanced motion, and there is no question of any of that stress being directly transmitted to great depth in order to be taken up in some other way.

It is then only necessary to find a southward flow to supply, through upwelling, the northward Ekman transport. This can be a deep, net geostrophic flow because the ridges spanning the circumpolar zone can support the required pressure difference as integrated circumpolarly. This pressure difference is thus proportional to the wind stress, as in the Munk–Palmén algebra, but not because that is intrinsic to the dynamics of the Circumpolar Current, which has not even been mentioned here; rather, the proportionality is required simply for meridional mass balance, and it would still be required whatever the size or direction of the Circumpolar Current. It is for this reason that we believe form-drag talk is obscurantist in relation to the Circumpolar Current: the physics that it really describes seems to have nothing to do with that great current, but instead with mass conservation in the quite independent meridional circulation.

[Of course, if the ocean bottom were level, and the upwelling were supplied by southward flow in a bottom Ekman layer, then the meridional circulation would dictate a zonal flow (e.g., Gill 1968). But there is no such linkage when the southward flow is geostrophic.]

It is doubly odd that interest should have continued to focus on form drag, because a Sverdrup-dynamics interpretation of the Circumpolar Current was sketched long ago by Stommel (1957), was elaborated by Wyrtki (1960), and has been supported quantitatively by Baker (1982), Godfrey (1989), and Chelton et al. (1990). In its eastward course from the western South Atlantic to the eastern South Pacific, the current veers southward by some 10° of latitude, which seems entirely consistent with local forcing by negative wind-stress curl. Its equatorward return flow along South America, the Falkland Current, looks like a subpolar western-boundary current, crossing a zone as broad as that traversed by the Gulf Stream between southern Florida and Cape Hatteras. The pattern of the circulation is just twisted drastically out of shape from the canonical gyre of the rectangular basin.

In this dynamical scheme the volume transport of the Circumpolar Current in Drake Passage, which is the upstream end of the western boundary current, is proportional to the circumpolar integral of the wind-stress *curl* around a latitude near Cape Horn. Baker (1982) calculated this integral along lat 55°S from long 40°W eastward to long 70°W for two sets of wind data, and obtained southward transport values of 173 and 190 $\pm$ 60 ($\times 10^6$ m^3 s^{-1}). For the Drake Passage, Whitworth et al. (1982) reported combinations of ISOS current records and density sections made to estimate the volume transport of the current at four different times (in the summer season for three years, fall for one). Their results were 117, 124, 137, and 144 ($\times 10^6$ m^3 s^{-1}), for an average of 130 $\times 10^6$ m^3 s^{-1}, and a variation of not more than 11% from the average. A time series of volume transport, estimated from a calibrated, four-year record of cross-passage pressure difference, showed a standard deviation also of only about 8% of the mean, but occasional departures as large as 35 $\times 10^6$ m^3 s^{-1} (Whitworth and Peterson 1985). Baker's (1982) wind-stress-curl values of transport are higher, but are not unbelievably discordant if, as he suggests, his wind-stress estimates were uncertain by 50%.

Using the more recent and presumably more reliable wind-stress compilation of Hellerman and Rosenstein (1983), Godfrey (1989) repeated Baker's (1982) calculation, for lat 54°S, and obtained an annual-average transport of 128 $\times 10^6$ m^3 s^{-1}. For lat 55°S Chelton et al. (1990) obtained 114 $\times 10^6$ m^3 s^{-1} from these data and they attributed the difference between the two values to their smoothing of the wind-stress field. From the same set we calculated circumpolar integrals of 107 $\times 10^6$ m^3 s^{-1} at lat 55°S and 130 $\times 10^6$ m^3 s^{-1} at lat 60°S. The uncertainty in these results is large: Hellerman and Rosenstein estimated error of around 50×10^6 m^3 s^{-1} in their Sverdrup-transport calculations south of about lat 40°S, and Chelton et al. (1990) believe that the error in those authors' wind-stress values in high southern latitudes is actually worse. As another gauge of the uncertainty, the range in monthly mean values of the circumpolar integral at lat 55°S derived from Hellerman and Rosenstein's (1983) *monthly mean* estimates is some 40×10^6 m^3 s^{-1} about the annual mean. Thus, this recent close approach of wind-stress-curl estimates of the Drake Passage volume transport to the average measurement of Whitworth et al. (1982) must be partly fortuitous, but it gives impressive support nonetheless to the Sverdrup-dynamics theory of the Circumpolar Current.

Sverdrup balance assumes that the vertical velocity is zero on the floor of the ocean. However, that condition must be violated locally in the Southern Ocean because certain perturbations to the path of the Circumpolar Current are correlated with bottom topography (Gordon et al. 1978); this correspondence suggests that the current reaches to the bottom, flows upslope and downslope to some extent, and thus experiences nonzero vertical velocities at the bottom. But these can be only local features, with upslope flow across ridges compensated by subsequent downslope flow, so that the regionally averaged vertical velocity must be zero, and the Sverdrup relation should hold in that regional sense.

It might be noted that Sverdrup dynamics is not possible in models of zonal flows in zonal channels (e.g., Straub 1993) because those models neither accommodate the large meridional excursion of the flow nor support the required western boundary current. To that extent they do not seem applicable to the Circumpolar Current. Furthermore, quasigeostrophic channel models (e.g., Treguier and McWilliams 1990; Wolff et al. 1991) do not permit alteration of water properties in their respective layers. Because the real ocean is not thus constrained by a layerwise conservation of mass, as is evident and necessary in the Deacon picture of cross-current circulation, these channel models too lack essential features of Circumpolar Current dynamics.

There are unresolved issues in the wind-stress-curl theory, such as how the Scotia Arc functions as an "eastern boundary," and what sort of pressure field must be associated with the streamline field, but the theory does explicitly predict the volume transport of the Circumpolar Current with tolerable accuracy, from clear physics. Equation (3), concerning the stress rather than its curl, coexists with the Sverdrup relation as a valid balance, although the two govern different components of the flow field. Adding a constant to the field of zonal wind stress, for example, would alter the meridional circulation, but, leaving the curl unchanged, would not affect the transport of the Circumpolar Current. Therefore, the cross-ridge pressure difference equated to the wind stress in (3), misleadingly called form drag or the like, does not control the Circumpolar Current either; instead it supports the meridional transport balance in the zone of Drake Passage.

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