Essay Review

Georg Wüst. "Quantitative Untersuchungen zur Statik und Dynamik des atlantischen Ozeans; sechste Lieferung: Stromgeschwindigkeiten und Strommengen in den Tiefen des atlantischen Ozeans". Wiss. Ergebn. deutsch. atlant. Exped. "Meteor" 1925–1927, **6**(2): 260–420. 1957.

The ocean is the great stabilizer and works much like the fly-wheel of an engine. Like a fly-wheel it responds to the systems it controls and so there is constant variation around a mean or steady state. If the oceanic fly-wheel were perfect we should have neither weather nor fluctuations in the output of our sea-fisheries. To understand our weather and many other practical problems we therefore need to understand the fluctuations in the nature and circulation of the ocean about the mean state.

Although the fluctuations are all that really matter and the mean state is a statistical fiction, the assessment of the mean circulation is one of the most important and most difficult tasks that the oceanographer is called upon to tackle. We need to know the mean geostrophic circulation of the oceans and seas everywhere on earth and a statistical evaluation of the nature of the waters being circulated. We must not restrict ourselves only to horizontal movements but must consider the structure and the vertical movement in the ocean as well.

One technique only is available and able to give even an approximate estimate of circulation in a whole ocean, a study of the field of mass derived from serial measurements of temperature and salinity on water samples captured by instruments suspended by wires from ships. Then with the Bjerknes circulation theorem the geostrophic currents may be calculated providing that at some depth a reference surface in which no motion occurs may be recognized. The early applications of the theorem were to surface currents, assuming some level surface of no motion. Although arguments in favour of defined "depths of no motion" were often made, one suspects that often the criterion was the amount of wire carried by the ship's hydrographic winch or the amount of time that could be spared for working each station.

In spite of the shortcomings, the circulation theorem with an arbitrary level of no motion usually at a depth between 1000 and 2000 m has given a most useful picture of the behaviour of a surface layer 100 m or so thick, but deep water movements could be studied only in the crudest outlines from the distribution of readily measurable properties such as salinity and oxygen content.

A critical advance was made 20 years ago by Defant (1941) and described in English in his text book (1961). For neighbouring stations in the ocean, Defant compared a large number of curves of differences in dynamic depth of the pressure values to show that in each profile there is a layer of considerable vertical thickness in which the differences in dynamic depth are constant or almost constant. He suggested that this prominent layer is motionless or almost motionless so that the reference surface of no motion should lie within it. The reliability of this method is much increased if individual reference depths determined from a large number of station pairs may be combined to give an extensive topography of the reference surface. Broadly speaking, this surface slopes from a depth around 500 m on the equator to depths well in excess of 2000 m towards the poles.

DEFANT, having provided the means, applied it in the upper layers or troposphere of the South Atlantic. WÜST, in the paper under review, has applied it with great success to the deepest waters or stratosphere of the South Atlantic, using the station data obtained by the "Meteor" in 1925–27.

His most remarkable finding was that currents as strong as 3–17 cm/sec. (0·06–0·33 knots) occur in the deep sea between 3000 and 5000 m but only along the western margin of the South Atlantic, so conforming to the theoretical requirements of STOMMEL (1948, 1958). The southerly components of these currents occur in the North Atlantic Deep water which may be traced backwards from about 45°S Lat. along the continental slopes of South America, the Antilles, and North America, around the bottom of the Labrador Sea and Greenland to its main origin in the Denmark Strait where heavy, cold Norwegian Sea water overspills. One component towards the North consists of Antarctic bottom water, much of it from the Weddell Sea. Nothing comparable was found in the eastern basin of the South Atlantic.

Striking was the computing of a secondary level of no motion near 4000 m depth, just where it was required at the boundary between the North Atlantic deep current and the Antarctic bottom current but, again, only on the western side of the profiles.

The volume transports of water through each of the South Atlantic "Meteor" sections were computed. The southerly North Atlantic deep current almost exactly compensated the combined northerly transport in the upper layers and in the Antarctic bottom water. The principle of continuity is well satisfied.

The markedly different distributions of temperature, salinity and oxygen in the eastern and western basins of the South Atlantic also find a sufficient explanation in terms of the very different deep circulations.

Geostrophic calculations which are generalized and statistical in nature require confirmation by direct measurements of deep currents. Direct measurements, though of high precision, are not readily generalized beyond the place where, and the time when, the measurements have been made. Greatest confidence comes when the two approaches can be integrated as in the western North Atlantic off Blake Plateau by SWALLOW and WORTHINGTON (1961). Here, using two ships, direct measurements with neutral buyoancy floats and the geostrophic method, were combined with great effect. Currents between 9 and 18 cm/sec (0·18–0·35 knot) were observed very similar to those computed by WÜST in similar situations in the South Atlantic. This consistent southerly current off the Carolinas contrasted with the very variable and much weaker currents with a large tidal component found by SWALLOW and HAMON (1959) in the Eastern North Atlantic. Western intensification occurs in both North and South Atlantic Oceans, as required by STOMMEL.

Now that DEFANT has shown how to derive a realistic surface of no motion at depth, the work of WÜST and of SWALLOW and his associates from the Woods Hole Oceanographic Institution and the National Institute of Oceanography, has shown that direct measurements of deep current are well in step with currents computed from the distribution of mass. In the Atlantic a sufficient assessment of the mean circulation at all depths is within sight so that the next stage, study of the deviations from the mean, can be begun using very largely the same techniques. These have the necessary precision but fall far short of the necessary accuracy. Precision and accuracy will have to be brought nearer together by a series of successive approximations.

The geostrophic method applied to the deep sea requires that the measurement of density should attain a relative accuracy of 1 part in 1 million. Indeed, measurements accurate to 1 part in 10 million could be put to good use in waters deeper than 4000 m.

Relative precision in measurements of density to 1 part in a million, is not easily attained even under optimum conditions ashore so that a quantity more readily measured must be used. The choice lies between precision titration of chlorinity (HERMANN, 1951; BATHER and RILEY, 1953) and conductivity (SCHLEICHER and BRADSHAW, 1956; COX, 1958). The relative precision of these two methods is very similar and both are better than the traditional Knudsen titration by an order of magnitude. There is no question but that conductivity provides much the easier measurement, notably so at sea. COX *et al.* (1962) have shown that deviations of computed from observed densities are considerably less using conductivity rather than chlorinity as the basic measurement. The computations of density have to be made by adding additional ciphers to the basic terms given in KNUDSEN's hydrographical tables. These terms expanded in this way can have no true physical or chemical meaning but, nevertheless, they are an expedient to enable us to use the more refined data we are now getting to arrive at more refined geostrophic relative assessments of ocean currents.

DITTMAR (1884), who made the analyses of sea salt on which most of our fundamental thinking is based, realized that the relative proportions of the several ions in sea water varied slightly in various parts of the world ocean but it became convenient to assume, for purposes of geostrophic calculation, that the composition of sea salt is invariant. In course of time the convenient assumption became dogma. A great debt is due to CARRITT and CARPENTER (1959) for recalling what DITTMAR wrote and the dangers of the dogma now that we can make measurements of greater precision.

The small but important departures of the true density from density computed from temperature and either chlorinity or conductivity arise from four main causes (a) variations in the composition of the salts in drainage from the continents, (b) fractionation of sea salts during concentration by freezing out of ice in polar regions, (c) precipation and solution of specific salts, notably calcium carbonate, (d) fractionation of hydrogen and deuterium, ^{16}O and ^{18}O in water during its meteorological cycle. Fractionation of other isotopes, such as those of carbon and sulphur, is relatively trifling. Since the causes of the deviations of computed from true density are so many and various, a direct empirical attack on the problem is required. Each main deep water mass of the world ocean will need to have its own characteristic relationship established between density and pressure and the determinations on which they are based. Only a very extensive investigation, such as that in progress at the National Institute of Oceanography and the University of Liverpool, Great Britain, has any chance of providing this. But once provided, the geostrophic method gives a means of keeping tab on the fluctuations of the deep oceanic circulation, very likely to affect climate and fisheries in shallow seas.

Table 1

Potential density at Station Cavall (46°30′N, 8°00′W) between depths of 3138 and 4656 m based on salinities with two NIO salinometers

	Date and range of depths	Number of samples	Mean potential density and standard deviation
A.	All depths on 3. November, 1958 ("Discovery II")	8	27·9094 ± 0·0014
	All depths on 14. May, 1960 ("Sarsia")	20	27·9092 ± 0·0013
	Both combined	28	27·9093 ± 0·0010
B.	Sub-divided by depth ranges		
	3138–3440	6	27·9092 ± 0·0012
	3707–4060	9	27·9097 ± 0·0046*
	4156–4656	13	27·9089 ± 0·0027
C.	Further sub-divided		
	3138–3169	3	27·909
	3232–3440	3	27·910
	3707–3904	3	27·908
	4000–4060	6	27·911*
	4156–4200	2	27·908
	4300–4356	2	27·909
	4400–4656	9	27·909

* The larger standard deviations and higher mean values are due to two determinations (27·915, 27·915) at 4,030 and 4,040 m where anomalous oxygen results were also obtained. Without these the mean potential density between 3,707 and 4,060 m would have been 27·908.

Many oceanographers have had doubts as to the dependability of the tables for correcting the effect of pressure at depth on density (e.g. Eckart (1959)). These have been based on the work of Ekman (1908, 1914) and Helland-Hansen (cit. Sverdrup *et al.*, 1942, Chap. 3, Table 12). Recent work by Crease, Catton and Cox (1962) has shown that these tables seem to be remarkably accurate. Indirect evidence has also been obtained in the Bay of Biscay (Table 1), where potential density has been computed from conductivity salinities and potential temperatures derived from the observed temperatures with the aid of these tables. The conductivity measurements on R.R.S. "Discovery" were made on board by Mr. J. Crease, whereas those from R.V. "Sarsia" were made ashore in the Plymouth Laboratory with another NIO salinometer by Mr. E. I. Butler. Their mean measurements differ only by 0·0002 on σ_θ or 0·0000002 on the density. Within the limits of error of our measurements all the water between 3150 m and the bottom in 4700 m is in neutral adiabatic equilibrium with a potential density (σ_θ) 27·9093 ± 0·0010.

These findings are offered with some misgivings since the parallel chemical work has shown assured evidence for quite strong stratification in the deep water. Either the chemical stratification is maintained by differences in density of the order of 1 in 10^7, beyond our power to measure, or the adiabatic correction coefficients are slightly too small, or, in the deep water, there is a systematic small error in computed density.

It is a measure of the speed of advance and of the stimulus to study deep circulation provided by Wüst that already we should be able to formulate practical means of studying not only the mean state but the all-important deviations from this mean.

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