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An observation of a deep countercurrent in the Western North Atlantic

J. C. SWALLOW* and L. V. WORTHINGTON†

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INTRODUCTION

THE DEPTH of the level of no motion under the Gulf Stream has been a matter of interest to oceanographers for many years, and widely-varying estimates of the volume-transport have been arrived at, according to the choice of reference level. DIETRICH (1936) used the oxygen-minimum as a reference surface, at a depth of about 800 metres, but it was pointed out by ROSSBY (1936) that this choice implied the existence of a deep counter-current with a transport more than double that in the upper layers. ISELIN (1936) in his study of the circulation of the western North Atlantic, presented currents calculated relative to the 2000-decibar surface, and from the observations available at that time it appeared that the dynamic topography in the deeper water was relatively weak. DEFANT (1941) chose a surface of no motion sloping downwards from west to east, at a depth of about 1500 m under the Gulf Stream off Cape Hatteras, and calculated southward velocities of the order of 3 to 6 cm/sec at 2000 m depth. The high oxygen content of the deep water in this region was recognized as being consistent with a southward flow (ISELIN, 1936) but the notion of a current increasing in velocity towards the bottom was rejected by many oceanographers as unlikely (SVERDRUP *et al.*, 1942, p. 456).

Three recent developments led to renewed interest in this problem and to the measurements described here. During the last few years the R.V. *Atlantis* of the Woods Hole Oceanographic Institution has made a number of more detailed hydrographic sections across the Gulf Stream, paying particular attention to the deep water, and it has become clear that the steep temperature gradients across the stream, already well known in the upper layers, persist all the way to the bottom (FIG. 1, Cape Romain Sect.). This implies, if the geostrophic equation is valid, a relatively strong narrow flow in the deep water, northward if the reference level is at the bottom, or southward below any mid-water reference level. No satisfactory choice could be made from considerations of continuity from one section to another, nor from the temperature-salinity relationship of the water masses themselves.

At the same time, STOMMEL (1956, 1957) devised a theoretical model of the circulation of the Atlantic Ocean which required, as an essential feature, the existence of a narrow southward current in the deep water at the western boundary.

A suitable instrument for making direct measurements of such a current was recently developed (SWALLOW, 1955, 1957) in the form of a neutrally-buoyant float, and it was arranged that the R.V. *Atlantis* should co-operate with the R.R.S. *Discovery*

*National Institute of Oceanography.
†Woods Hole Oceanographic Institution.

II of the British National Institute of Oceanography in current-measuring and water-sampling, in an attempt to detect the south-going current and to locate the level of no motion.

Parachute drogues (VOLKMANN, KNAUSS and VINE, 1956) were also available for current measurements at all depths, and a new method had been devised by A. S. LAUGHTON for measuring currents very near the bottom using an underwater camera.

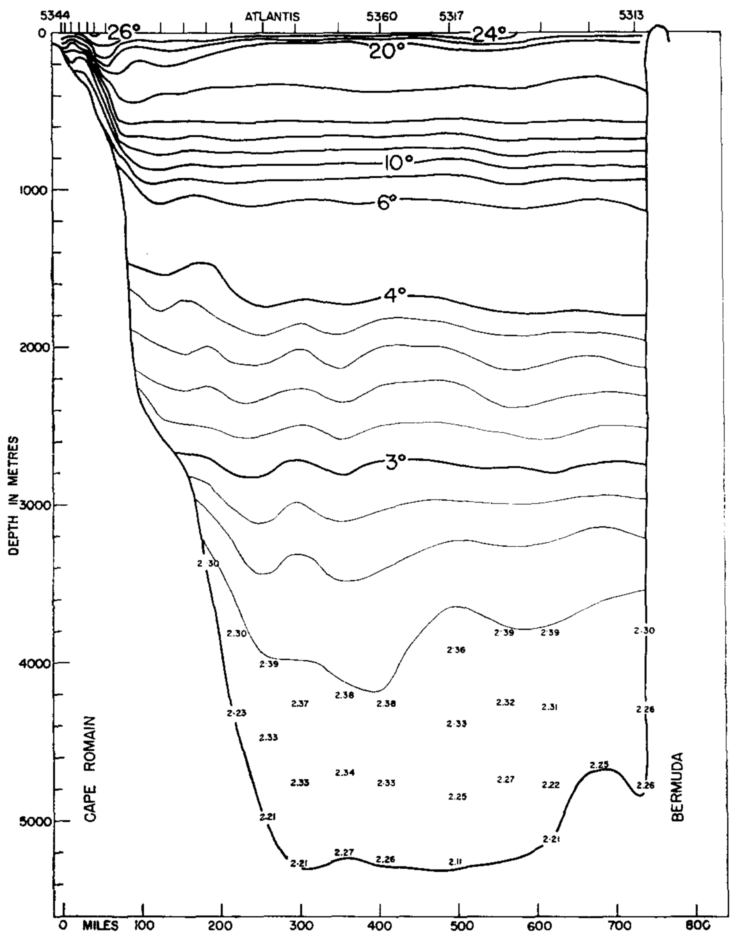

FIG. 1. Temperature Section, Cape Romain to Bermuda, made by *Atlantis* in June 1955.

It was planned that the *Atlantis* would first find the region of steep temperature gradients in the deep water, and then the *Discovery II* would make current measurements whilst the *Atlantis* occupied more hydrographic stations in the same area. Previous work had shown that, north of Cape Hatteras, the deep temperature gradients lay under the surface stream, and it was thought that surface currents of several knots would make deep current-measuring difficult. STOMMEL suggested that the most favourable area would be off Cape Romain, South Carolina, where the surface stream flows over the Blake Plateau, and the deep temperature gradients are found farther off-shore at the foot of the continental slope. Farther south, the deep temperature gradients become weaker, and it was decided to work as near as was practicable to the point of departure of the surface stream from the Blake Plateau.

METHODS OF MEASURING CURRENTS

In tracking neutrally-buoyant floats or parachute drogues, the accuracy of the measurement is limited by uncertainties in fixing the ship's position. The area chosen for the present measurements was within the ground-wave coverage of Loran stations 1H6 and 1H7, and good fixes could be obtained from them at most times. For the greater part of the work, the Loran readings were combined with radar fixes on an anchored buoy. This allowed averaging over groups of Loran readings, giving an improvement in accuracy and an estimate of the remaining uncertainty.

The parachute drogues, and their method of use, have been fully described by VOLKMANN, KNAUSS and VINE (1956), and the same technique was used in the present measurements, except that more use was made of radar in fixing the drogues. Although it seemed possible that a fairly strong deep countercurrent, of the order of 10 cm/sec, might be detected by this means, the results with the deep drogues were rendered uncertain by strong variable winds and surface currents, and only the measurements in the upper layers are considered reliable. Surface currents were measured occasionally by releasing fluorescein dye near the anchored buoy, and tracking it for a few hours.

One successful measurement of current near the bottom was made by the photographic method. A deep-sea camera (LAUGHTON, 1957) was mounted on a tripod looking vertically downward, and took a series of 60 photographs while on the bottom for 24 minutes. The orientation and tilt of the tripod were obtained from a combined compass and clinometer in the field of view. Two methods of current registration were used. First, drops of neutrally-buoyant aniline were ejected into the water from a spring-loaded hypodermic syringe. The velocities were higher than expected however, producing a large number of small drops (more than one per picture) so that only the direction and not the speed of the current could be obtained from them. The second method used was to observe the deflection from the vertical of a free-flooding ping-pong ball suspended on a thin thread. This angle could be calculated from the photographs, dimensions of the camera tripod and ball suspension being known. A calibration was made by towing a similar ball suspension in a tank at known velocities and measuring the deflection. Unfortunately, in an attempt to make another current measurement, the entire assembly was lost. The calibration could not therefore be made on the original ball, and one having the mean properties of six different makes was used instead.

The neutrally-buoyant floats, and the equipment used for receiving their sound signals, have already been described (SWALLOW, 1955, 1957). Only a brief account has been given previously of the method of operation, and some further details are given below.

First, the float has to be loaded to make it sink to the required depth. This load is calculated as follows. Each complete float is made neutrally-buoyant in a tank of salt water of known density and temperature. This adjustment is made to within 1 gm in a float weighing 10 kg, or 1 in 10^4. Suppose the density is ρ_0 at temperature T_0. Then the density of the float at depth D meters, where the temperature is T, will be approximately

$$\rho_0 (1 + kD - \alpha (T - T_0))$$

where k is the bulk compressibility of the float, per metre depth change, and α is the coefficient of volume expansion of the float material. For the tubes used, k was $15 (\pm 1) \times 10^{-7}$ per metre, and α was 69×10^{-6} per °C.

If the density of the sea water at the required depth D is ρ *(in situ)* obtained from observations of temperature and salinity, then the extra load needed to make the float sink to that depth is

$$\delta M = M \left(\frac{\rho \, (in \; situ)}{\rho_0 (1 + kD - \alpha (T - T_0))} - 1 \right)$$

where M is the mass of the float.

An allowance has to be made for movement of the rubber sealing rings in the end plugs, under pressure, and for the extra compressibility of the insulation on the transducer winding. Previous experience suggested that the calculated load should be reduced by 6 gms to allow for these effects.

Each float was tested for leaks, and for satisfactory working of its sound transmitter, by lowering it on a wire to more than the required working depth (usually 3000 m). If all was well when the float was recovered, the calculated load would be added and the float dropped over the side at a chosen position. They take three or four hours to sink to the stable depth, and usually no fixes were taken during that period.

Bearings could be obtained on the acoustic transmitter by stopping the ship head-to-wind and putting two hydrophones over the side, about 30 m apart. As the ship fell away from the wind direction, the path difference between the signals arriving at the two hydrophones was observed as a function of the bearing of the ship's head relative to north. FIG. 2 is a block diagram of the receiving equipment. The path difference was measured as a time delay on an oscilloscope. It was usually

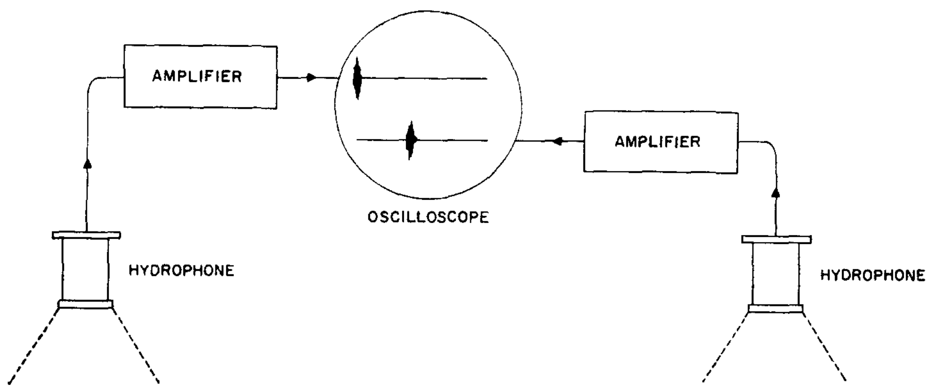

FIG. 2. Block diagram of receiving equipment.

in the range 0 to 20 milliseconds and could be estimated with an accuracy of about ½ milli-second. The amplifiers were tuned to 10 kc/s, with a bandwidth of about 500 c/s. In taking a bearing, a series of from six to ten readings of the time difference t was taken, while the ship's head swung through an arc of about 60°. If t_{max} is the time difference that would be observed with the ship heading directly towards the float, then

$$t = t_{max} \cos (\phi - \theta)$$

where ϕ is the heading of the ship, and θ is the bearing of the float, relative to north (FIG. 3). To determine θ and t_{max} from the series of observations of t and ϕ, rays were plotted at successive values

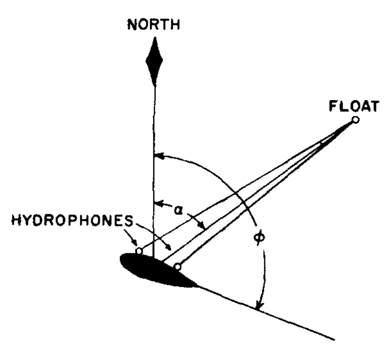

FIG. 3. Taking a bearing on a float.

of ϕ, and lines were drawn at right angles to them at distances from the centre proportional to the corresponding values of t. These lines should all pass through a point, at a distance proportional to t_{max} from the centre and on the ray making an angle θ with north (FIG. 4a).