Currents and temperature stability over the Antarctic Bottom Current at 10° N

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ABSTRACT

Instrumented neutrally buoyant floats were used to investigate some characteristics of the temperatures and currents associated with the Antarctic Bottom Current at 10° N. An average current of 4.6 ± 0.5 cm sec^{-1} in the 325° direction was found over a period of 5 days at a distance 540 m off the bottom. This and other values obtained during the sinking of a float are in substantial agreement with the predictions of G. Wüst. Vertical water motions at the hovering float did not exceed 28 m rms and so compare with those present in quieter locations.

Temperature profiles were obtained from floats which sank to the bottom and then returned to the surface. First and second derivatives of the temperaure with respect to depth are reported for 100 m depth intervals. Stabilities E were determined and log E varied linearly with depth near the sea floor.

Antarctic Bottom Water spills over the sill of the Para Rise at 2° S off the coast of South America and flows northward down a gentle bottom slope while it mixes slowly with overlying North Atlantic Deep Water. The Bottom Water contains 60 percent of its Antarctic source water near the equator and that amount is reduced to 28 percent at 10° N. Many of the characteristics of the Antarctic water, from its source to its last stages of dissipation near 40° N, have been investigated by Wüst (1933, 1957) who applied classical oceanographic techniques to data based in large measure on the results of the *Meteor* Expedition of 1925–1927. Little has been done to verify the general velocity structure which he predicted for the deep waters of the Atlantic or to profit from studying details of the clearly defined mixing processes involved. This report describes measurements of variations in the deep temperature structure of the water column at 10°24′ N, 50°22′ W as well as measurements obtained from a neutrally buoyant float placed in the upper portion of the Antarctic Bottom Current at that location.

Instrumented neutrally buoyant floats were used to obtain the data. Such floats transmit temperature, pressure and distance information to a listening ship by means of sonic pulses. They provide detailed temperature profiles when adjusted to sink slowly to the bottom of the ocean or they supply information on temperature and pressure fluctuations when they are tethered to the sea floor or allowed to hover freely at a desired depth. Further details are described in past work by the author (Pochapsky, 1963). A new modification of the float structure is described in the next section along with the results obtained from a hovering float.

Hovering float experiments

Only one hovering float was used in the present work. This was placed at a depth of 4390 m, 540 m off the bottom. Position, temperature and pressure information was obtained from it during the time it sank and while it remained at depth for a period of 5 days.

This float had a large bag of water attached in order to bring about a closer correspondence between the movements of the float and the water around it. The bag, made of polyethylene sheet, filled near the 1000 m level and formed a closed cylinder of water, 61 cm in diameter and 73 cm long. It has a negative buoyancy of 680 gm and was connected to the buoyant aluminum-shell float above it with a 15-m length of line. The volume of water encased was 212 liters as compared with the 23.6-liter volume of the instrumented float itself. As a result of this

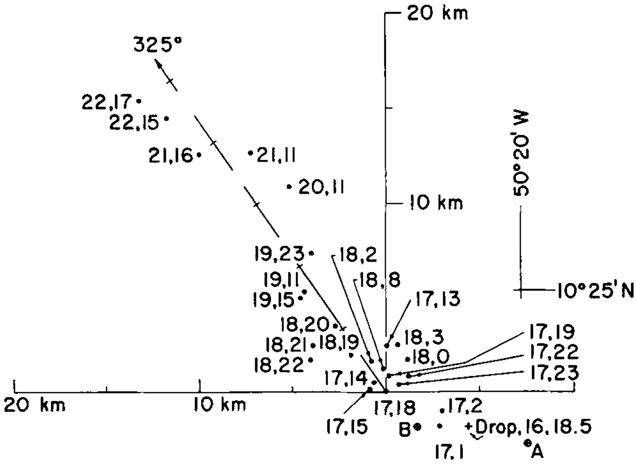

Fig. 1. Geographic position of hovering float. The first numeral represents the date, October 1967, and the second the hour, UT. The float reached equilibrium depth, 4390 m, at 17, 18.

construction (1) the effective float compressibility was increased, (2) the float resonant frequency was decreased and (3) viscous drag was increased. All these factors increase the coupling between the float and the surrounding water but despite the large volume of water in the bag, the float should still attain only 17 percent of the vertical amplitude of surrounding water at depth for low frequencies having periods measured in days. Near and below the float resonant period, 1.1 hr, the correspondence between vertical water and float motions would be almost complete. To the extent that the float does not follow the water, this procedure makes uncertain any deduction of water response from float response. As an example, temperature changes of the encased water have to be considered; it is estimated that the thermal time constant of that is approximately 15 hr.

The geographic position of the float was determined during the later stages of sinking and then while it drifted at equilibrium depth. Observations are illustrated in Fig. 1. Positions A and B, noted in that figure, mark two reference floats which were anchored so as to be 15 m off the sea floor. Near these positions, the location of the hovering float could be determined to ± 1 km. During the last few days of the experiment, however, reliance was placed on star fixes and dead reckoning back to position B. At that time, the uncertainty of location was approximately ± 2 km.

The hovering unit was inserted in the water on 16 October, 1967 at 18 h 30 m UT. For convenience, positions are marked by the day, followed by the hour, and so the entry position is denoted by 16, 18.5 in Fig. 1. At 17, 1, or 6.5 hr later, the float was at a depth of 2300 m and sinking at a rate of 6 cm sec^{-1}. At 17, 13, it was at a depth of 4200 m and sinking at approximately 1 cm sec^{-1}. Between these two last positions, the average horizontal velocity was 12 cm sec^{-1} in a NW direction. At 17, 18, equilibrium depth, 4390 m, was reached and the motion during the preceding 200 m was southerly at approximately 14 cm sec^{-1}.

While the float sank, it communicated occasionally with the reference A. From the signal transit times it was possible to calculate velocities relative to A. The precision was not great but a dozen values ranging from 8 to 12 cm sec^{-1} away from A were obtained between the 1000 and 3000 m levels. Near the 4000 m level, there was a velocity component *towards* A at a few cm sec^{-1}, in substantial agreement with the southerly course noted above.

During the 5 days of movement at equilibrium depth, the free float moved with an average velocity of 4.6 ± 0.5 cm sec^{-1} in the 325° direction. For convenience, this direction is drawn in Fig. 1 and marked into five equal divisions. The erratic nature of the float track is believed to be real, at least during the first few days when the navigational accuracy was good.

These results for the velocity compare as well as may be expected with the geostrophic velocities computed by Wüst (1957) for this loca-

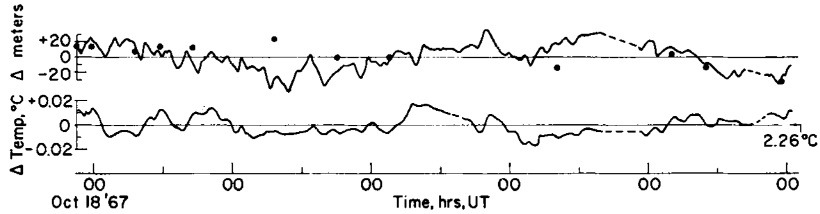

Fig. 2. Temperature and vertical position at hovering float as a function of time. ● represent changes in distance above an assumed flat bottom. The upper curve is seriously contaminated with pressure gage noise and is an upper limit to depth changes.

tion. He predicts the average velocity to be in a northwest direction in most of the deeper water and expects velocities slightly in excess of 7 cm sec^{-1} between 2500 m and 4300 m. Upwards, they decrease so as to be less than 1 cm sec^{-1} at 1000 m but they remain near 5 cm sec^{-1} below 4300 m. A strict comparison can only be made for the 5-day average at the 4400 m level where Wüst predicts 7 cm sec^{-1} compared to our 4.6 cm sec^{-1}. The fluctuating character of the ocean is such that the velocities obtained during the descent of the float may be far from true averages; the southerly component between 4200–4400 m is probably only a temporary condition. With these reservations, our results confirm the direction expected by Wüst but suggest that the velocities may be 50 percent higher at mid-depths and 20 percent lower at a location 500 m off the bottom.

Temperature and pressure values were each obtained at 20-sec intervals during hovering except at the times when the ship was in full motion. In addition, whenever the ship was in a position almost over the float, measurements were made of the distance between that instrument and the sea floor. The total depth of water was practically constant, but there were undulations which give a ± 15 m uncertainty to the location of the true bottom beneath the float as determined on the ship's fathometer. The relative height of the float, obtained from echo times, is uncertain to no more than the ± 10 m, which is associated with a 5° uncertainty in the depression angle from the ship to the float. All float depths obtained when a constant bottom depth is assumed are consequently uncertain to ± 25 m. The measured variations, illustrated as solid points in Fig. 2, lie within these limits and so the measurements restrict any vertical movements to be less than a total of 50 m.

Attempts to obtain changes in depth from pressure readings were thwarted by a noisy gage. There was a linear trend of 470 m over 5 days and a strong component having a 25 m amplitude and 94 hr period; both were obviously spurious. When these were removed from the data, analysis showed the spectrum density decreased with the frequency, f, as $f^{-1.2}$ below the float resonant frequency and f^{-2} above it. The latter rate is particularly disturbing because it can be associated with the noise spectrum of metal films such as were used in the gage. The standard deviation of the detrended data was 11 m which, because the gages were operating at full sensitivity, represents the sum of vertical float movements and gage noise. Most of this took place at the lowest frequencies where the characteristic periods are of the order of a day. The contribution from movements at frequencies between f_v and $f_{v/2}$ where f_v is the Väisälä-Brunt frequency, was only 3.8 m so that in this region, where the float and water tend to move together, the water amplitude was quite small. The Väisälä period was 2.3 hr. No definite vertical jetting of the water was observed in the original data, which are illustrated in Fig. 2. The limitations of these linearly detrended depth data should be kept in mind.

Observed temperatures are also illustrated in Fig. 2. Analysis showed that the spectrum density fell with the -2.3 power of the frequency for periods between 2 days and 2.3 hr. At higher frequencies, the density fell much more rapidly for the next decade of power, at which point the noise threshold was reached. Periods longer than 1 day contributed 0.0113°C to the rms deviation while all periods longer than the stability period contributed 0.0129°C. Any further contributions down to the Nyquist period, 10 min, were ignorable.

Float A, anchored 15 m above the bottom at the position shown in Fig. 1, transmitted single temperature and pressure readings every 5 min. Over a period of 8.5 days, starting early on 13 October, the total linear drift was $+0.0085$°C. Variations about this trend, averaged over 6-hr periods, were less than ± 0.004°C except for readings that were ± 0.01°C higher during 20 October. Outside of these latter high readings, most of the variation can be considered to be possibly of instrumental origin. It can be concluded that the temperature fluctuations at an Eulerian point close to the bottom did not exceed ± 0.01°C over a period of more than a week. As usual during these experiments, the pressure data were of poor quality but there were no fluctuations exceeding the instrumental noise threshold of ± 2 m.

The small temperature changes at position A suggest that even smaller changes would take place at a moving point. If so, any temperature changes at the hovering float would be the consequence of relative vertical motions between the float and water. This is not necessarily entirely true because hovering took place appreciably farther from the bottom. On the assumption that it is, a low-frequency rms amplitude of 0.011°C at the float in the existing adiabatic temperature gradient 0.48×10^{-6}°C cm^{-1} implies a relative water amplitude of 23 m. From the float characteristics, it can be shown that the true water amplitude is 28 m rms while the float amplitude is 4.7 m. The latter figure is half that shown by the noisy pressure gage.

This section can be concluded by stating that the temperature measurements suggest that the rms amplitude of water motion at a distance of 540 m from the sea floor may be as large as 28 m for periods of the order of a day but they are less than 4 m in the vicinity of the Väisälä frequency. These results for a region of considerable deep water activity do not differ impressively from the results obtained in the presumably quieter deep waters at 12° N, 27° W. Observations at the latter location showed low frequency motions of 20 m amplitude (Pochapsky, 1968).

Temperature profiles

Instrumented floats were allowed to sink to the sea floor and then return to the surface. Meanwhile, they transmitted temperatures and pressures. The temperatures revealed some interesting features as a function of the time of descent, or ascent, but these could not be related directly to the depth because none of the pressure gages operated satisfactorily. Nevertheless, the gages did establish that the velocity of a moving float varied smoothly with the time and so the depth should have varied smoothly. The velocity could be calculated on the basis of the density difference between a float and the surrounding water and the depths obtained on this basis yielded temperature profiles consistent with those obtained with hydrographic casts. It is believed that the depths calculated in this way are correct to within 10 m near the surface or sea floor and to within 50 m at mid-depths. Errors in temperature gradients determined on this basis should not exceed 0.2 percent as a consequence of the uncertainty in depth. The thermal time constant was 5 sec for either the thermistor probes or for the float shell.

As an example, the depth D in meters during descent for a nominal sinking weight was calculated to be

$$D = 1000 + 32.24\ t - 0.0259\ t^2 + 1.38 \times 10^{-5}\ t^3$$

A constant drag coefficient, 0.6, was employed. t is the time in minutes after reaching the 1000-m level. The coefficients were modified in each case to be in accordance with the measured total time of descent. That time was known to an accuracy of a fraction of a minute and was established with the help of echoes from the bottom to determine the time of path reversal.

The absolute accuracy of the temperature readings was better than ± 0.05°C, but the relative accuracy was approximately ± 0.01°C. The least count was ± 0.001°C. At a constant temperature, the standard deviation attributable to overall system noise was 0.0036°C, as suggested by data obtained in the hovering experiment. Consequently, 95 percent of the readings would lie in a band of ± 0.007°C about the true temperatures.

Temperature profiles obtained in this fashion are compared in Fig. 3. The temperature scale applies directly only to the left profile, a. All other curves are shifted in increments of $\frac{1}{2}$°C relative to that scale and individual readings for these are drawn as circles with a scale diameter of 0.015°C. Drop b was started on 14 October at 1800 hr. The others, d, f and h, took place 21, 40 and 49 hr later, respectively.

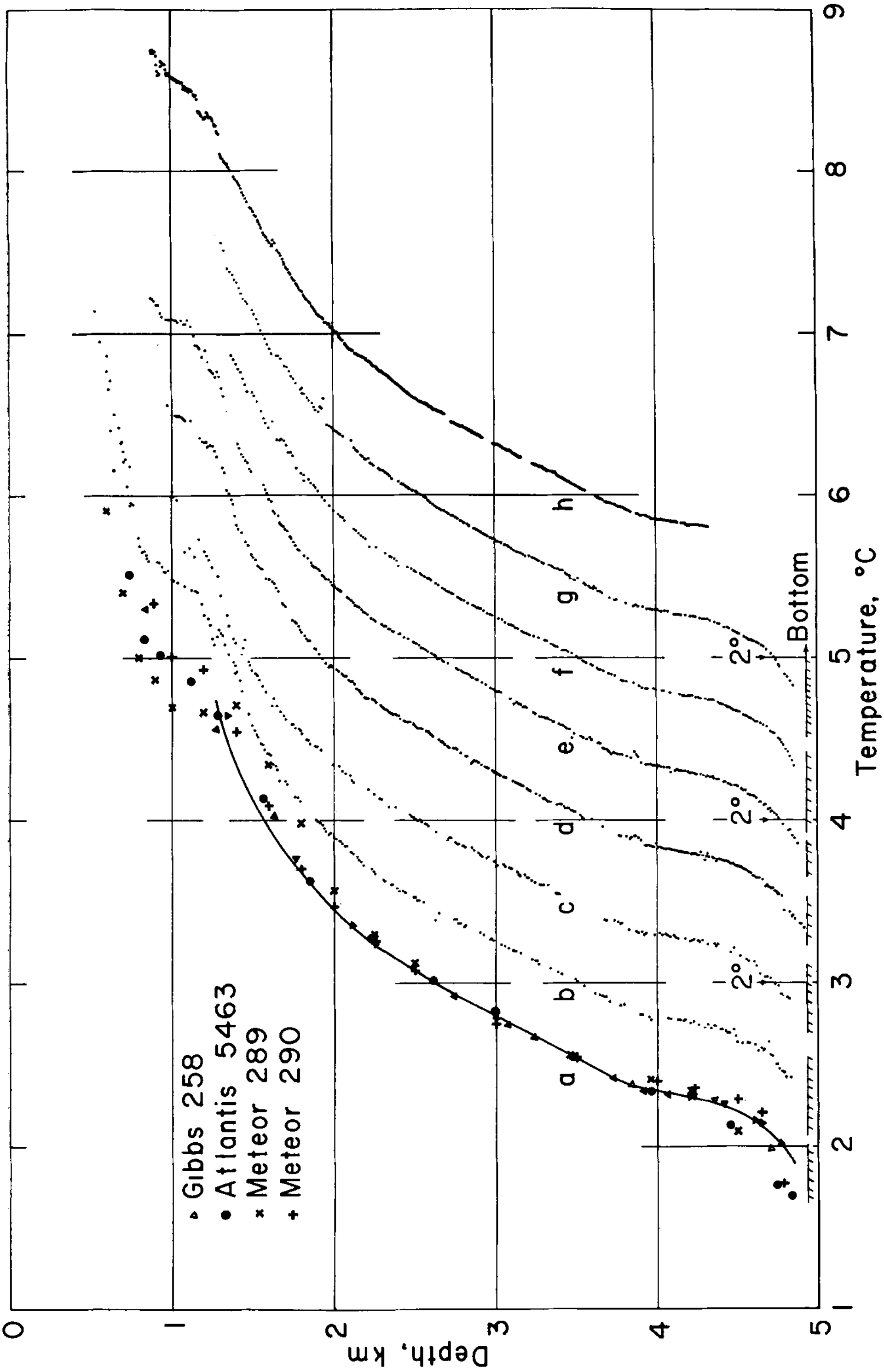

Fig. 3. Temperature profiles. Curve *a* is a reference obtained from the others and is marked with hydrographic thermometer readings. The temperature scale applies directly only to these. The other curves, obtained from sinking and rising floats, are displaced in increments of ½°C.

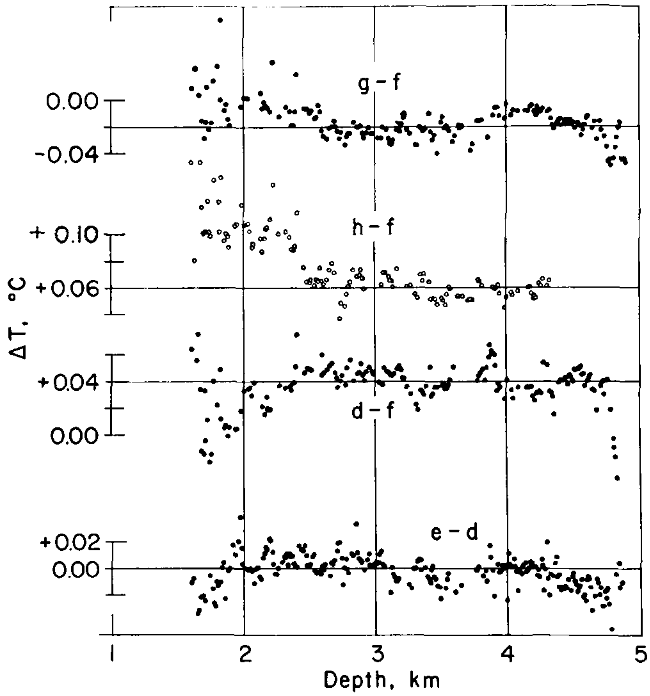

Fig. 4. Temperature differences between some profiles of Fig. 3. as a function of depth.

Profile *a* is a reference curve and represents the average for all drops to which a constant, 0.025°C, was added in order to agree more closely with the hydrographic thermometer measurements. The latter are represented by individual symbols along that curve and illustrate values obtained at the present Gibbs station as well as in earlier work at Atlantis station 5463 (Capurro, 1961) and Meteor stations 289 and 290 (D.A.E. 1932). Except in the deepest and shallowest water, the *Gibbs* thermometer readings agree, to the expected precision of 0.01°C, with the Atlantis results at the nearby location, 10°17′ N, 50°31′ W, occupied 11 years earlier. The present readings are on the cooler side. The Gibbs station lay between the two Meteor stations, 289 at 11°2′ N, 49°33′ W and 290 at 9°7′ N, 50°27′ W. If it is proper to interpolate between these, a definite cooling has taken place since 1927 except near a depth of 3500 m. Substantial differences, measured in tenths of a degree, occur at the different times within the deepest 500 m of water.

Curves *b* and *c* were obtained during the sinking and ascent of an instrument less sensitive than the others used. At one-fifth the usual sensitivity, most of the scatter in these readings can be considered of instrumental origin. It is also probable that the surfacing time was incorrectly recorded; this would explain a temperature slope in the shallower water during ascent which is inconsistent with the other profiles. The high density of readings along profile *h* was obtained for the slowly sinking hovering float. Readings obtained within 150 m of the bottom during the drops suffer in accuracy because of interference from bottom echoes. There was no indication of isothermal conditions to within tens of meters of the sea floor.

Below a depth of 2000 m, temperatures are seen to decrease in a fairly smooth fashion with depth. Measurements which deviate significantly from the trend are best ignored unless the deviations are common to the descent and ascent of a given drop, i.e. to *b* and *c*, *d* and *e*, or *f* and *g*. The most conspicuous anomaly took place near a depth of 3850 m for drop *d*, *e* where temperatures exceeded the trend by as much as 0.05°C.

There are small undulations in the profiles which suggest that there could have been large temperature differences at a given depth among the profiles. That this is not so is illustrated in Fig. 4 where temperature differences are plotted

Table 1. *Temperature relations and stability as a function of depth*

Depth (meters)	Temperature In situ (T, °C)	Temperature Potential (θ, °C)	Salinity[a] (‰)	$d\theta/dz$[b] °C cm^{-1} $\times 10^5$	$d^2\theta/dz^2$ °C cm^{-2} $\times 10^{10}$	E meter^{-1} $\times 10^8$	Stability period (hours)
1600	3.96	3.82	34.980	1.736	2.09	23.6	1.15
1800	3.68	3.52	34.975	1.318	1.73	16.8	1.36
2000	3.455	3.27	34.970	1.046	1.04	13.1	1.54
2200	3.275	3.18	34.965	0.900	0.84	11.1	1.67
2400	3.14	2.92	34.960	0.711	0.53	8.72	1.88
2600	3.015	2.79	34.953	0.687	0.15	8.06	1.96
2800	2.905	2.65	34.946	0.650	0.17	7.04	2.10
3000	2.80	2.52	34.938	0.618	0.09	6.96	2.11
3100	2.75	2.46	34.934	0.616	0.03	6.96	2.11
3200	2.695	2.40	34.930	0.613	0.15	7.14	2.09
3300	2.645	2.34	34.926	0.586	0.02	6.71	2.15
3400	2.595	2.28	34.922	0.609	−0 23	7 21	2.08
3500	2.540	2.21	34.918	0.633	−0.07	7.77	2.00
3600	2.485	2.15	34.913	0.623	0.40	7.81	1.99
3700	2.43	2.08	34.911	0.553	0.53	6.77	2.14
3800	2.39	2.03	34.908	0.518	0.54	6.27	2.23
3900	2.36	1.99	34.904	0.444	1.14	5.09	2.47
4000	2.34	1.96	34.900	0.290	0.83	2.81	3.32
4100	2.32	1.92	34.898	0.278	−0.26	2.88	3.29
4200	2.30	1.89	34.895	0.342	−0.50	4.07	2.76
4300	2.28	1.85	34.893	0.379	−0.70	4.79	2.57
4400	2.25	1.81	34.890	0.483	−1.11	5.89	2.30
4500	2.21	1.76	34.885	0.601	−1.51	6.82	2.13
4600	2.16	1.70	34.879	0.784	−2.10	9.38	1.82
4700	2.08	1.60	34.870	1.021	−3.62	11.41	1.65
4800	1.97	1.48	34.857	1.509		17.87	1.32

[a] Based on Atlantis Station 5463. The T–S relationship was used below 4000 m.
[b] z, distance upward.

as a function of depth. For the uppermost set of points, values obtained at different depths during the ascent f were subtracted from the corresponding readings, necessarily interpolated to obtain the same depths, during descent g. The other plots were obtained in a similar fashion. When consideration is given to the fact that scatter in the measuring system places 95 percent of the readings at a constant temperature within a band of ± 0.01°C, few significant trends are apparent in the range 2000 to 4750 m. The high scatter of values in depths shallower than 2000 m suggests appreciable turbulent fluctuations. It was hoped that this procedure would reveal internal wave modes, but it is now clear that an order of magnitude improvement in temperature reliability is needed.

The temperature measurements at small depth intervals are summarized in Table 1 together with salinity values obtained by interpolating the data of Atlantis station 5463. Our own salinity determinations tended to agree with the Atlantis results but showed more scatter. Adiabatic temperatures and their first and second derivatives with respect to height, as well as stabilities and local stability frequencies, are included in the table. Each gradient is the average of individual gradients found graphically for each profile. Individual gradients varied about the average by approximately ± 10 percent but the variations of second derivatives were ten-fold larger.

The foot of each temperature curve has a shape which suggests that there may be a simple analytic relationship between temperature and depth in the deepest waters. Neither a simple power nor an exponential relationship could be found. On the other hand, the logarithm of the stability is linearly related to depth within the precision of our measurements as shown by the

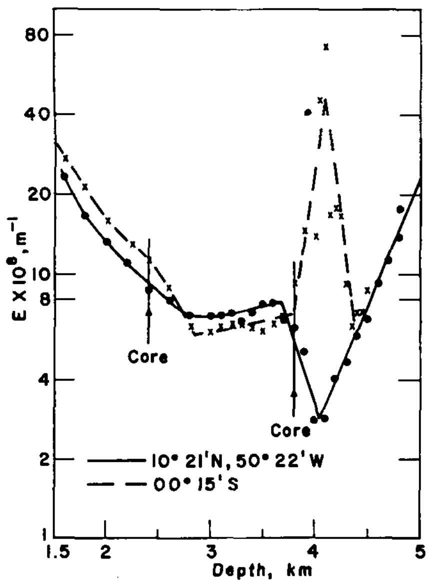

Fig. 5. Stability as a function of depth at two locations.

solid curve of Fig. 5. The water core positions noted in that figure were obtained from clearly defined breaks in the T–S curves.

The logarithmic relationship may have a more general validity than a fortuitous one at the Gibbs station. To explore this possibility, use was made of the extensive data obtained on Crawford Cruise 22 (Capurro, 1961). Five stations on a track along 00°15′ S between 31°W and 31°27′ W had impressively similar profiles and stabilities were calculated on the basis of an average profile. Nevertheless, the depth intervals were large for present purposes and errors were particularly large in the region of steep gradients at depths between 4000 and 4200 m as well as near the bottom. The results are not precise enough to judge the correctness of a logarithmic relationship but they are included in Fig. 5 as a dashed line to show that such may be possible.

In moving from near the equator to 10° N, the layer in which the stability increases with depth descends toward the sea floor and leaves behind a region of low stability. At any fixed point within the former layer, the sign of $d^2\theta/dz^2$ requires a substantial loss of heat downwards for a constant diffusion coefficient of 22 cm^2 sec^{-1} (Wüst. 1957). Heat is also lost to warm the waters downstream. This suggests that heat must be convected from above, almost as though the bottom current aspirates water into itself. It would be of interest to find the magnitude of the horizontal current as the bottom is approached. The present value at 540 m above the floor may be substantially different from the values within the gravity current nearer the bottom. Some inconclusive measurements of our own suggest the possibilities of velocities of a few cm sec^{-1} a meter off the floor and of more than 20 cm sec^{-1} at distances of 10 and 100 m. In any event, the heat balance is strongly dependent on convective terms and thorough measurements of the velocity structure are needed before conclusions can be reached as to the magnitudes of the diffusion coefficients in this interesting region.

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СКОРОСТЬ И ТЕМПЕРАТУРНАЯ УСТОЙЧИВОСТЬ НАД АНТАРКТИЧЕСКИМ ПРИДОННЫМ ТЕЧЕНИЕМ НА 10° с. ш.

Для изучения некоторых характеристик температуры и потоков, связанных с Антарктическим придонным течением, на 10° с. ш. были использованы нейтрально плавающие затопленные буи. На расстоянии 540 м от дня был найден поток со средней за 5 дней скоростью 4,6 ± 0,5 см/сек в направлении 325°. Эти и другие величины, полученные в процессе опускания буя, находятся в хорошем согласии с предсказаниями Г. Вюрста. Вертикальные движения буя не превосходили в среднеквадратичном смысле 28 м, что сравнимо с величинами, наблюдавшимися в более спокойных районах.

Температурные профили были получены с буев, которые опускались на дно, а затем возвращались на поверхность. Первая и вторая производные температуры по глубине измерялись через 100-метровые интервалы. Была определена величина устойчивости E и было найдено, что $\log E$ вблизи дна меняется с глубиной линейно.