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REVIEW PAPERS:

I. "CURRENT - EFFECT" ON SEA LEVEL

II. OCEANIC EDDIES

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I. "CURRENT-EFFECT" ON SEA LEVEL

INTRODUCTION

There is now a considerable body of literature on changes in sea level that are not astronomical in origin. Some of these changes have been ascribed to the effects of variation in currents near the sea level stations. The aim here is to review what is known of this "current-effect" on sea level.

It is not easy to define what is meant by the current effect, nor to separate it from the effects of other variables, especially wind stress. It is known however that some major current systems, such as the Kuroshio and the Gulf Stream, are variable in both position and time, and that their variations do not seem to be related to local weather. We are mainly interested in the effects of such variations on sea level, especially at coastal stations. Thus we wish to ignore the regular seasonal changes in sea level (although admitting that these may be associated partly with seasonal current or circulation changes), and also ignore the changes in sea level that are directly correlated with local weather. On the other hand, sea level changes associated with changes in density of the water column ("steric" sea level changes) are of interest, to the extent that they are due to variations in currents, rather than to local heating or cooling, precipitation or run-off.

REVIEW

The following areas will be considered in turn: Pacific coast of Japan, east and west coasts of U.S.A., south-east coast of Australia.

Pacific Coast of Japan

Suggestions of a current effect on sea level on the Pacific coast of Japan were made as early as 1927 (Nomitsu and Okamoto 1927). Miyazaki (1955) noted that the amplitudes of annual (Sa) and semi-annual (Ssa) sea level variations at particular stations varied widely from year to year. In the light of later work, such variations can probably be ascribed to the current system.

Moriyasu (1958) studied monthly mean sea levels (adjusted to a fixed value of atmospheric pressure) at four stations around Kii peninsula (Fig. 1). The most interesting result was the large fluctuation in sea level difference between Kushimoto and Uragami, which are only 15 km apart. This difference had a range of more than 30 cm for the period 1951-53, but the range decreased to only about 2 or 3 cm from late 1953 to the end of 1955. The difference in sea level between the two stations was shown to be fairly well related to the position of the axis of the Kuroshio. When the axis was more than about 50 km offshore, the sea level difference was small and relatively constant. The large and more variable differences appeared when the Kuroshio axis was less than 50 km offshore. (The 200 m isobath is only about 15 km offshore in this area.)

In the same paper, monthly mean sea levels at Kushimoto were shown to be closely related to dynamic height anomalies relative to 800 decibars at a station about 20 km offshore. The rangesof dynamic height and monthly mean sea level were between 40 and 50 cm, and the departures from a unit-slope line (which would represent perfect isostatic adjustment) were within ±12 cm. Better agreement would not be expected, since monthly mean sea levels were compared against dynamic heights computed from only one station per month.

In a subsequent paper, Moriyasu (1960) examined sea level data from additional stations between Aburatsu and Maisaka (Fig. 1). He presented two more graphs showing fairly close agreement between monthly mean sea levels at these two stations and dynamic height anomalies about 100 km offshore from each station. Monthly sea level anomalies* were computed for each month, 1951-56 for nine stations, and the standard deviations of the anomalies discussed in relation to season, year and position along the coast. Correlations between sea level anomalies at different stations were also computed.

Interpretation of the standard deviations and correlations was difficult, but the main conclusion seemed to be that the sea level anomalies were greatest on those parts of the coast (between Shionomisaki and slightly east of Maisaka, approximately) where the oceanographic conditions were known to fluctuate noticeably.

In 1961, Moriyasu extended the graph of the sea level difference between Kushimoto and Uragami. The period of large and variable difference in sea level which had begun at the end of 1955 continued until May 1959. Its ending coincided with a sudden movement of the Kuroshio away from the coast in this region. This behaviour, both of sea level difference and of the current, was very similar to that found in July 1953. The sea level difference between these two nearby stations, and its relation to the known movements of the Kuroshio, are the clearest evidence to date of a connection between coastal sea levels and offshore currents.

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* "Anomaly" for e.g. January 1953 = (Monthly mean sea level for January 1953) - (mean of all available January monthly means).

In the same paper, Moriyasu explains the sometimes large difference in sea level between stations only 15 km apart as due to the Bernouilli effect associated with a strong southward current which diverges rapidly from the coast between the two stations. Dominance of the Bernouilli effect implies a balance between

$$v\ \partial v / \partial y \quad \text{and} \quad -g\ \partial \zeta / \partial y$$

in the horizontal equation of motion, where ζ is sea level and v the longshore (y -axis) current velocity. Numerical values chosen to suit data from an admittedly small number of stations (five) off the Kii peninsula suggest 14 cm change in 15 km, for a current of maximum value 4 kt. The sign seems to be correct: the observed difference Kushimoto minus Uragami becomes more positive in periods of high variability, compared to its value in quiescent periods, as would be expected on the Bernouilli effect hypothesis.

On the other hand, although it is of the right sign and order of magnitude, the Bernouilli lowering appears too small to account entirely for the > 30 cm range in sea level difference between Kushimoto and Uragami. Also, this mechanism on its own would imply a large variance in sea level at Uragami, relative to Kushimoto, but the standard deviations in the 1960 paper are almost the same at the two stations. The question of isostasy was not discussed.

Moriyasu considered the connection between actual monthly mean sea levels (as well as the difference discussed above) and the position of the Kuroshio. His conclusions (1960) were that at Aburatsu, sea level rises when the Kuroshio moves inshore, whereas the opposite is true at Maisaka. The difference in behaviour appears to be connected with the appearance of a "cold water region" (counter-clockwise eddy) off Maisaka, when the Kuroshio moves offshore; this feature is not found off Aburatsu. Schematically, the variation in sea level along sections normal to the coast off Maisaka and Aburatsu appears to be as shown in Figure 2, where the full lines apply when the Kuroshio is further offshore, and the dashed lines when it is nearer the shore.

Shoji (1955, 1957) gives daily mean sea levels (25 hour mean, adjusted to fixed atmospheric pressure) for an island in the path of the Kuroshio (Hachijo-jima, Fig. 1), and also for a number of coastal stations, for 1954 and 1955. He also shows graphs of daily mean sea level against dynamic height, relative to 800 decibars, for Hachijo-jima and Shionomisaki. The range of values at Hachijo-jima is about 100 cm, and the conditions are isostatic to within about ±10 cm. (There is slight evidence that a deeper reference level would have given a better fit.) Off Shionomisaki the range is about 50 cm, and again conditions are isostatic, this time to within ±6 cm, except for one observation, 12 cm from the line.

The daily sea levels are presented graphically, and discussed qualitatively. The main conclusions are:

(i) The sea level graphs are fairly smooth, justifying the neglect of wind effects.

(ii) Sea level variations at the island station (Hachijo-jima) are very large compared to those at coastal stations.

(iii) Sea level changes at Hachijo-jima sometimes show quasi-periodic changes, with a period of about two months.

(iv) Sea level changes at the coastal stations (except Onahama) appear to be highly correlated. They also appear to be negatively correlated with changes at Hachijo-jima.

(v) At Onahama, sea level variations are smaller and more regular than at other coastal stations. This is thought to be due to the different current system (Oyashio current) off Onahama.

In a later paper (Shoji 1961) daily sea level graphs are given for additional stations. These results support the hypothesis that relatively large sea level changes are found on the parts of the Japanese coast bordered by the Kuroshio. The Bernouilli effect is again discussed briefly.

A large, relatively local change of sea level of 30 cm in five days was mentioned. Although a typhoon struck Japan at this time, Shoji considered this large change was not a simple wind effect. On the graphs presented, this large change appears as part of several cycles of varying sea level, lasting for about five weeks.

Shoji refers to a tendency for disturbances (individual peaks or troughs in the sea level record) to travel westward along the south coast of Japan. This has been confirmed by Isozaki (1968), who suggested they are not due to continental shelf waves, since the shelf is very narrow and the bottom topography complex.

To summarise, a connection between coastal sea levels and variations in the Kuroshio near the southern coast of Japan seems to be clearly established. On present evidence, the connection seems to be baroclinic rather than barotropic, since a close correlation has been shown between coastal sea levels and nearby dynamic heights. But the observations have not yet been fitted in with theory. The simple Bernouilli effect in a barotropic model ocean does not offer a satisfactory explanation.

Time scales between several days and at least several years seem to be involved, and there is some evidence of long period modulation of the shorter-period effects, as well as a direct long-period variation. Space scales from at least 15 km upwards are indicated.

East Coast of U.S.A.

In spite of the many years of study of the Florida Current and Gulf Stream, their connection with observed sea level variations is not clear. The following factors make it more difficult to study such a relation off the east coast of U.S.A. than off Japan. (i) Off U.S.A. the continental shelf is wider, so that the effects of local winds are expected to be greater. (ii) The Florida Current is constrained by bottom topography, so that only fluctuations in flow rate can be expected to show up in sea levels. (iii) The Gulf Stream appears to follow a relatively fixed path from Florida to Cape Hatteras (see locality map, Fig. 3).

Montgomery (1937) compared sea levels at Charleston and Miami with dynamic heights off Chesapeake Bay. Agreement was reasonable, but only five dynamic heights were used, and there were large separations (420 and 720 miles, respectively) between tide station and oceanographic station. In a later paper (1941b) Montgomery compared sea levels at Key West with dynamic heights at four stations on the left hand (north) side of the Florida Current, and within 60 miles of Key West. Agreement was "perhaps slight". The only extensive study of the connection between sea level and dynamic height was at Bermuda (Shaw and Donn 1964). The agreement here was good, but this is not particularly relevant to the present discussion.

Cross-stream differences of sea level have been studied by several workers. Hela (1952) compared the difference in monthly mean sea levels between Cat Cay and Miami with long-term average surface currents. There was very good agreement in shape between the two curves. The sea level curve was the mean over three years. Plotting the sea level differences for each year separately shows surprisingly good agreement from one year to the next, except for the one value (November 1940). This agreement, if supported by further data, would imply a degree of seasonal regularity that is in marked contrast with the Kuroshio. In the same paper, Hela discusses the connection between sea level difference and actual sea level at Miami. It appears that, on average, a 10 cm rise in sea level at Miami corresponds to a simultaneous 3.4 cm rise at Cat Cay. Being in the same direction, these changes do not fit in with the idea that they are "solely due to fluctuations in the Florida Current". The simultaneous rise at stations on both sides of the Florida Current is contrary to Shoji's qualitative inference from sea levels at Hachijojima and at nearby coastal stations, but the difference in conditions should be kept in mind. The Florida Current is confined to flow between Miami and Cat Cay, whereas the Kuroshio is free to move onshore or offshore.

Montgomery (1937, 1941a) and Hela (1957) discuss the longitudinal slope of the Florida Current in relation to sea level. Montgomery (1941a) concludes that "there is no justification for using the difference in sea level from Key West to Miami as an indication of the strength of the Florida Current". The effect of local winds on sea level at the two stations was considered large enough to mask changes in longitudinal slope.

O'Hare et al. (1954) compared monthly mean sea level at Miami with monthly mean surface currents in the Gulf Stream east of Florida, for the period May 1950 - April 1951. There is fair agreement. The interpretation is that sea level in the Sargasso Sea remains relatively constant, so that an increase in strength of the Gulf Stream should be accompanied by a lowering of sea level at the coast. The scales chosen for their Figure 9 correspond to a Gulf Stream effective width of about 50 km, which seems reasonable.

Fuglister (1951) also compared sea levels at Miami and Charleston with estimates of current derived from ship's drift. He found "only fair" agreement.

Stommel (1953) examined differences between long-term average monthly mean sea levels at Cat Cay, Key West, Havana and Miami Beach. The observations were discussed briefly in relation to a two-layer model with axial acceleration and with conservation of vorticity. Stommel claimed that vertical shrinking of the upper moving layer, in the direction of flow, produces an anticyclonic shear, and "magnifies the transverse geostrophic slope of the free surface". It is not clear if this magnified slope differs from what one would infer from the greater surface velocities in the narrower and shallower Miami-Cat Cay Section.

Mysak and Hamon (1969) showed that the spectrums of sea level at two North Carolina stations showed broad peaks centred around a period of 15 days. These did not seem to be connected with local weather. It was suggested that they might represent an effect of variations in the Gulf Stream, but no comparison with the speed or position of the Gulf Stream could be made.

West Coast of U.S.A.

Off the west coast of U.S.A., currents are relatively weak, so that a connection between current and sea level will be hard to establish. An interesting start is being made at Oregon, where continuous records of currents on the shelf and slope have been made. In a recent thesis (Collins 1967), the connection between sea level and current was examined by regression, and significant regression coefficients were found. But the currents in this area are also related to wind stress, which in turn has a direct effect on sea level, so that a multivariate analysis will be necessary before the direct connection between current and sea level can be stated confidently.

Sturges (1967) gives clear evidence that sea level is isostatic (to within about ±8 cm) off San Diego (Cal.) and Neah Bay (Oregon). Saur (1962) reported on monthly mean sea levels and atmospheric pressures at six stations in the eastern North Pacific. He assumed that residual sea level (after allowing for atmospheric pressure) represents variation in steric level, and that this, in turn, reflects the variability in the ocean current along the coast. With these assumptions, he claims that the year-to-year difference (in current) in the same month may be as great as the mean seasonal change in current. No observations of steric sea level or of currents were used in this paper.

South-East Australia

The circulation off the south-east coast of Australia is strong, complex and variable, at least as far east as Lord Howe Island (Hamon 1965). The mean sea level variance at Lord Howe Island is very high (Hamon 1968) especially for periods between a few months and one year. It is almost certain that this is mainly an isostatic effect, but there are too few stations near the island for this to be demonstrated at present.

At coastal stations (Sydney 34°S., Coffs Harbour 30°S.) and for periods in the range 20 to 50 days, the spectrums of mean sea level show about three times the variance of the corresponding atmospheric pressure spectrums, assuming hydrostatic response. Wind stresses are expected to be small at these periods, and in any case sea level changes due to wind are small on this coast due to the narrow shelf. It appears very likely that the large sea level variances in this range of periods is connected with the strong, variable offshore circulation.

Some preliminary work has been done on sea level differences between coastal stations. Figure 4 shows differences in four day mean sea levels between Sydney and Port Kembla. The separation between these two stations is 80 km. The figure shows a large range (35 cm) and that changes of the order of 10 cm in two weeks are not uncommon. There was a curious quasi-permanent change in average level towards the end of 1963, which is reminiscent of the sudden changes in sea level difference between Uragami and Kushimoto, noted above. In the present case, however, cruise data covering the period September 1963 to September 1964 (Hamon 1965) do not show any obvious changes in current regime that could account for the sea level difference change at the end of 1963. Even if this sudden change is disregarded (as perhaps due to instrumental defects), the remaining sea level difference variations are considered too large to be due to weather effects or continental shelf waves, and are therefore presumed to be connected with the offshore circulation. But again, comparison with cruise data for the period shows no obvious connection.

The connection between daily mean sea level and dynamic height anomaly just beyond the edge of the shelf near tide stations has been studied for Sydney (Hamon 1957) and Eden (Fig. 5, also Hamon and Stacey 1960). In both cases, there seems to be much less agreement than has been found in other parts of the world. We conclude tentatively that conditions in the deep water off south-east Australia are not isostatic. We would expect appreciable pressure changes on the bottom in deep water.

THEORY

Very little has been done to provide a theoretical background against which the significance of observed sea level changes, related to circulation, can be assessed. As discussed above, the simple Bernouilli effect seems to be inadequate.

Papers on the theory of ocean circulation generally give scant attention to the problem. Most theoretical papers assume steady conditions, whereas the observational evidence discussed above relates to variations with time. The results of steady-state theories can be checked against observation only by comparison with the difference between long-term mean sea levels at several coastal stations, and the results of precise levelling. As has been shown recently for the U.S. Pacific and Atlantic coasts (Sturges 1967, 1968) there are serious difficulties in levelling with enough precision for this, although one might expect better results over short distances along coasts with strong off-shore circulation.

Intuitively, one feels that, at least for barotropic ocean models, the coastal sea level variations would depend to first order on friction. Since friction cannot be modelled realistically, there is perhaps less chance of accounting for observed coastal sea levels than for other features, such as volume transport.

More progress in this field might come from computing coastal sea levels that are dynamically consistent with observed features of the off-shore circulation, rather than from attempts to solve problems of circulation in closed model basins. In this approach, the intervening continental shelf could probably be allowed for realistically, although the problem of friction remains.

CONCLUSION

There seems to be no doubt that ocean currents affect coastal sea levels to a measurable extent, at least in areas of strong circulation. It is also clear that we have not yet really begun to understand the underlying mechanism.

The greatest barrier to progress is the lack of an appropriate theoretical framework. This is needed to help interpret existing data, and to suggest what new data would be the most useful. Lacking guidance from theory, one can only speculate on the lines that field work and analysis should take.

I think there are three avenues for development or improvement:

1. Greater use of recording current meters, both on the continental shelf and on the continental slope, and eventually in abyssal depths.
2. Use of bottom-mounted pressure recorders to check on the isostatic hypothesis.
3. Wider use of the more sophisticated analysis techniques that are now available, for example
    - 3.1 Multiple time series analysis techniques for better elimination of the effects of local weather (see e.g., Groves and Hannan 1968; Cartwright 1968).
    - 3.2 Appropriate numerical filtering to define more precisely the parts of the spectrum being studied. In this connection, it seems unlikely that the monthly mean sea levels now being published regularly by the Permanent Service for Mean Sea Level will be regarded as satisfactory, since much of the sea level activity due to circulation takes place at periods less than two months. Any serious work in future should start with daily or semi-daily sea levels.

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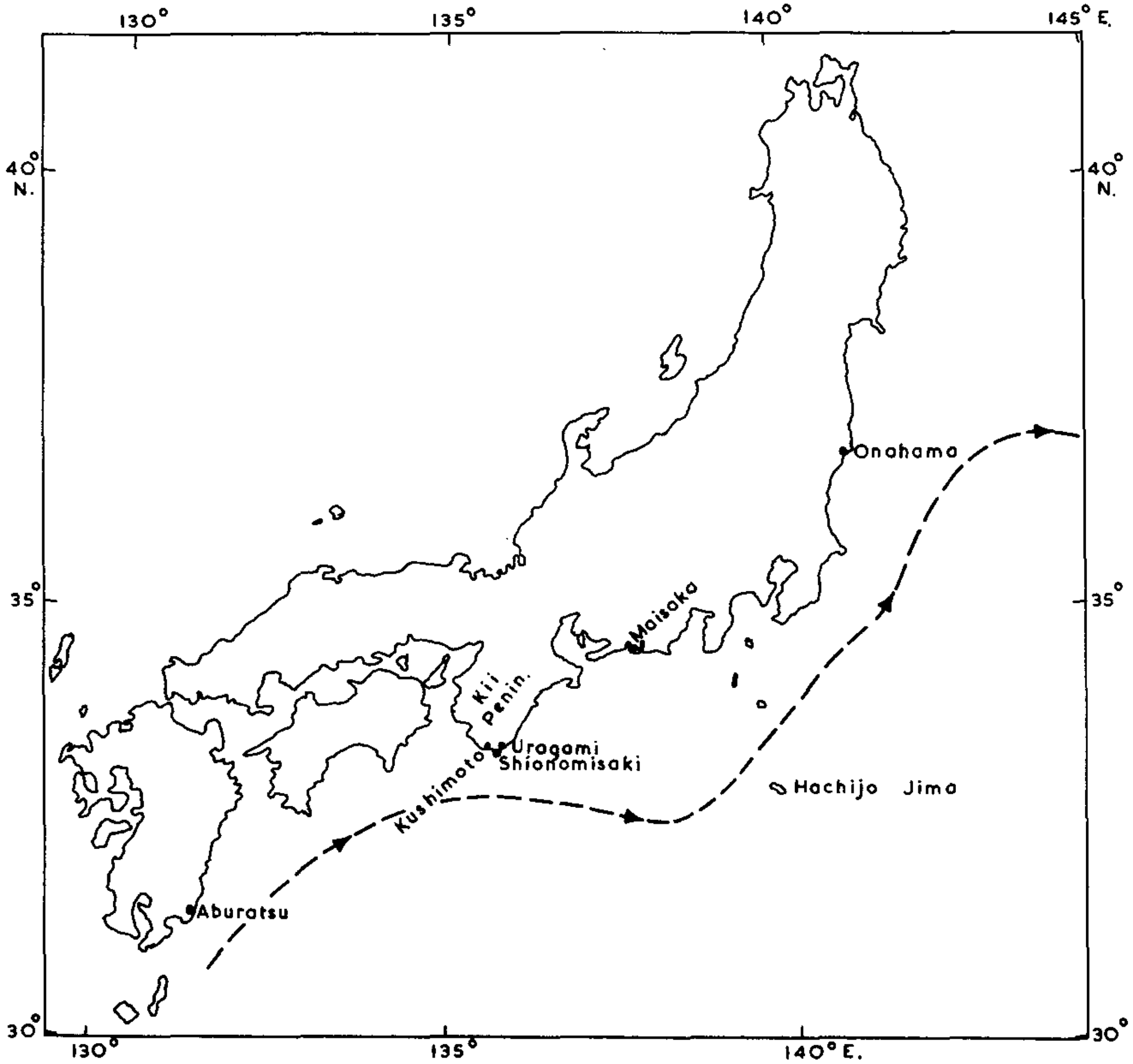

Fig. 1.- Map of Japan, showing tide gauge stations. The dashed line shows the mean position of the Kuroshio axis (after Masuzawa, 1960).

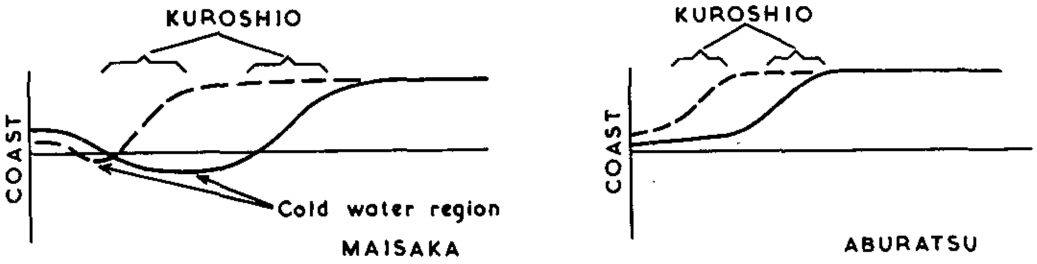

Fig. 2.- Variation of sea level along lines normal to the coast of Aburatsu and Maisaka, Japan. (Schematic)

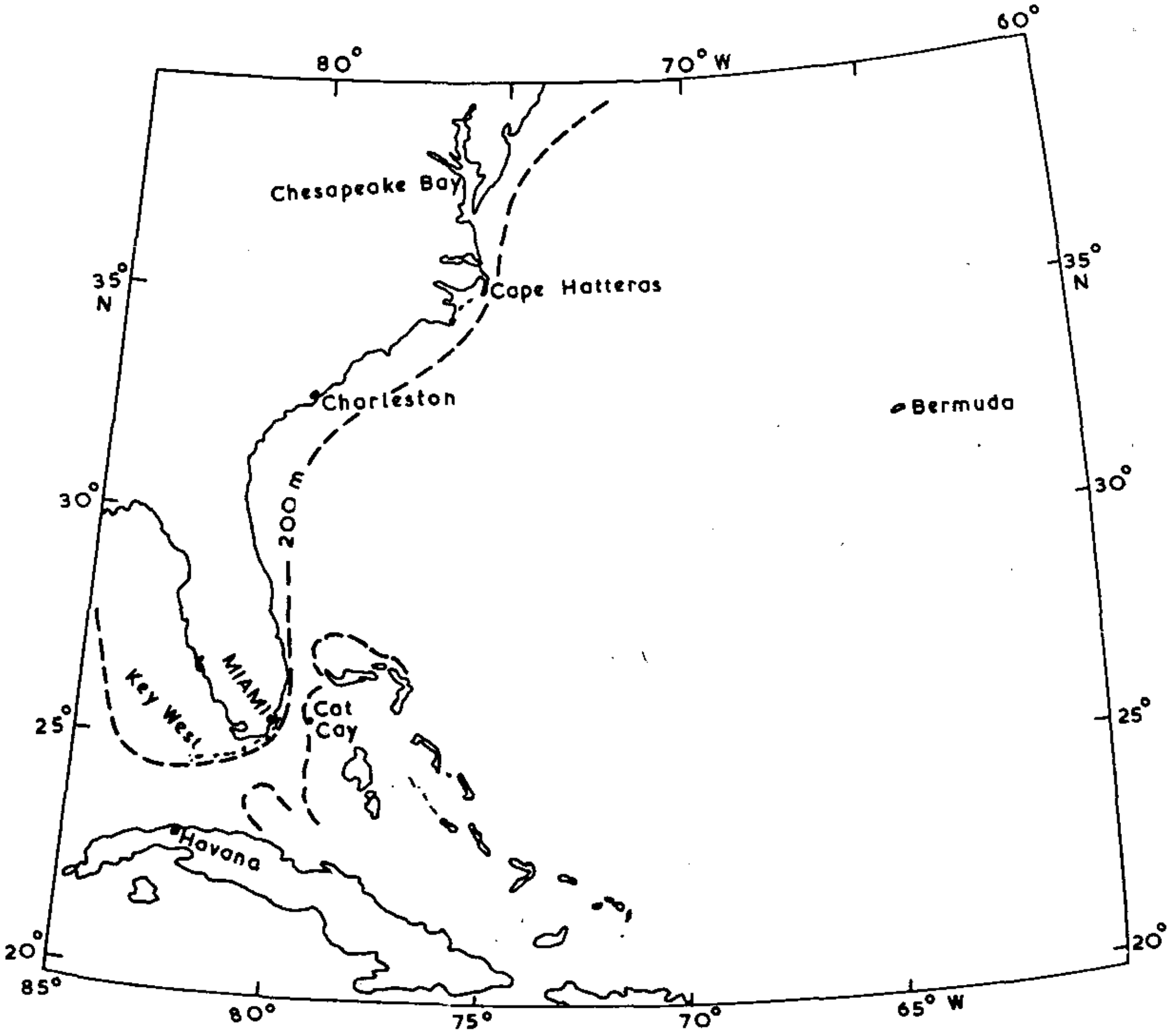

Fig. 3.- Locality Map - Western Atlantic Ocean.

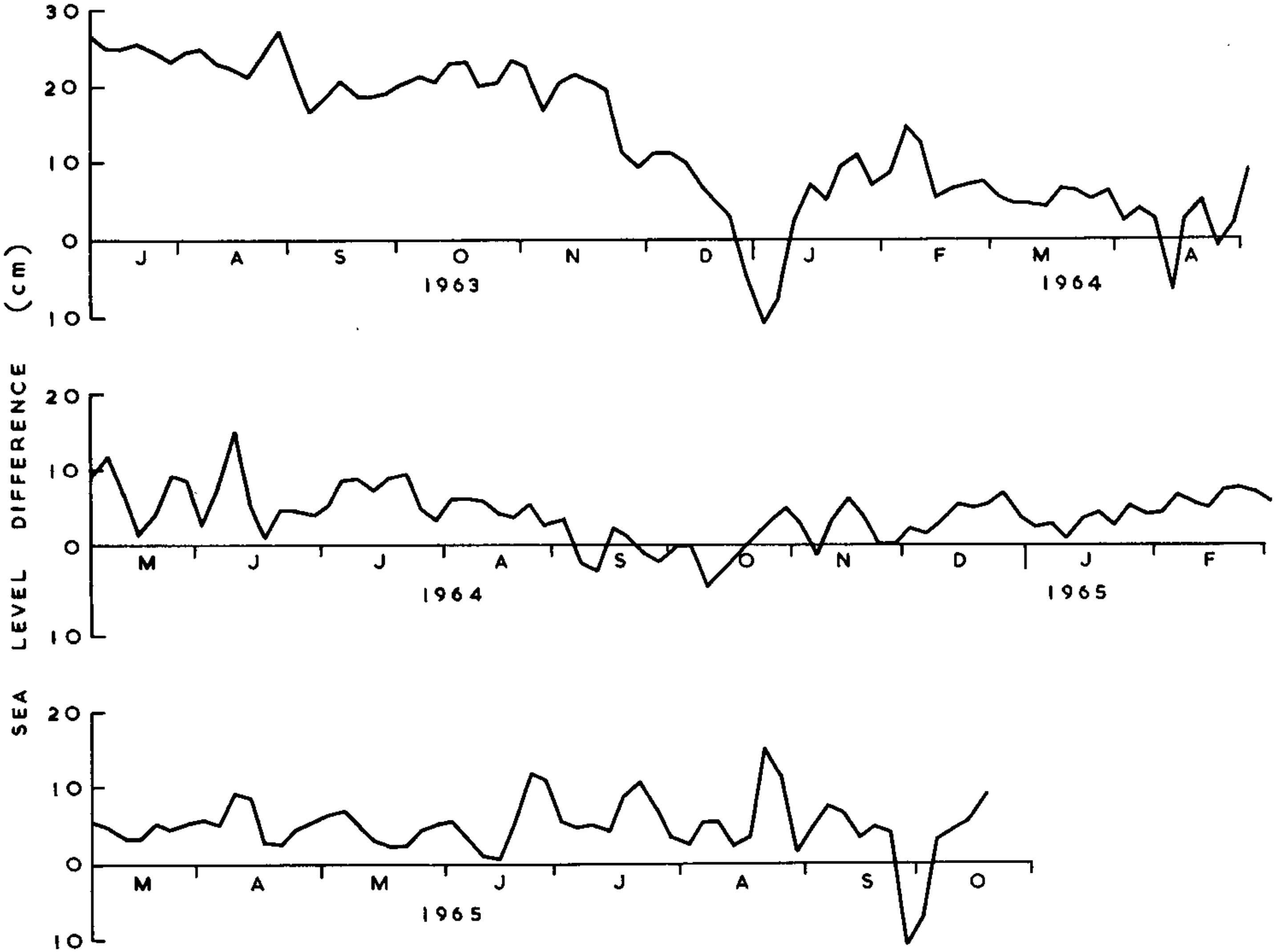

Fig. 4.- Four day mean sea level differences, (Sydney)-(Pt. Kembla), July 1963 - October 1965.
(No adjustment for atmospheric pressure).

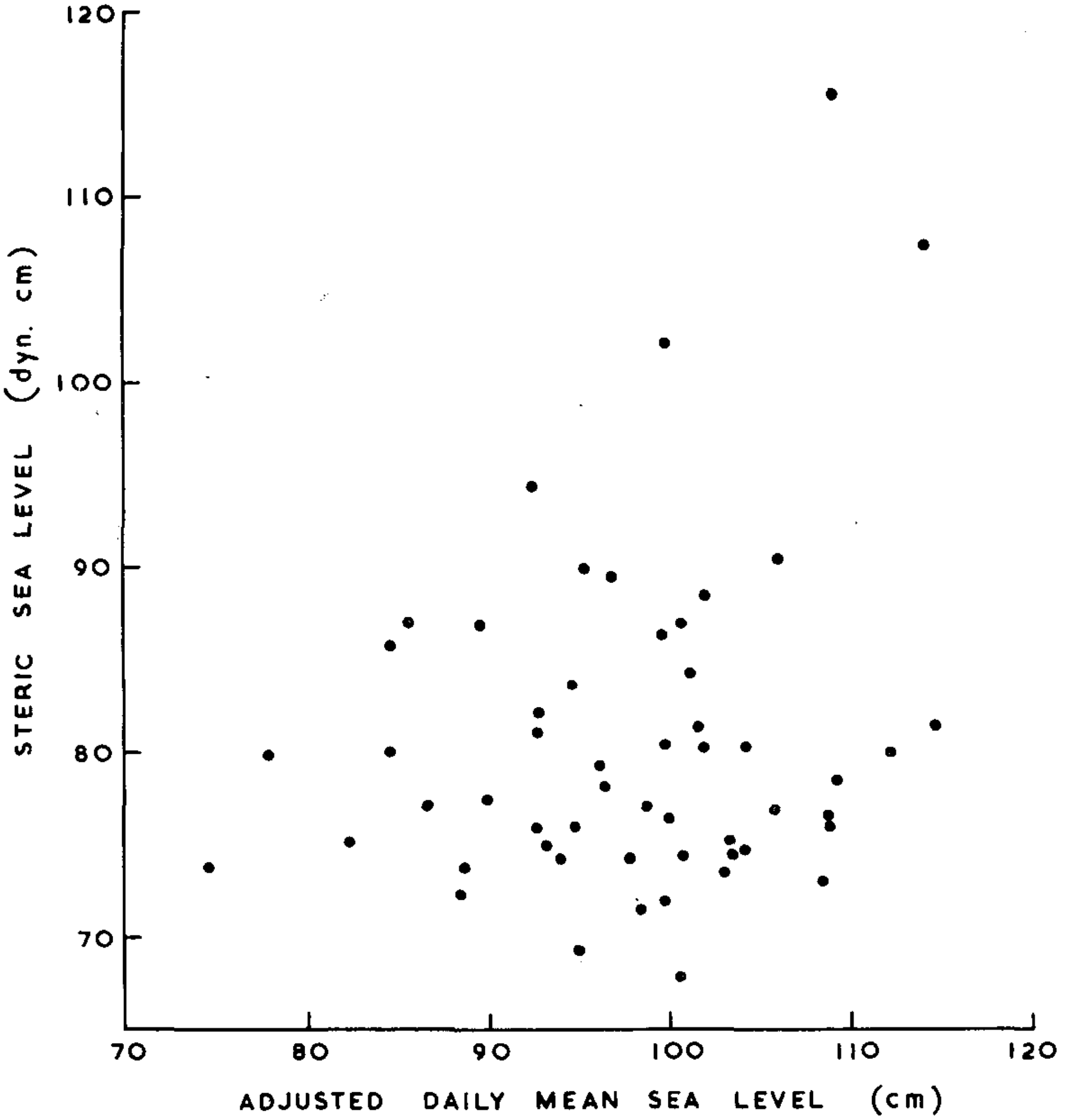

Fig. 5.- Steric sea level, relative to 500 decibars, versus daily mean sea level (adjusted to fixed atmospheric pressure) at Eden (37°04'S., 149°59'E.). The steric sea levels are from a station on latitude 37°04'S., but at varying longitudes in the range 150°21'E. to 150°29'E.

II. OCEANIC EDDIES

INTRODUCTION

The available literature on oceanic eddies is reviewed. In selecting items, I have had in mind particularly the large-scale (~100-500 km diameter) geostrophic eddies, such as we have frequently observed off east Australia. Eddies or gyres with dimensions approaching those of the larger ocean basins (~5000 km) are not considered. The main purpose in compiling the review was to present an annotated bibliography for the benefit of fluid dynamicists who might not be familiar with, or have ready access to, the oceanographic literature.

Terminology

Most workers have used the terms "eddy" or "vortex", but at Woods Hole there is a tendency to refer to large-scale eddy-like features of the Gulf Stream as "current rings". The change is intended to emphasise the difference in structure. The word "eddy" suggests a vortex, with maximum velocity near the centre (e.g. Rankine vortex), but the observed features in the ocean have maximum velocities relatively further away, and in the case of anticyclonic features, have maximum horizontal shear near the outside.

REVIEW

A small "low" off California (Reid et al. 1963). This feature was examined with drogues and by the usual water-bottle technique, with sampling to only about 600 m. A summary of the main features observed is as follows:

Diameter	90 km
Distance of centre from shore	170 km
Distance of centre from 100 fm line	150 km
Surface currents	21-36 cm/sec
Time for drogue to make one circuit	88 hr near centre 134 hr near edge

The current seemed to be confined to the upper 125 m. There was no marked horizontal variation in properties (temperature, salinity, density, thickness) of the mixed layer, which was about 45 m thick. The structural feature leading to the low dynamic height at the centre of the eddy was "a steepening of the vertical gradients in the pycnocline, so that the temperature, salinity and density found between 80 and 125 m at adjacent stations were found between about 50 and 100 m at the centre of the eddy". The eddy motion was geostrophically balanced.

Other small cyclonic eddies were observed in the same area, but were not studied in detail.

McEwan (1948) gives examples of 100 km diameter eddies off California in 1940 and 1941. These eddies start in spring, and show a decrease in velocity and increase in size in a period of about two months. Angular momentum was approximately conserved. The kinetic energy was found to decrease according to a formula of the dissipation of energy caused by lateral eddy viscosity. Observations were in general agreement with a mathematical model, based on the diffusion of vorticity, initially taken to be constant over a circle of appropriate area. The angular momentum per unit thickness was of the order 5×10^{21} c.g.s. units (sum of the values at 0 and 50 m depth). The use of angular momentum and kinetic energy per unit depth, here and in the papers by Ichiye, seems of doubtful validity, but is probably consistent with the assumption of only horizontal flow (McEwan, p. 205).

Gulf Stream Eddies

Most of those observed have been cyclonic, but it seems that this might be simply due to most cruise tracks favouring the right-hand side of the stream.

A section across a cyclonic eddy was given by Iselin (1940) (see Newton 1961, Fig. 4). This is a deep structure (down to at least 1200 m depth). The overall diameter is ~190 km, and the diameter to current maxima about 100 km. Once again, there is no marked structure (at least in the temperature section) above about 70 m depth. A similar section with temperature structure discernible this time down to 3500 m, is shown by Fuglister (1963, p. 297).

When I visited Woods Hole in July 1968, Fuglister told me he had estimated about 14 "current rings" per year might form between Cape Hatteras and Grand Banks. When first formed, a current ring has a horizontal surface temperature structure of 8-10 degC. This disappears first (? six months), but the deeper structure remains for up to 18 months.

Newton (1961) discussed a cold Gulf Stream eddy, especially in its formation stages, and compared it with similar features in the atmosphere. He claims that the forming eddy increases in size by flow of water along the cold-water "tongue" just before separation. His Figure 18 shows that the horizontal area of the eddy increases with depth, down to at least 1000 m, and also shows the vertical distribution of heat deficit.

Warren (1967) looked at dynamical aspects that might determine the translation of an eddy. His model is essentially baroclinic, but the velocity at the bottom must be non-zero (p. 507). Growth or decay of an eddy is ignored, as are lateral boundaries, but bottom topography is dealt with. Only cyclonic eddies are considered. The results are a tendency to westward translation due to the beta-effect, and translation (along slope (?)) due to the topographic effect. In both cases, translation velocity seems to be proportional to the square of the radius.

In comparing his results with observation, he points out that westward translation at about the expected speed ($\sim$10 cm/sec) had been observed early in the life of some eddies, but that this could equally well be explained by advection, of the whole eddy structure, embedded in a large-scale westward drift in the Sargasso Sea.

He notes (p. 522) that a puzzling feature of at least one observed eddy is a persistent elliptical shape, whereas his theory indicates that to first order, only circular eddies can persist without change in shape. He also implies (p. 521) that eddies decrease in size as they age, but the evidence for this is not clear. (McEwan (1948) claimed an increase in size with age.)

The question of interaction between a geostrophic vortex and a uniform wind stress has been discussed by Stern (1965). The main idea seems to be that conservation of vorticity will inhibit Ekman transport in near surface layers, and result in a coupling of the motion and a transfer of energy to greater depths. This paper might be relevant to the problem of explaining how the horizontal temperature structure in the surface layers is destroyed soon after the formation of an eddy.

Phillips (1966) dealt with the theory of "large-scale eddy motion in the Western Atlantic", but the word "eddy" here seems to mean "turbulent", and I am not clear if this paper is relevant to the "current ring" problem. A point that could be of interest is Phillips' interpretation of neutrally-buoyant float results. He points out that these show "the existence of considerable energy in transient motions whose period is so long that the currents must be quasi geostrophic". By combining float results and geostrophic velocities obtained from hydrology soundings, and considering the two linear geostrophic modes of a two-layer ocean, he deduced that "78% of the kinetic energy is in the barotrophic mode". This is a surprisingly large fraction.

Iselin and Fuglister (1948) point out that "the eddies that have been found both north and south of the Gulf Stream fall into two general classes. The first are frictionally driven or shear-zone eddies, often roughly 17 miles in diameter.....The second.... are considerably larger, rotate in the opposite direction, and are frequently found many miles away from the Gulf Stream". This paper gives a bathythermograph section (to 900 ft depth) across a large (200 mile long, 60 mile wide) cyclonic eddy, found in June 1947. It also includes some surface velocities in an anticyclonic eddy on the north side of the Stream. This eddy was apparently about 100 miles in diameter. This is the only reference I know to an anticyclonic Gulf Stream "eddy".

Vortices off Japan

Ichiye (1955a) gives dynamic heights and temperatures at 100 m depth for an anticyclonic vortex observed east of Japan, September 1953 - November 1954. Surface temperatures were not discussed. The vortex was about 400 km diameter, and probably formed from the eastward-flowing Kuroshio by a cut-off process. The main vortex appeared to split up, then recombine. It moved towards NE.

The formation of the vortex is discussed in terms of stability criteria. It was concluded that the model of Kuo (1949), might explain the initial stages of formation. (This model considers instability of planetary waves due to nonuniform vorticity distribution in a zonal current.)

The decay of the vortex is also considered. Figures 10(a) and (b) of the paper show respectively velocity and absolute vorticity as functions of distance from the centre, on 4 successive cruises, though it is hard to assess their accuracy, or the reality of some of the features shown. As "a possible process of dissipation", Ichiye suggests (p. 127) that in the initial stage, diffusion by eddy motion works only within the vortex, tending to smooth out the vorticity distribution. Exchange of vorticity with the outer region is controlled (= inhibited) by the large horizontal stability zone at the boundary. Then the vortex is divided into several parts, and the small vortices are more easily diffused because of their small scale, and because they have no barrier to the exchange of vorticity with the outer region.

A table on p. 128 gives the changes in angular momentum and kinetic energy with time. Between February-March and November 1954, angular momentum decreased from 3.5 to 0.8 x 10^{23} c.g.s. units (cm^4 sec^{-1} (?)) and kinetic energy decreased from 6.9 to 1.8 x 10^{17} c.g.s. units (cm^4 sec^{-2} (?)). (These values seem to be per unit depth, and with density = 1.)

The effect of the coast on the diffusion of vorticity is also discussed theoretically. It is claimed (p. 129) that the half-life time for vorticity decay is of the order 10 to 24 days when away from the coast, and 40 days when near the coast.

Horizontal mixing was thought to be more marked at 600-800 m than at shallower depths. It was claimed that this was associated with less "horizontal stability" (= vorticity gradient?) at such depths than in the near-surface layers.

In another paper, Ichiye (1955b) discusses, both practically and theoretically, a "cold water mass" (cyclonic eddy) that is sometimes present near the Japanese coast (centre approx. 33°N., 138°E.). This cold water mass is a well-known feature, and is mentioned in many Japanese papers. It exists continuously for many years, then disappears in a few months. It seems to be topographically anchored, and is probably not typical of the free eddies or current rings observed in the western Atlantic or off east Australia.

Eddies in other Areas

Duncan (1968) gives data on a large (300-400 km diameter) anticyclonic eddy, centred at 40°S., 15°E. The associated temperature and salinity structure extend down to at least 1200 m depth, but the station spacing of about 200 km is too coarse to give detail of the structure. The feature "disrupted" the subtropical convergence. The surface dynamic topography showed a dome of about 60 dyn. cm., and surface temperature varied appreciably. Darbyshire (1964) gives more details of the dynamics of the area, and refers to eddies but they seem to be relatively weak and ill-defined features (surface currents $\sim$ 10-20 cm/sec) in a complex topography.

Eddies have been observed to the west of the Hawaiian Islands (Patzert 1969). These have diameters in the range 50-150 km, volume transports up to 8.10^6 m^3, and are mainly confined to the upper 150 m. Both cyclonic and anticyclonic eddies have been observed, but cyclonic ones predominate. Total energies are in the range 1.10^{22} to 5.10^{22} erg. They were originally thought to be wake phenomena, but Patzert (1969) suggests they are "driven by strong local winds blowing through the restricted passage between the islands of Miami and Hawaii".

Bruce (1968) reports that a relatively small but intense anti-cyclonic gyre forms off the Somali coast in summer (southwest monsoon period), but is absent in winter. The gyre appears to be about 600 km diameter, but there are not enough observations to give a clear idea of its structure. This observation is interesting to the extent that the gyre forms as a direct response to a seasonal wind pattern.

Numerical Experiments

Bryan and Cox (1967, 1968a, 1968b) have reported the most elaborate numerical solutions yet attempted for the oceans. Briefly, their numerical model has the following properties:

1) Three dimensional
2) Non-linear
3) Constant depth, rectangular ocean
4) Both wind stress and surface heating as driving functions. These are functions of latitude only (i.e. no time variation)
5) Primitive equations are used
6) System starts from rest.

Of interest in the present context is their discussion of time-dependent motions that appear superposed on the steady solution. Figure 7 of the 1968b paper shows eddy-like disturbances in the western boundary current region. These are anti-cyclonic and move poleward, at the same time either growing or decaying. A typical diameter is 1000 km (i.e. four times as large as those observed off east Australia). An individual "disturbance" appears to have a half-life of the order of two to four months, which although small is at least of the order of magnitude observed in the ocean. Note however that in the numerical solution, the disturbances appear to be embedded in, and moving with, the mean flow; they are not produced by a loop cut-off process, as observed in the Gulf Stream and at least postulated off east Australia. The numerical model disturbances have a very large amplitude - nearly twice the total mean transport of the western boundary current. It is very hard to know how far one can expect model and real ocean to agree. We might note that the 1968 papers deal with a model in which the parameters are chosen to accentuate non-linear effects in the western boundary current region - in particular, the Rossby number (2×10^{-4}, as defined by Bryan and Cox) is an order of magnitude higher than for the real ocean. Bryan and Cox note (1968b, p. 977) that, except for the large amplitude there are many similarities between the disturbances in their baroclinic model and those found earlier in a barotrophic model (Bryan 1963). In the barotrophic model, the travelling disturbances only appeared when the Reynolds number was increased, and appeared to be generated by shear flow instability.

The disturbances in the baroclinic model affected the whole water column, with bottom velocities about half the surface velocities (1968b, Fig. 8), so that they are largely barotrophic. This seems to be at variance with what is known of vertical current profiles in the real ocean, although I do not know of any direct measurements of current beneath eddies or current rings. (Parker (1963) reports 1000 m depth drogue measurements showing clockwise (anticyclonic) motion with a diameter of ~ 300 km, at 30°N., 62°E., but there is no information on the current, temperature or density structure in the area at that time.)

Robinson and Niiler (1967) and Niiler and Robinson (1967) have developed the theory of steady free inertial jets in a stratified ocean with varying bottom topography. They compute the path of the stream axis, given the direction and curvature at one point as initial conditions. Numerical solutions are obtained for the Gulf Stream. Under some conditions, a solution may show a path looping back over itself. These solutions "are unphysical; however, they tend to point out the conditions which lead to the formation and detaching of eddies in the stream".

In the present context, there are two points of interest: what are these conditions? and what size of eddy would be expected? The first question cannot be simply answered, as far as I can see. The looping back in their Figures 3-6 (pp. 277-278) depends on the initial conditions (direction, curvature, bottom topography, vertical current profile near bottom) in a complex way, and no simple summary appears possible. (Note that in these four figures, the change in Coriolis parameter with latitude is neglected, i.e. beta = 0.) Application to a real ocean situation would be difficult, since some of the necessary parameters (e.g. stream curvature, bottom current or current shear near the bottom), would be very hard to estimate.

Since these papers imply that eddies form by a loop cut-off process, one would expect the eddies to have diameters of the order of twice the radius of curvature of the better-developed meanders in the computed paths. This leads to diameters in the range 600-1200 km, from the numerical solutions given in Part II. Observed eddies are generally smaller than this, but perhaps it is to be expected that estimating from the radius of curvature would give an upper limit.

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