Journal of Marine Research, 79, 109–117, 2021

On the process of upwelling

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The phenomenon of upwelling which is present along many coasts has been discussed from various points of view. A general explanation of the phenomenon has been given by application of Ekman's theory of wind driven currents (Thorade 1909; McEwen 1912), and emphasis has been put on the velocity of the vertical motion (McEwen 1934) or on the depth from which water is brought to the surface (Sverdrup 1930; Gunther 1936). Defant (1936) has discussed the horizontal and vertical motion and the distribution of density within the region of upwelling off the coast of southwest Africa, but his analysis is based on average surface conditions and on the results of a few, widely spaced oceanographic stations. His schematic representation of the currents in a vertical section at right angles to the coast appears misleading, since the vertical motion shown in the region of upwelling necessitates the existence of a horizontal divergence within this region. But a divergence is indicated only at the offshore border of this area.

A more detailed analysis of the process of upwelling off the coast of California can be undertaken by means of observations which were made by R. H. Fleming and E. G. Moberg of the Scripps Institution of Oceanography on board the "Bluefin," the vessel of the California State Fish and Game Commission during three cruises in March, May, and June, 1937. On all cruises a section was located nearly at right angles to the coast of California off Port San Luis (lat 35°10′N, long 120°45′W; see Fig. 60). The stations of March 25–27 were reoccupied on May 5–6, and on June 26–27. Although the exact locations differed slightly, we shall in the following discussion deal with the data as if the locations were identical.

Figure 61 shows the dynamic profiles on cruises I and II of the 0-decibar surface, the 100-decibar, and the 200-decibar surfaces relative to the 500-decibar surface. It is seen that above the 200-decibar surface a very conspicuous change took place: on March 25–27 the surface rose more or less irregularly from the coast to a distance of 180 km from the coast, but on May 5–6 the surface was nearly flat from the coast out to station 4,100 km from

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Originally published April 9, 1938, in the *Journal of Marine Research* 1(2), 155–164.

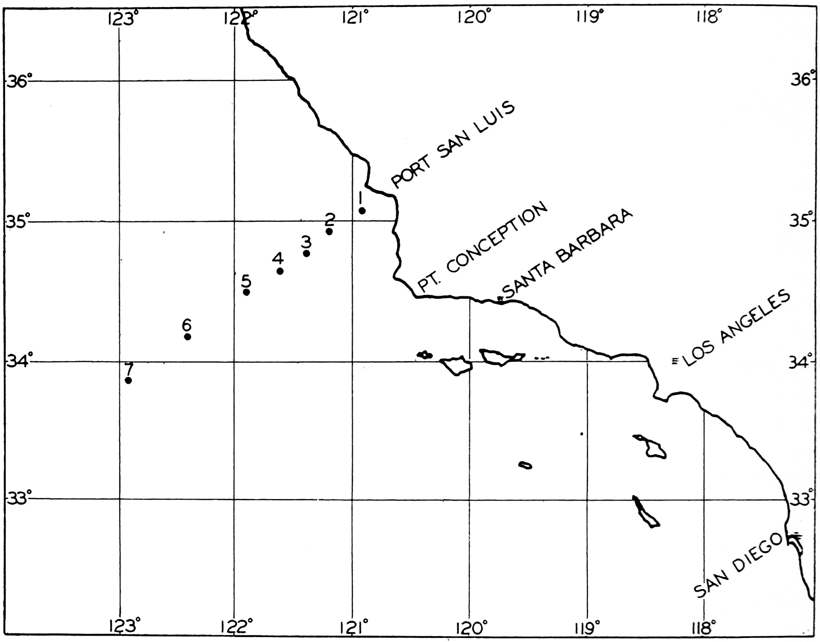

Figure 60. Chart showing the location of stations occupied on the northern line during the "Bluefin" cruises in March, May, and June, 1937.

the coast, and then rose steeply, not less than 0.116 dyn m in the 28 km between stations 4 and 5. The profile of the 100-decibar surface was changed similarly but less, whereas the 200-decibar surface remained nearly unaltered. It is evident that these changes must be due to a transport away from the coast and a banking-up offshore of the light surface water, and it is probable that the transport is caused by wind.

The wind direction and velocity in latitude 35° N, longitude 122° W, was derived from the weather maps constructed twice daily at the Lindbergh Aerological Station (San Diego). The direction of the wind was assumed to deviate 17° from the direction of the isobars and the velocity was obtained by multiplying the gradient wind by the factor 0.62 (Rossby and Montgomery 1936). During the five days prior to March 25, the average wind was 6.0 m/sec from S 85° W and, thus, the wind had blown from a direction which would transport the surface waters parallel to the coast. On March 23 and 24 the wind was strong enough to delay the beginning of the cruise. On March 25 the wind changed to NNW and during the following six weeks it blew from directions between N 55° W and N 20° E, occasionally reaching velocities as great as 12 to 14 m/sec. During the period of 41 days between March 25 and May 5 the average wind was 6.7 m/sec from N 23° W, and this wind which was nearly parallel to the coast would cause transport of the surface layers away from the coast. That such a transport took place is evident from a comparison of the distribution of density.

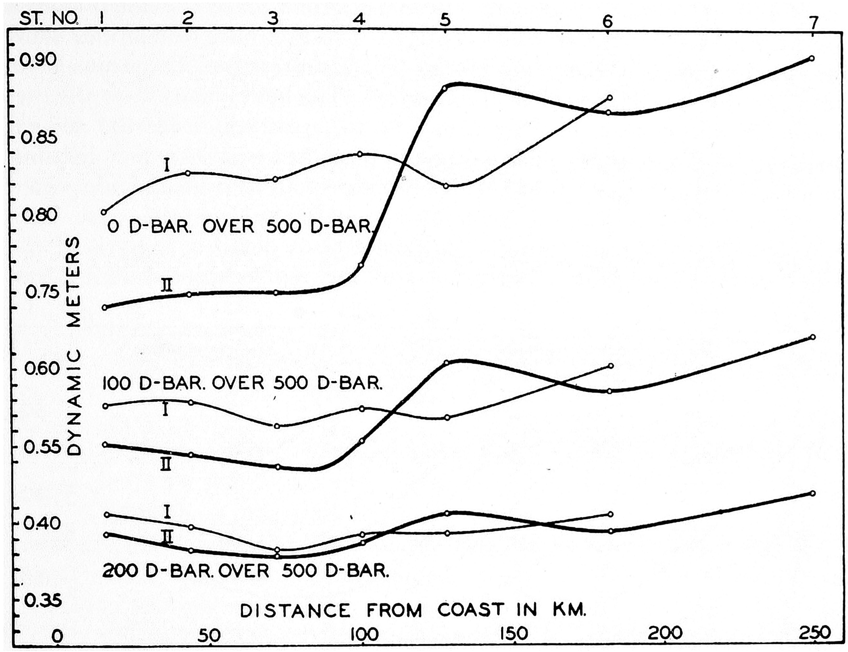

Figure 61. Profiles of isobaric surfaces relative to the 500-decibar surface on cruises I and II.

The thin lines in Figure 62 show the distribution of density on March 25–27 represented by means of σ_t which, in this case, give sufficient accuracy. The heavy lines show the corresponding distribution of May 5–6, and the arrows indicate the general character of the motion which may have brought about the new distribution. It is quite evident that the light surface-water has been transported outwards and banked up at a distance of about 100 km from the coast.

A somewhat different picture is obtained by a detailed analysis of the data. From the vertical displacement of the temperature, salinity, and oxygen curves the average vertical velocities during the 41-day interval were computed, assuming that at the surface the average vertical velocity was zero. Knowing the vertical velocities, the horizontal flow was found by means of the equation of continuity, assuming uniform velocities parallel to the coast. The details of the computations will be reported elsewhere. Here only the final result is shown in Figure 63. The lines with arrows represent the average stream lines in the 41-day time interval, and do not represent trajectories. The scale is arranged to show the inclination of the stream lines exaggerated in the same proportion as the vertical scale of the section (400:1).

The horizontal velocities are indicated by curves. The computed maximum offshore velocity is 11 cm/sec, in good agreement with the probable velocity of the wind current. The average outward velocity of the boundary region between the upwelled water and the

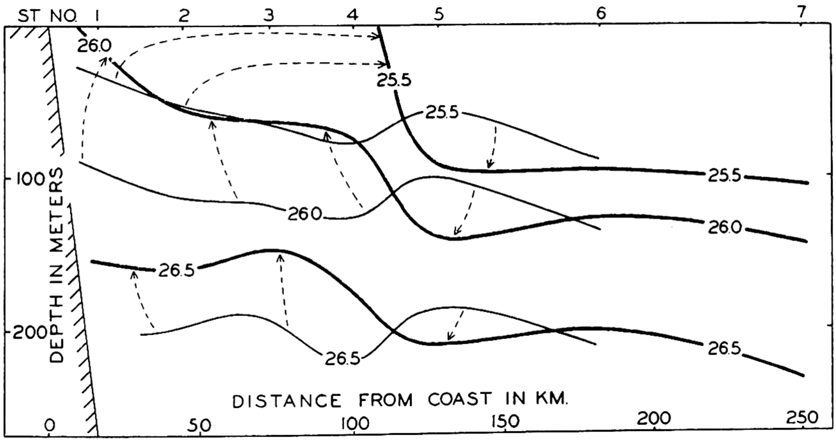

Figure 62. Vertical distribution of density (σ_t) on cruises I (thin lines) and II (heavy lines).

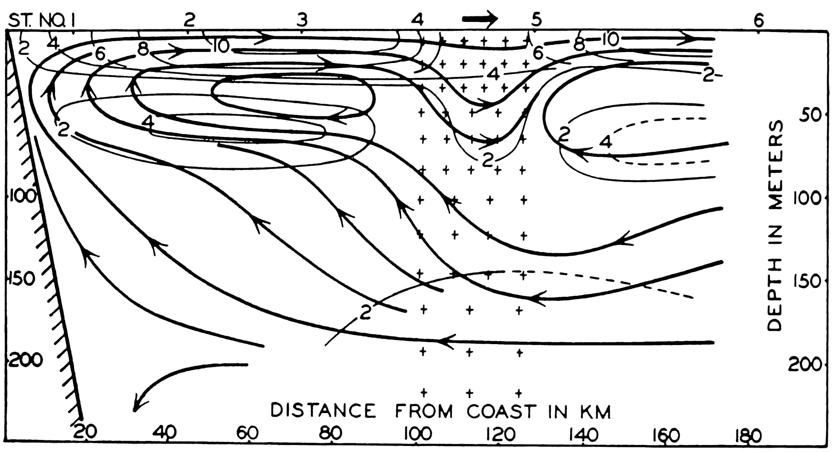

Figure 63. Computed vertical circulation in the time interval between cruises I and II. Direction shown by heavy lines with arrows; horizontal velocities shown by thin lines; areas with x's represent region with swift current parallel to the coast and away from the reader.

offshore surface water must however, in the period March 25–27 to May 5–6, have been only about 3 cm/sec since water, which on March 25–27 was found near the coast, was on May 5–6 found at a distance of 90 km from the coast. At a distance of 75 km from the coast the outward transport above a depth of 40 m, as computed from the equation of continuity, is 26×10^3 cm^3/sec, whereas a value of 25×10^3 cm^3/sec is obtained from the average tangential stress of the wind in the 41-day period, assuming a roughness parameter of 0.6 cm (Rossby and Montgomery 1936). The thickness of the upper homogeneous layer is 41 m

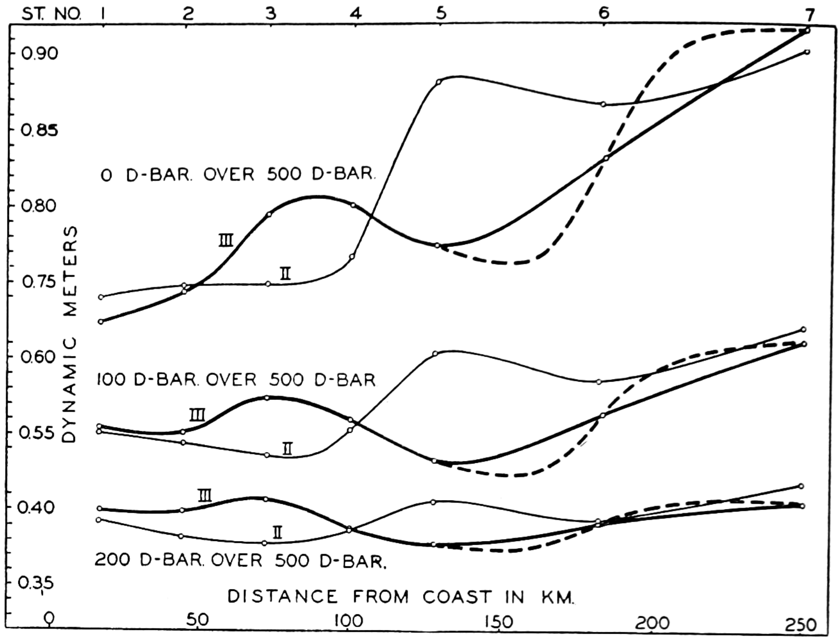

Figure 64. Profiles of isobaric surfaces relative to the 500-decibar surface on cruises II and III.

and Rossby's theoretical value of this thickness is 39 m (Rossby and Montgomery 1935). These agreements lend support to our representation of the flow.

The picture shows several striking features. Between the coast and the banked-up light surface-water we find a cellular circulation which reaches to a depth of about 80 m and within which the water flows away from the coast above 40 m, and towards the coast below 40 m. Near the coast the motion is ascending, near the offshore boundary it is descending. The offshore boundary is, however, moving slowly away from the coast and the inner cell is, therefore, being fed from below by water which flows toward the coast. This inflow appears to take place at depths not exceeding 200 m.

Figure 64 shows the profiles on cruises II and III of the 0-decibar, 100- decibar, and 200-decibar surfaces relative to the 500-decibar surface. It is evident that on cruise III the light offshore water was encountered at an even greater distance from the coast, but the banking-up effect does not appear so strikingly. When drawing the profile in the simplest manner, we find a gentle rise of the surface between stations 5 and 7, but between these stations we may actually have a steep rise on a short distance, as indicated by the dashed curves. Assuming that such was the case, we find that the boundary region has moved 70 km away from the coast in the 51 days between the cruises, and that inside the boundary region an eddy of a diameter of about 75 km is present.

A comparison of the density distribution (Fig. 65) shows that near the coast some upwelling has continued and that vertical motion has taken place directly inside of the

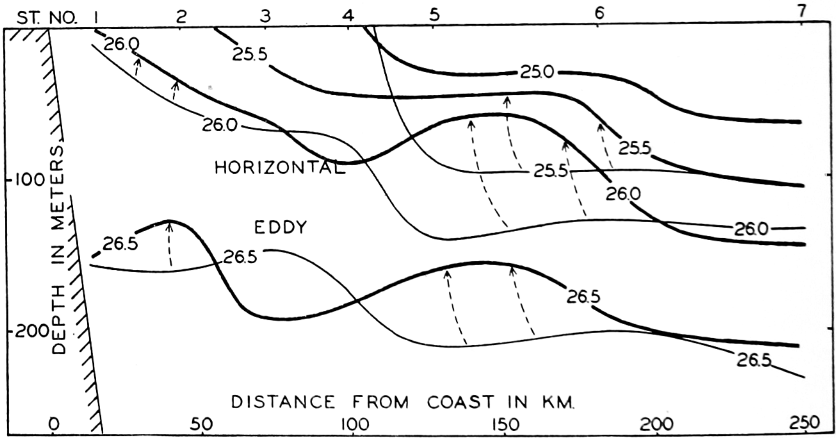

Figure 65. Vertical distribution of density (σ_t) on cruise II (thin lines) and III (heavy lines). Arrows indicate types of motion which have brought about the changes in the distribution.

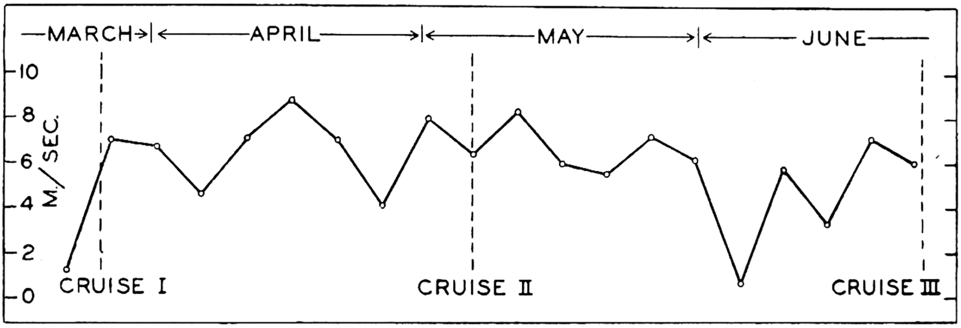

Figure 66. Component of wind parallel to the coast (from N 20° W in the period March–June).

boundary, but between these regions of vertical motion water masses of a different character have been embedded. A closer study indicates that these water masses are similar to those present at the coastal side of the boundary region on cruise II or, in other words, that an eddy had detached itself from the coastal side of the boundary. The low density of the water directly inside the boundary on cruise III (between stations 5 and 6) is due to heating, since the salinity is high. It has not been attempted to calculate any vertical or horizontal velocities since the processes are too complicated.

In the period May 5–6 to June 26–27 the wind was weaker and from a more westerly direction. In Figure 66 average 5-day values of the component of the wind parallel to the coast (from N 20° W) have been plotted for the time interval March 20 to June 28. It is seen that conditions were more favorable to upwelling between cruises I and II than between cruises II and III, and that between the latter cruises there was a period in which the component parallel to the coast was nearly zero. It is possible that the variable wind conditions in the time interval between cruises II and III were responsible for the breakdown of the boundary region and the development of the eddy, but it will be pointed out that other possibilities exist.

It does not seem feasible to make the described conditions the subject of a mathematical analysis, but certain features can be dealt with by simple reasoning. Let us assume that a tangential stress directed parallel to the coast and away from the reader is exerted upon the surface of the water within which the distribution of density is that shown by the thin lines in Figure 62. Under these conditions the uppermost light water will be transported away from the coast and must be replaced by water of greater density from some subsurface depth. The upwelled water will in turn be carried away from the coast and, since this water has a greater density, a current parallel to the coast must develop and flow in the direction away from the reader. If within this current the velocity decreases with depth, as was the case on May 5–6 (compare the slopes of the 0-decibar surface and the 100-decibar surface and observe that the profiles of these surfaces cannot very well be drawn with the same slope), part of the tangential stress of the wind will be balanced by the stress due to the shearing motion of the current. This means that the area within which the current has been developed must move offshore at a speed which is smaller than the speed by which the surface layers inside the current are transported outwards since there the full amount of the stress of the wind will be active. Thus, the region of the current will act as a barrier to the outward transport and inside this barrier a cellular circulation must develop. Since the barrier is moving, this cellular circulation on must either become thinner or it must be fed from below. It is not probable that it will become thinner since the depth of the offshore flow must be dependent on the wind velocity and be relatively independent of the motion of the barrier. We conclude, therefore, that water from greater depth is being drawn into the inner cell as the barrier moves away from the coast.

Outside the region of the current we must obtain a somewhat similar system. We must find ascending motion near the barrier which will act as a secondary coast, but the ascending motion is restricted to the upper layer of light water and has not the character of upwelling, that is, no deeper water is brought to the surface. Below the top layer we must find descending motion feeding the current which flows toward the coast below the boundary layer. If this were not the case, it would mean that the effects of the upwelling could be detected at very great distances from the coast.

The region of the current parallel to the coast must be a region of intensive mixing since it moves away from the coast at a velocity which is smaller than the surface velocity inside the region, and since the kinematic boundary condition, that the velocity normal to the boundary shall at all levels be equal to the velocity of the boundary, cannot be fulfilled. This may explain why our data cannot very well be interpreted as showing a sloping boundary surface between the light and the heavy water such as should be expected if we were dealing simply with the boundary between two currents, but that the simplest interpretation shows a region of transition characterized by a strong current parallel to the coast within which the velocity rapidly decreases with depth. Observations within such a current are, however, desirable in order to establish this feature definitely.

It is perhaps conceivable that this decrease of velocity may become so great that the stress of the wind is completely balanced against the stress due to the shear in the water. In this case the region of the current would take the character of a solid stationary barrier and inside of the barrier a stationary cellular circulation could be present. It seems, however, more probable that a stationary stage is not reached and that conditions such as were found on cruise II cannot exist for any length of time. A change in the wind velocity or direction which leads to a change in the transport of the surface waters may alter conditions to such an extent that large eddies develop on the boundary between the upwelled water and the offshore surface water. In our case it must also be considered that the current flows from north to south in a region within which the deflecting force of the earth's rotation changes rapidly with latitude. This may lead to unstable conditions and give rise to the formation of eddies. It also should be considered that, since the kinematic boundary condition is not fulfilled, intensive mixing must take place along the border of the boundary region, and this process may lead to the formation of larger eddies.

It is hoped that in the future it will be possible to undertake special series of observations during periods of upwelling in order to obtain better knowledge of the phenomenon and to answer many questions which now must be left open. The data from the three cruises in 1937 have, however, shown that the details of the process of upwelling are much more complicated than any previous records have revealed.

Summary

Observations on three cruises off the coast of California in 1937 indicated that during the first part of a period of upwelling the light surface-water is banked up at some distance from the coast. Between the heavier upwelled water and the light offshore water a boundary region forms with a strong current parallel to the coast. This boundary region moves away from the coast, but so slowly that it acts as a barrier against the outward transport of the upwelled water. Within this a cellular circulation develops which, since the offshore boundary moves away from the coast, is being fed from below.

This appears to be the initial stage, but as the boundary region moves out conditions become more and more unstable and large eddies develop on the coastal side of the boundary. The processes are far more complicated than assumed on the basis of earlier data.

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