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The Distribution of Ocean Temperatures along the West Coast of North America deduced from Ekman's Theory of the Upwelling of Cold Water from the Adjacent Ocean Depths.

By

Dr. George F. McEwen,
Physicist of the Marine Biological Station of San Diego.

With 21 Text-figures.

Sonderabdruck aus
Internationale Revue
der gesamten Hydrobiologie
□□□ und Hydrographie □□□

Herausgeber: BJÖRN HELLAND-HANSEN (Bergen),
W. A. HERDMAN (Liverpool), G. KARSTEN (Halle),
CHARLES A. KOFOID (Berkeley), L. MANGIN (Paris)
SIR JOHN MURRAY (Edinburgh), ALBRECHT PENCK
(Berlin), E. M. WEDDERBURN (Edinburgh). C. WESENBERG-LUND (Hilleröd), FRIEDR. ZSCHOKKE (Basel)
und R. WOLTERECK (Leipzig-Gautzsch), Redakteur

Verlag von Dr. Werner Klinkhardt, Leipzig

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The Distribution of Ocean Temperatures along the West Coast of North America deduced from Ekman's Theory of the Upwelling of Cold Water from the Adjacent Ocean Depths.

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Introduction.

The presence along the west coast of North America of a belt of cold surface water having at any point a much lower temperature than is normal for the corresponding latitude, has long been known. And several papers have been written in which a diversity of merely qualitative explanations of this interesting and perplexing phenomenon have been given. The present paper is an attempt to explain quantitatively the temperature distribution by means of a new theory of oceanic circulation, developed by V. W. Ekman (1) of Kristiana.

The contents of this paper fall under the following nine heads:

I. A brief summary of some important and generally accepted facts concerning oceanic temperatures and circulation.

II. A brief review of the theories that have been proposed to account for the cold-water belt along the west coast of North America.

III. An abstract of the most important part of Ekman's theory of oceanic circulation needed in attacking the above mentioned problems.

IV. Some general qualitative applications of his theory to a variety of temperature problems.

V. The formulation of a temperature problem in such a way that a quantitative estimate of the mean monthly surface-water temperature for any given place can be made by means of the physical theory of heat and circulation.

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VI. The solution of the above problem for four very different regions along the Pacific Coast, and a comparison of the observed and calculated values.

VII. A discussion of the results, and additional test of the theory using the observations made by the Marine Biological Association of San Diego in a much more limited area.

VIII. Some remarks on the influence of ocean temperatures on the coast climate of California.

IX. Summary and conclusion.

I. Some generally accepted facts regarding oceanic temperatures and circulation.

Under normal conditions, such as prevail in mid-ocean, the surface temperatures are subject to periodic changes with the seasons similar to the temperature[1]) variations in the air over the land as shown by the curves (Figs. 1 and 2). But the range of the ocean temperatures is only about half as much. The estimated daily range for the ocean is about 2°, while that for the air is about 10°. Observations show that in the ocean the daily temperature change does not extend below a depth of 10 to 20 meters, and the annual change can not be detected below 200 meters. The curve (Fig. 3) illustrates in a general way how the temperature varies with the distance below the surface. Water more than 200 meters below the surface has a temperature decreasing from 10° as the distance downward increases, often reaching the value zero on the ocean bottom, even at the equator. But the presence of several limited regions in which the temperature distribution differs notably from the normal, both with reference to depth and time of year has long been recognized, and considerable discussion relative to the reason for their existence has been carried on.

The question of the distribution of temperatures in the sea is so intimately connected with that of the character of its currents that it is practically impossible to separate them entirely. Consequently, as soon as one leaves the general question of the mean temperature of a given area, and wishes to decide whether that temperature is normal, and if not, from whence and how it is derived, the features of oceanic circulation must at once make part of the discussion.

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[1]) All temperatures are in the Centigrade scale.

As the motion of oceanic waters is partly determined by their temperatures, so their paths may often be traced out by the isothermal curves showing the surface temperatures. As the motion, when normal, is usually less accurately measurable than the temperature, and is much more rapidly lost, it often happens that the distribution of current water can be more accurately determined by a study of its temperature than in any other way. Thus, by means of its temperature, the existence of a given current can be determined with certainty over an area far exceeding that in which it can be proved to have a perceptible constant motion in any given direction.

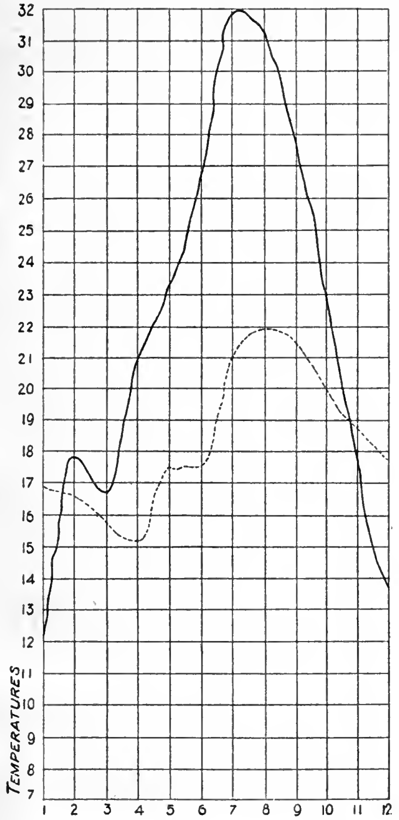

Fig. 1.

······ Curve for Surface Temperature of the Ocean, Lat. 32° 30′, Long. 140°.

— Curve for Air Temperature at Yuma, Arizona, where the Latitude is the same.

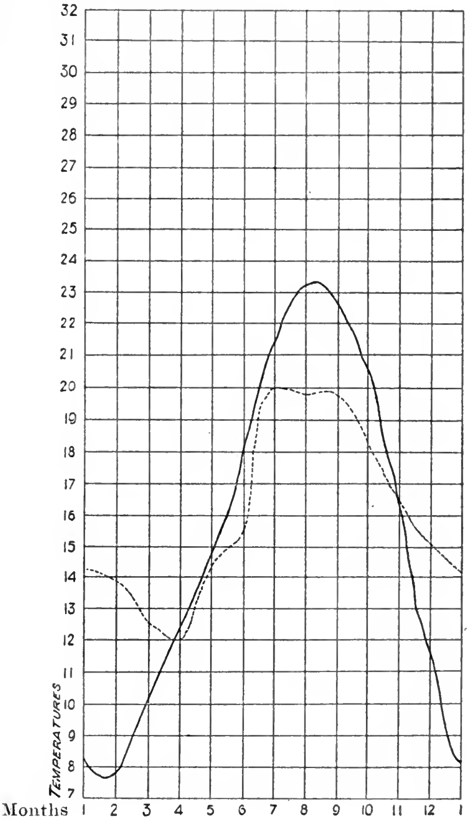

Fig. 2.

······ Curve for Surface Temperature of the Ocean, Lat. 37° 45′, Long. 140°.

— Curve for Air Temperature at Sacramento, where the Latitude is the same.

Conversely, if a large body of water be shown to have a nearly uniform summer temperature, corresponding in general with the normal value for the latitude, and with local circumstances in its particular portions, this is, of itself, evidence that no large body of water intrudes within its borders from a region of a different normal temperature. In other words, in the general oceanic circulation, a stream of water with a temperature normal to one latitude can not move to a region where another temperature is normal without exhibiting its presence by a deflection of isothermal curves.

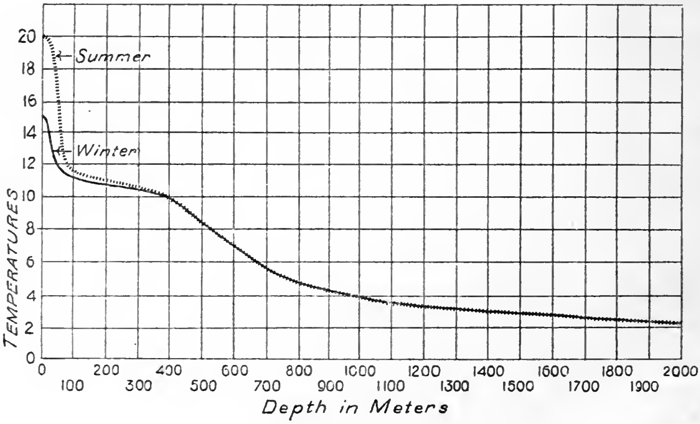

Fig. 3.

The source of the energy required to set and keep in motion the vast mass of ocean water has been productive of endless discussion. The attractive force of the moon, the lag of the water itself, the difference in temperature and density of the equatorial and polar regions, the unequal distribution of the atmospheric pressure has each in its turn been proposed and strenuously advocated as the true and only cause of ocean currents. To the seaman, however, the cause of the ocean currents has always been the winds, the motion of the waters of the sea taking its origin in the region where the winds attain their maximum constancy, that is, in the region of the trade-winds.

The following results of Zöppritz's (2) theory of ocean currents have been widely used by hydrographers. A perfectly steady wind acting continuously on the surface of the sea will thru friction, give rise to a movement of the surface waters in the same direction as the wind, itself. If the latter continues for a sufficient length of time, the impulse, first felt only at the surface, will gradually communicate itself downwards, owing to the viscosity (internal friction) of the water, and the lower strata to a successively greater and greater depth will thus partake of the movement until it is finally shared by the whole mass, the velocity diminishing as the depth increases. The rate, however, at which the motion is communicated to the depths of the ocean is exceedingly slow. For example, Zöppritz (2) estimated that in a depth of 4000 meters a surface current of a given velocity would require a period of about 100000 years to transmit a velocity of one half this amount to a point half way toward the bottom, and 240 years would be needed for the layer at the depth of 100 meters to attain one half the surface velocity. A period of 24 hours is required before a change in the direction of the wind affects the water at a depth of 5 meters. Similarly, when once established, these submarine currents exhibit a corresponding reluctance to undergo any change in direction or intensity. In general, according to Zöppritz, the time (t) in years required for a given constant surface velocity to produce a velocity one half as great at a depth of (x) meters is given approximately by the formula

$$t = 0{,}024\,x^2.$$

But observations show that perfectly steady winds do not exist, even in the region of the trades. The winds are constantly changing, and the surface currents change with them. The lower strata of the ocean, however are insensible to these changes, and at a considerable distance below the surface, the waters of the ocean have probably a slow but steady motion, the direction of the motion probably agreeing closely with the resultant surface winds.

James Page (3) summarises the conclusions based on Zöppritz's theory as follows. We have, therefore, in the body of the sea two distinct sets of currents; first, those at the immediate surface, which move practically at the obedience of the surface winds, sometimes in one direction and sometimes in another, second those of the lower strata, which are constant in direction and velocity and represent the aggregate effect of the winds that have blown for ages past.

II. Review of the theories that have been proposed to account for the cold-water belt along the west coast of North America.

1. Dall (4) has discussed the dependence of the cold in-shore water on the Japan Current somewhat as follows. The Japan Current is produced by the impinging of the Pacific North Equatorial Current on the

eastern shores of Formosa and the adjacent islands. While the larger part of the Equatorial Current passes into the China Sea, a portion of it is deflected northward along the eastern coast of Formosa, until reaching the parallel of 26° it bears off to the northeast, washing the whole southeastern coast of Japan, and increasing in strength as it advances to a limit which appears to be variable.

At the very outset the Japan Current must force its way thru the barrier of the Loochoo Islands and little later thru that chain of rocks, shoals and islets extending from Yokohama to the Bonin Islands, then it has nearly 6000 miles to traverse before reaching the opposite shore of the Pacific. Thruout this distance it is opposed by the northeast monsoon from the end of September to the end of February. Hence it is not surprising that its force should be checked, and its continuity as an eastward current should be for the time almost obliterated. In fact, according to many reports of navigators the current is subject to serious fluctuations which appear to be due to the monsoons, and as compared to the Gulf Stream is cooler and has a much smaller volume.

After reaching the northwest coast of North America near Sitka, Alaska it has a southward branch which, by the time it reaches San Francisco has become a cold current rather than a warm one, simply because it is intruding on a normally warmer area.

2. Richter (5) attributed the low in-shore temperatures to a polar current, and his argument may be summarised as follows. The question not as to the existence, but as to to the character of the ocean currents contiguous to the coast of California is still an open one. Some of the most recently published maps show that a cold current of great width washes our shores, and others again indicate that it is the deflected warm Japan Current which is passing this country in its southward movement. A third opinion gives the surface waters to the "Kuro Siwo" or Japan Current, and identifies the sub-stream with the polar current.

But his own conclusions are briefly as follows: The northern or arctic currents are powerful enough to alter materially the direction of the Japan Current. They sweep against the warm water as the polar waters meet the Gulf Stream on the north of Scotland. The arctic waters predominate on the surface by superior force until the "Kuro Siwo" gives a strong wall, which causes the cold current to pass underneath in the direction of the equator. One branch of this artic current continues south along the coast as far as Point Conception, latitude 34°.

The bulk of the Kuro Siwo trends eastward, but perhaps nowhere washes the shores of the United States, being separated from them by the narrow cold stream, and yet is near enough to exercise a powerful influence on the coast climate. Both streams flow to the south and produce a narrow eddy current next to the shore flowing north.

The directions of the observed currents can, according to Richter, be accounted for on the basis of Zöppritz's theory from the fact that the coast of the United States tends northeastward[1]) from Cape Mendocino, latitude 40° to Tatoosh Island, latitude 49°, and tends southeastward to San Diego, latitude 32° 45′, while the prevailing winds are easterly near Tatoosh Island and westerly near San Diego.

3. Bishop (6) assumed a flow of cold antarctic water northward along the ocean bottom, and explained how it would cool the inshore water as follows. It has been the custom to call the cold stream which is of a very low temperature, immense volume, and great velocity, and flows along the west coast of North America, a continuation of the Japan Current, but there are two insuperable objections to that solution. First, the Japan Current must necessarily pass out and dissipate itself in the vast breadth of the Pacific Ocean long before reaching Alaska. Second, the Japan Current is entirely too warm, and the Alaskan glaciers are inadequate to materially cool the adjacent waters.

But the antarctic continental glacier extending 4000 miles in length along the antarctic circle cools the ocean water, and because of its increased density this cold water sinks to the bottom and flows northward, being forced on by the accumulated supply behind it. After passing the Tropic of Cancer, the Pacific Ocean contracts in breadth. In latitude 45°, it is only half as wide as at the equator. Hence, the northward moving water is accelerated to twice its previous velocity. From the rotation of the earth this current would be forced to the east against the North American continent in the latitude of Sitka and Vancouver's Island. The resulting pressure drives the water up the continental slope and it then flows southward, having no other outlet. It continues along the coast, being held there in opposition to the westward deflecting force due to the earth's rotation by the eastward pressure due to the bottom current which continually pushes up from the deep sea under the easterly globe-thrust along the entire west coast of North America.

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[1]) The coast actually trends slightly to the west.

4. Dall (7) refuted Bishop's argument and reasoned as follows in favor of the commonly accepted view (8), that the Japan Current is a cold stream bordering the California coast and has a relatively low temperature because of its passage thru high latitudes. The northeast trades, blowing hard and steadily for ten months in the year carry the warm water, which the Japan Current delivers in mid-ocean, to the northwest coast of North America which it reaches at latitude 54° near Sitka, Alaska. There, the stream divides into two branches, a northerly and a southerly one. At this point the maximum temperature is 20°, but the average is about 15,6°. As it moves down the coast, it loses its heat, and produces the fogs and rains of the Oregonian regions cooling off so that when it reaches the latitude of the Golden Gate, it is colder than the sea water under normal conditions in that latitude. That this essentially superficial stream is not due directly to the impinging of cold antarctic water on the northwest coast seems to be certain from the fact that the temperature of the latter is nearly zero, while the current when it reaches the coast is 17° or more warmer than that; and that the water of the current is warmer in latitude 54° than it is in the more southern part of its course, whereas, if it was abyssal, we should expect it to be colder, and to gradually warm up as it moved south-ward exposed to the action of the sun.

The Pacific is open without stint to the influx of antarctic cold water, and in it probably goes on a great system of true oceanic circulation such as might occur were there no continents. The general system of oceanic circulation is influenced by the rotation of the earth, differences of density and temperature in the oceanic mass, tides, the pressure of the atmosphere, and various minor causes, and there are no indications that would lead us to believe that such movement are other than very slow and gradual, or that they have any marked effect in producing the superficial streams of rapidly moving water which we call currents.

5. Holway (9) showed that the hypothesis of a surface stream flowing south and being abnormally cold because of its passage thru high latitudes can not be reconciled with observations, and attributed the cold surface water to an upwelling of bottom water as follows. The cold in-shore water cannot be due to a surface current flowing parallel to the coast, as observations of the surface temperatures have indicated narrow belts of alternately warm and cold water directed normally to the coast, the temperature differences being 2° or 3°. Also the coldest part of the cold-water belt is near Cape Mendocino latitude

40°, there being warmer water both to the north and to the south of this region. But a cold current flowing down from Alaska would continually rise in temperature as it moved southward. The only remaining explanation is that there is a belt of cold water upwelling from the adjacent ocean depths.

The above surface temperature relations present two problems. First, what is the cause of this cold-water belt? We have already seen that it must be due to an upwelling from the adjacent ocean depths, but what is the cause of this upwelling? And secondly, why should the coldest portion of this area be in the vicinity of Cape Mendocino, instead of farther north?

Andrees (10) attributed the cold-water areas to a vertical current caused by the winds blowing off shore and driving the surface water to the leeward, thus causing a return drift at the bottom of the ocean and an upwelling near shore. Hann (11) accepted this view and added "the sharp deflection of an ocecan current off shore may cause a rise of cold water from below". He also assumed that the whole coast of California belonged to the area of the constant trade winds. Buchanan also accepted the theory of Andrees. But observations show that the trade-wind region does not extend as far north as California and the component of the wind velocity normal to the coast blows toward the land except for a few weeks in the winter time, the season when the temperature is nearest to the normal.

Another peculiar fact concerning the temperature distribution is shown by the charts made by Sir John Murray (12). They indicate that water temperatures over the continental shelf are notably less than those at the same level but farther out over the great ocean depths.

The observations and hypotheses referring to the relatively cold water lying along the west coast of North America may, according to Holway, be summarised as follows:

a)¹) The definite belt of cold water can not be traced south of Point Conception.

b) In the summer, the coldest part of this belt is in the vincinity of Cape Mendocino.

c) The source of this cold coast water is in the ocean depths to the northwest of Cape Mendocino.

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¹) The observations a) and b) and the theory of Zöppritz have an important bearing on the hypotheses c) and d).

d) This cold water at or near the ocean bottom has a slow drift agreeing in direction with the average direction of the surface drift, and is driven to the surface on striking the slope of the continental shelf. Local variations in the temperature of the cold coast water are due to the submarine valleys and other irregularities in the slope of the continental shelf.

6. A study of the preceeding investigations shows that the general method employed was to first examine the data and then to formulate a hypothesis to account for the observed results, the hypothesis being supported partly by the fact that it accounted in a qualitative way for the particular observations used, and partly by the theory of Zöppritz. Some writers used the effect of the earth's rotation and others did not. Sometimes erroneous statements regarding the observed facts were used in support of the explanation. The temperature distribution was regarded as constant thruout the year. No quantitative tests were attempted, and the explanations given by the different writers are inconsistent with each other. So the question is still an open one.

III. Ekman's theory of oceanic circulation.

It is well known that the motion of large masses of water in the ocean is accompanied by irregular vortex motions that cause the computation of the actual frictional forces according to the methods of rational hydrodynamics to be worthless. There is, in fact, an obvious disagreement between the results of hydrodynamics on the one hand and experience on the other. Therefore, for practical purposes we have had to be satisfied with purely empirical formulae, which because of their very limited field of application, have been unable to afford any help to oceanography. But all of the factors affecting the motion of ocean water can be taken into account if problems of a sufficiently simple type are formulated. By devising a series of such simple typical problems and solving them by exact analytical methods in which all of the factors are used, a sort of framework can be formed about which a theory of ocean currents corresponding to the different actual problems can be built. The actual problems, however complex they may be, can then be most readily attacked, not by analytical methods, but by suitably combining the proper typical problems that have been solved.

By proceeding in this way a number of very important results have been obtained which are the immediate consequences of the general principles of mechanics, and are therefore independent of any hypothesis as to the laws of fluid friction.

The following discussion will include only a brief statement of the assumptions made by Ekmann (1), and a few of the most important results that he obtained. It is a well known fact that the rotation of the earth upon its axis tends to deflect (13) a body moving along its surface to the right of its path in the northern hemisphere and to the left in the southern hemisphere. Therefore, in addition to the forces given by the conditions of the problem, this deflecting force must be introduced. It can be proved that a correct solution of a problem concerning the motion of a body on the earth's surface will result if we assume no rotation of the earth upon its axis, but introduce the force.

1. $$F = 2\,V\,\omega \sin \Phi$$

into the equations of motion and solve in the usual way. Velocities and displacements are measured relative to the earth's surface, (V) is the velocity of the body (ω) is the angular velocity of the earth, (Φ) is the latitude of the place, and (F) is the force perpendular to the direction of motion, as shown in (Fig. 4).

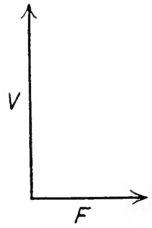

Fig. 4.

That is, the deflecting force due to the earth's rotation is proportional to the velocity of the body relative to the earth's surface and increases from zero at the equator to a maximum at the poles, and for a horizontal velocity in any direction is directed normally to the motion. (Fig. 4) corresponds to the northern hemisphere, but the arrow (F) would have to be reversed to correspond to the southern hemisphere.

First typical problem. Imagine a large ocean of uniform depth and without differences of density affecting the motion of the water. The influence of neighboring ocean currents and continents is left out of account so that water can freely flow in or out of the region considered. Suppose the water surface to be impelled by a steady and uniform wind equal in strength and in direction over the whole region, and that these conditions have continued long enough to establish a stationary state of motion. Introducing the deflecting force (F) into the equations of motion of a viscous fluid and solving in the usual way gives the following results for the northern hemisphere. The velocity of the surface water is directed at an angle of 45° to the right of the wind velocity (Fig. 5). The magnitude of the velocity of the water decreases as the distance below the surface increases, the direction of the velocity of each layer being to the right of that above it. At the depth

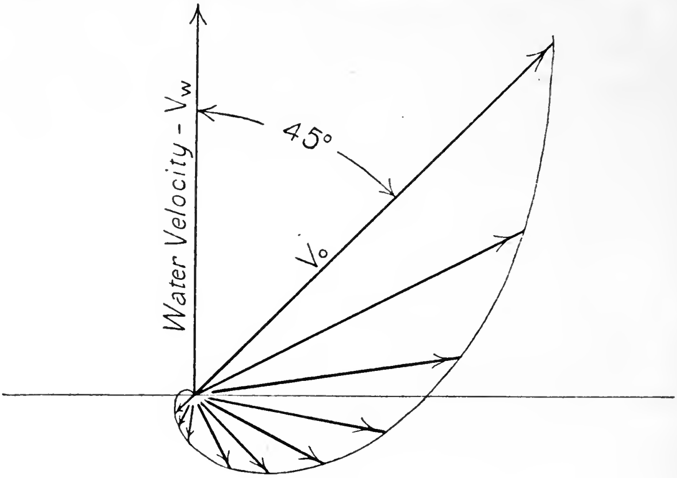

Fig. 5.
The arrows show the velocities of the water at the depths 0,0 D, 0,1 D, etc., below the surface. V_0 is the surface velocity.

2. $$D = \pi \sqrt{\frac{\mu}{q\,\omega \sin \Phi}}$$

where (μ) is the coefficient of viscosity and (q) equals the density, the magnitude of the velocity is only $^1/_{20}$ of that at the surface, and is in the opposite direction. This distance (D) is called the "depth of the wind-current", and the motion at greater depths is assumed to be negligible. Another result is

3. $$V_0 = \frac{T_1}{\sqrt{2\mu q\,\omega \sin \Phi}}$$

where (V_0) is the surface velocity of the water and (T_1) is the tangential pressure at the surface due to the wind.

The general character of the motion may be illustrated as follows. Imagine a spiral stairway so situated that the edge of the top step is directed at an angle of 45° to the right of the wind velocity, thus coinciding with the arrow (V_0) of (Fig. 5). Now, if as we descend, the successive edges are shortened so as to have the successive lengths of the arrows of the diagram (Fig. 5), each edge will represent in magnitude and direction the velocity of the water at that depth. And by the time a half turn had been made the edge of the step, that is, the velocity of the water would be only $^1/_{20}$ of its value (V_0) at the top, and from there downward the velocity would be still smaller. So, for practical purposes we can neglect the motion below that point.

The total momentum of the “wind-current„ is directed at right angles to the wind itself. The flow or volume of water per unit time transported[1]) parallel to the wind is zero, and that perpendicular to the wind is

4. $$\frac{V_0 D}{\pi \sqrt{2}} = \frac{T_1}{2 q \omega \sin \Phi}.$$

Therefore, the direction and magnitude of the flow depends only on (T_1) and not upon (μ). Within this surface layer of thickness (D) it is evident from (Fig. 5) that water is actually transported in various directions and that the amount carried per unit time in a layer of given thickness decreases as the depth increases. But when we add up the amounts flowing parallel to the wind in the whole layer, it is found that there is as much water flowing with the wind as against it, so the total quantity per unit time, that is, the flow parallel to the wind is zero. While when we sum up the amounts moving normal to the wind the value $\left[\left(\frac{V_0 D}{\pi \sqrt{2}}\right) = \frac{2}{9}(V_0 D)\right]$ is obtained, the direction or the flow being to the right of the wind. That is, a uniform wind blowing over deep water in a region remote from any obstructions, will transport water only at right angles to its own direction, the flow being the same as if the velocity of all the water in the layer of thickness (D) was directed to the right of the wind and the magnitude was $^2/_9$ of its value at the surface.

The experimental value of (μ) for laminar motion is about 0,014, and when substituted in (2) gives the value

5. $$D = \frac{44}{\sqrt{\sin \Phi}} \text{ centimeters.}$$

The theory of Zöppritz is based on exactly the same assumptions as the above except that the influence of the earth's rotation is neglected,

---

[1]) Thru a vertical rectangle of unit width and perpendicular to the direction of the motion.

and his result is entirely different. But as was stated before, the motion is not laminar, but is turbulent and the ordinary value of (μ) which gave the absurdly small value of (D) must be replaced by a virtual value much greater than 0,014. The virtual value

$$6. \qquad \mu = \frac{D^2 q \, \omega \sin \Phi}{\pi^2}$$

would be different under different conditions of wind, velocity etc., and can only be determined by current measurements and other observations carried out under varying circumstances. From the rough measurements now available[1]), a mean value of (D) would be 75 meters.

Experiments on the relative velocities of the wind and surface current led to the following approximate relation:

$$7. \qquad V_0 = \frac{0{,}0127}{\sqrt{\sin \Phi}} V^1_w$$

where (V^1_w) is the wind velocity and (V_0) that of the surface water. That is, at the latitude 45° the velocity (V_0) of the surface water is approximately $^1/_{55}$ times that of the wind (V^1_w) which causes it. This multiplier would be about 0,013 at the poles, and would increase as the equator is approached. If the wind velocity is in miles per hour and the value of (V_0) is required in meters per second the formula becomes

$$8. \qquad V_0 = \frac{0{,}00569}{\sqrt{\sin \Phi}} V_w.$$

From Ekman's theory, the time required for a steady current to produce any fraction of the final limiting value is independent of the value of (μ) and the current would be practically fully developed in 24 "pendulum hours"[1]). And thus outside of the tropics where this theory does not hold, only a few hours are required to set up a stationary state of motion, and the enormous times that Zöppritz computed on the basis of laminar motion and the ordinary value of (μ) are meaningless.

Second typical problem. Assume as before an infinite ocean of uniform depth (d), [(d) is greater than (D)] and of uniform density (q) Suppose the surface to be inclined at a constant angle (Φ). Solving as before, the following results were obtained. The current will consist

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[1]) Ekman deduced the following approximate formula for determining (D).

$$(D) = \frac{7.6}{\sqrt{\sin \Phi}} (V^1_w).$$

of a "bottom-current" of thickness (D) running more or less in the direction of the force, and above this a current reaching right up to the surface with the almost uniform velocity

9. $$U_0 = \frac{g \sin \Phi}{2 \omega \sin \Phi}$$

perpendicular to the force, where (g) is the acceleration of gravity, and (Φ) is the inclination angle of the surface with a horizontal. In order to represent the result graphically, it is sufficient to so turn (Fig. 5) that the longest arrow points perpendicularly to the left of the pressure gradient. Then add to each arrow in turn the constant velocity (U_0) to the right. The successive resultants will be the velocities at the distances 0,0 D, 0,1 D, 0,2 D etc. above the sea bottom. (Fig. 6) has been constructed in this way, (OY) being the direction of the pressure gradient. In the bottom-current the flow in the direction of the pressure gradient is

10. $$S_y = \frac{U_0 D}{2 \pi} = 0{,}159\, U_0 D,$$

while the flow normal to the pressure gradient is

11. $$S_x = \frac{2 \pi - 1}{2 \pi} U_0 D = 0{,}84\, U_0 D.$$

The significance of the above problem may be further brot out by the following explanation. Fluid motion can also be generated by difference in pressure, and in a region of the ocean far removed from obstructions and having a uniform depth greater than (D), it follows from Ekman's theory that if the pressure decreases as shown by the arrow in (Fig. 6) the resulting motion from the upper surface downward to a distance (D) above the bottom will be uniform, tho not in the direction in which the pressure decreases but at right angles to that as shown by the arrow (U_0) of (Fig. 6). And the motion between the bottom and a surface at the distance (D) above the bottom can be represented by another spiral stairway, the edge of the top step coinciding with (U_0) and the successive edges decreasing to zero at the bottom by having the succession of values of the arrows in (Fig. 6). By summing up, as before, the amounts flowing parallel to the pressure gradient and then the amounts normal to the gradient in the different layers of this "bottom stream" it will be found that the total flow or volume per unit time transported in the direction of the gradient is 0,159 U_0D while the flow in a perpendicular direction is 0,84 U_0D.

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[1] 1 pendulum hour = $\frac{1 \text{ sideral hour}}{\sin \Phi}$.

Third typical problem. Assume a steady uniform wind blowing in a constant direction everywhere outside a straight and infinitely long coast. Under these conditions, the depth of the ocean being supposed uniform, the current would be the same at any two places at the same distance from the coast, and no inclination of the surface can occur in the direction of the coast itself. Perpendicular to this direction a slope will arise and gradually increase until the total flow normal to the coast is zero. If the depth exceeds (2 D) there will be three distinct currents: first, a “bottom-current” of depth (D) moving more or less in the direction of the slope, but with a deflection to the right increasing from 45° at the bot tomto 90° at the top: second, a “midwater-current” of almost uniform velocity parallel to the coast and reaching from the top of the “bottom-current” to the depth (D) below the surface. (The velocity of this current is proportional to the component of the wind velocity parallel to the coast): third, a “surface-current” in which the volocities are equel to those of a wind-current” superposed on the velocity of the “midwater-current” „The bottom-current” and the “surface-current” will not be appreciably influenced by an alteration of the depth (d) as long as it exceeds (2 D), and the only effect then will be a corresponding alteration of the depth of the uniform “midwater-current“.

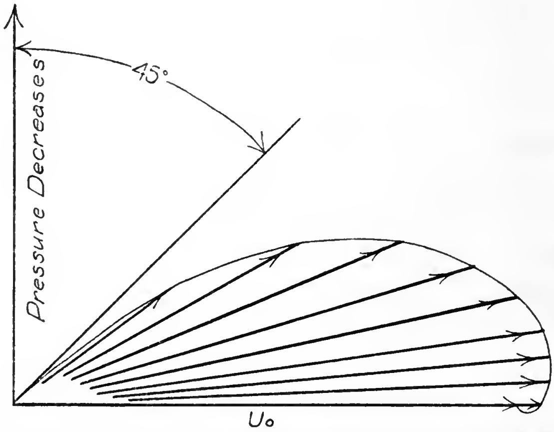

Fig. 6.

The most striking result of the coast’s influence is that a wind is able indirectly to produce a current more or less in its own direction from the surface down to the bottom, while in the absence of coasts the wind's effect would be limited to a comparatively thin surface layer, even if blowing steadily for any length of time.

The general circulation of the water in the neighborhood of a coast is shown in (Fig. 7) which represents a cross-section perpendicular to the coast. The wind is assumed to blow parallel to the coast, and normal to the paper from the reader. The arrows show the components of the velocity of the water in a plane perpendicular to the coast. The actual velocity would be the resultant of that shown and a component parallel to the wind.

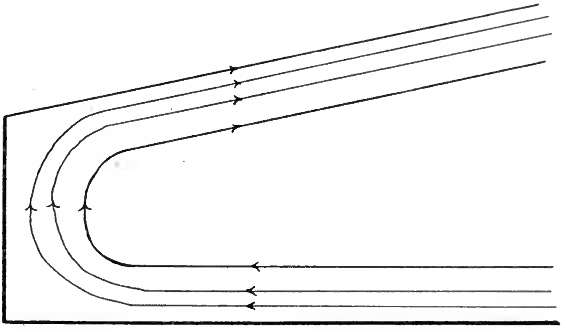

Fig. 7.
Cross Section Normal to the Coast.

If the wind should blow in the opposite direction to that shown, the motion of the water would be reversed. Thus the component of the wind velocity parallel to the coast causes an upwelling of bottom water under the conditions shown, or it carries surface water to the bottom if its direction is reversed. The rate at which water is carried up or down is proportional to the magnitude of the component of the wind velocity parallel to the coast ¹).

From a computation of the time required for the stationary motion to be established, the following general conclusions may be drawn. Assuming (D) = 75 meters, the stationary state of motion will be practically fully established to within several hundred kilometers from the

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¹) As the depth (d) diminishes from the value of (D) in problems 1 and 2 or diminishes from (2 D) in problem 3 the motion becomes less influenced by the earth's rotation, that is the flow due to a wind tends to have the same direction as the wind, and that due to a pressure gradient tends to flow from a region of high to one of low pressure. But no important alteration in the results worked out above will follow until (d) is less than (0,5 D) in 1 and 2, or less than (D) in 3.

coast in a few days if the depth does not exceed about 400 meters (that is over the continental shelves). In the deep ocean, on the other hand, particularly in the case of very broad currents, say 1000 or more kilometers in width, the "midwater-current" may require several months to become approximately fully developed[1]). Thus the effect of a wind is entirely different from what Zöppritz's theory shows, and the time required for producing the effect is measured in days or months, rather than in geological periods.

IV. Qualitative applications of Ekman's theory to several peculiar temperature distributions.

Since, if the preceeding theory is true, it is possible for a wind acting under proper conditions to bring bottom water to the surface, and since we know from observations that the bottom water is cold, abnormally low surface temperatures may be due, at least in some cases, to a combination of circumstances which would produce an upwelling of the cold deep water into the warmer surface water.

H. Thorade (14) has compiled, from all reliable sources available, a large amount of detailed information concerning the prevailing wind directions and surface temperatures over the North Pacific Ocean corresponding to each month of the year (Figs. 20 and 21 are copies of two of his temperature charts, and they also indicate the prevailing surface currents for the months of Jan. and Aug. respectively). He assumes that the low temperatures near the coast are due to the upwelling of cold bottom water, and points out a close correspondence between the prevailing wind directions and the temperature distribution He does not give the velocity of the winds, but only the prevailing directions, and devotes only one page to a qualitative application of Ekman's theory, the results of which are shown to agree in a general way with the observations on the seasonal distribution of temperatures. This is the first application of the new theory of upwelling to the California region.

The statement made by Holway (9) that during the summer the

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[1]) The approximate value of the time required for such a current to attain 0,7 of its final value is given by the formula

$$\left\{\frac{d}{D \sin \Phi} + 0{,}0036 \frac{x^2 \sin \Phi}{D}\right\} \text{days}$$

where (x) is the distance from the coast in kilometers across the stream and (d) and (D) are depths in meters.

coldest surface water was in the vicinity of Cape Mendocino, lat. 40°, is fully verified by the maps in Thorade's (14) article. And from the present theory we would expect the water there to be colder than that farther south because of the higher latitude and the upwelling combined. North of this region the wind component parallel to the coast is directed to the south but rapidly diminishes, being only 2 or 3 miles per hour off Vancouver, less than 500 miles farther north, while the velocity at latitude 40° is about 15 miles per hour. So we would expect the quantity of cold water upwelling to diminish proportionately in that distance. The normal temperature for the latitude of Vancouver is about 2° more than the actual temperature off Cape Mendocino from June to September, but is about 2° less during the rest of the year. The charts show as would be expected that off Vancouver the temperature is practically independent of the distance from the coast.

The presence of abnormally cold water on the continental shelf, referred to in Holway's article, would result directly from Ekman's theory, as the prevailing wind direction is such as to drive the bottom water toward the coast over this slope.

Along the west coasts of Africa and South America the prevailing wind directions (15) are such that a temperature distribution similar to that off the west coast of North America would be expected on the basis of Ekman's theory, remembering that in the southern hemisphere the deflecting force is directed to the left of the motion. This agrees with the observations which indicate a reduction of in-shore temperatures corresponding to the strength of the wind component parallel to the coasts.

The qualitative agreement being so satisfactory, it seemed very desirable that a detailed and quantitative test should be made. And I have attempted to so formulate the problem of the distribution of ocean temperatures along the west coast of North America that it could be put into mathematical language and solved with the aid of Ekman's equations of fluid motions.

V. The mathematical formulation of the temperature problem for a given locality.

Assumptions. The normal in-shore temperature for any latitude is the same as the actual temperature at a point in mid-ocean having the same latitude.

The difference between the actual temperature and the normal temperature is due entirely to the mixture of cold water from the adjacent ocean bottom with the surface water.

The cold water upwelling in a particular region is due entirely to the winds in that latitude, and no cold water from the surface in other localities enters the region.

Observations and deductions. An examination of the isotherms of Thorade's charts showed that the temperature of a surface layer of water is approximately constant for a certain distance (x_1) out from the coast, and increases in proportion to the distance from that point out to a point whose distance is (x_2) from the first, and remains nearly constant from this point to the limit of his map, longitude 140°. Therefore the amount of heat in this layer of unit width and thickness (y) would be

$$y\,\{x_1 T + \tfrac{1}{2} x_2 (T + t_2) + x_3 t_2\}$$

where (T) is the actual in-shore temperature, and (t_2) is the normal temperature for the latitude.

Assume that this amount of heat is the same as if the total volume $y\,\{x_1 + x_2 + x_3\}$ was, to begin with at the normal temperature (t_2) and a volume (xy) at the temperature (t_2) was then removed from the region and replaced by an equal volume (xy) upwelling from below at the temperature (t_1). (T) and (t_2) denote mean monthly temperatures corresponding to the latitude, and (xy) will be assumed to be the volume upwellling into the surface layer during the month's time ending with the date to which (T) and (t_2) correspond. Equating the two expressions for the amount of heat gives the equation

12. $$y\left\{x_1 T + \frac{1}{2} x_2 (T + t_2) + x_3 t_2\right\} = \left\{\overline{(x_1 + x_2 + x_3 - x)}\, t_2 + x\, t_1\right\} y.$$

The actual temperature at any instant is the result of the continuous action of various causes, absorbtion of heat from the sun, radiation, evaporation and the intrusion of cold water. But, in order to simplify the computation, it is assumed that the actual continuous process can be replaced by the following artificial one. Assume that whatever the temperature for any given month may be; the temperature a month later would be normal were there no intrusion of cold water; and assume that if all the water that actually intruded during the previous month were then quickly mixed with the surface layer the resulting temperature would be the same as the actual temperature. A time interval of one month was adopted because the computation of the mean temperature for each month can then be made directly, and compared with the observed temperatures, which are given for the same time interval.

From equation (12) the following equation for calculating (t), the difference between the normal temperature (t_2) and the actual temperature (T) was obtained:

13. $$t_2 - T = t = \left\{\frac{x\,y}{\left(x_1 + \frac{x_2}{2}\right) y}\right\} (t_2 - t_1) = \left\{\frac{x}{x_1 + \frac{x_2}{2}}\right\} \{t_2 - t_1\}$$

where (x) is the length, normal to the coast that multiplied by the cross section ($y \times 1$) of the volume considered, gives the amount of cold water intruding, and ($x_1 + x_2$) is the distance from the coast out to the point where the temperature is practically normal. Now if this quantity (x), which depends upon the amount of water upwelling in one month, can be determined, then the temperature reduction (t) can be calculated by substituting in equation (13) if the normal temperature and that of the upwelling water are known.

With the aid of Ekmans theory (x) can be computed as follows. The flow normal to the coast due to the surface current between the depths (z_1) and (z_2) is

14. $$s = \int_{z_1}^{z_2} u\,dz$$

where (u) is the velocity normal to the coast. From Ekman's equation this reduces to

15. $$s = \int_{z_1}^{z_2} V_0\, e^{-az} \cos(45^0 - az)\, dz$$

where

16. $$a = \sqrt{\frac{q\,\omega \sin \Phi}{\mu}} = \frac{\pi}{D}$$

Integrating between these limits and simplifying the result we have

17. $$S = \frac{V_0\, D}{\pi \sqrt{2}} (1 - 2^{-k\pi} \cos k\pi) = \text{approximately,}^{1)} \frac{V_0\, D}{\pi \sqrt{2}} \left(k\pi - \frac{[k\,\pi]^3}{3}\right)$$

where a surface layer down to the depth (k D) is considered.

Observations show that the temperature is practically uniform from the surface down to the depth of about 5 meters, so the value of (y) can be taken equal to 5 meters or 0.06 (D) where (D) is 75 meters. Substituting this value of (k) in equation (17) gives

18. $$S = \frac{0{,}186\, V_0\, D}{\sqrt{2\,\pi}} = 0{,}0418\, V_0\, D, \text{ and}$$

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[1] This approximation is true only for small values of ($k\,\pi$), the error being less than 3 % if ($k\,\pi$) is less than 0.5.

19. $$x^1 = \frac{S}{y} = 3.1 \frac{V_0}{\pi \sqrt{2}}.$$

Substituting for (V_0) its value in terms of the wind velocity from equation (8) we have

20. $$x^1 = \frac{3,1\,(,0057)}{\pi \sqrt{2} \sqrt{\sin \Phi}} V_W = \frac{0,004}{\sqrt{\sin \Phi}} V_W$$

where (x^1) is the average velocity in meters per second at which a surface layer of thickness 5 meters leaves the coast and (V_W) is the component of the wind velocity parallel to the coast in miles per hour.

Multiplying by the number of seconds in a month we have

21. $$x = \frac{10480}{\sqrt{\sin \Phi}} V_W.$$

Substitute this value of (x) in equation (13) and the result is

22. $$t = \frac{10480\,V_W}{\sqrt{\sin \Phi}\left(x_1 + \frac{1}{2} x_2\right)} (t_2 - t_1)$$

which is a theoretical relation between (t) the reduction of the temperature below the normal value, the wind velocity (V_W) parallel to the coast, the normal temperature (t_2), and the temperature (t_1) of the upwelling water which causes the abnormally low actual temperature (T). For any given place the latitude (Φ) is constant, and observation shows (t_1), (x_1), and (x_2) to be practically constant. So the only variables for a given locality are the wind velocity parallel to the coast and the normal temperature. The wind velocity must be in miles per hour and the distances (x_1) and (x_2) in meters.

VI. The application of the above theory to four selected regions, and the comparison of the observed and computed values.

Four stations on the coast were selected in which the factors entering into the computation differed widely. For (t_1) the value 8° was used north of latitude 36° and 9° was used for the region south, but the value of $\left(x_1 + \frac{1}{2} x_2\right)$ was chosen so as to give the best agreement between the computed and observed values, but was assumed constant for each statien. It seemed reasonable to use a value of (t_1) somewhat greater than the bottom temperature, as the water would become warmer as it rose and mixed with the layers above. The value chosen is about the average of the mean annual surface temperature and the bottom temperature. It is the actual value usually found at the depth of 500 meters.

Using for (V_W) the average wind velocity for the month's time just preceeding the middle of the month to which the mean temperatures (t_2) and (T) correspond, the value of (t_2) for the latitude, and a constant value of $\left(x_1 + \frac{1}{2} x_2\right)$ determined by trial, the temperature difference (t) and the actual temperature (T) of the in-shore water were calculated for each month of the year. And the results were compared with the observed values. The values of (V_W) were calculated[1] from the U. S. Coast Pilot Chart records,[2] and the temperatures were taken from Thorado's maps, on which the surface isotherms for each month of the year are plotted between the coast and longitude 140°, and between the latitudes 20° and 50°.

In the following tabulation of results San Diego is denoted by (1), Point Conception by (2), San Francisco by (3) and Cape Mendocino by (4). The same results are shown graphically by (Figs. 8, 9, 10 and 11).

Station No. 1, Lat. 32° 45′, Long. 118°, $\sqrt{\sin \Phi} = 0{,}736$, $t_1 = 9°$, $\left(x_1 + \frac{1}{2} x_2\right) = 915$ Kilometers, $t = 0{,}017\, V_W (t_2 - t_1)$

Month	V_W	t_2	$(t_2 - t_1)$	calculated t	observed t	calculated T	observed T	Differences
1	11,80	16,90	7,90	1,60	1,90	15,30	15,00	0,30
2	13,45	16,60	7,60	1,75	1,20	14,85	15,40	—0,55
3	14,10	15,80	6,80	1,65	0,20	14,15	15,60	—1,45
4	13,20	15,20	6,20	1,40	—0,10	13,80	15,30	—1,50
5	15,40	17,50	8.50	2,20	3,20	15 30	14,30	1,00
6	16,90	17,50	8,50	2,45	0,90	15,05	16.60	—1,55
7	18,20	20,90	11,90	3,70	4,90	17,20	16,00	1,20
8	18,40	21,90	12,90	4,00	3,90	17,90	18,00	—0,10
9	17.60	21,50	12,50	3,75	4,10	17,75	17,40	0,35
10	16,90	20,10	11,10	3,20	2,90	16,90	17,20	—0,30
11	15,55	18,70	9,70	2,55	2,20	16,15	16,50	—0,35
12	14,15	17,75	8,75	2,10	1,55	15,65	16,20	0,45

Average value of all of the observed values of t, 2,23
" " " the three greatest " " t, 4,30.

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[1] The frequency, or number of hours per hundred during which the average wind velocity in each 5 degree square has the direction N, NNE, NE, etc. for each month is denoted by the length of an arrow pointing in the direction of the wind. The average force of the wind is expressed in Beaufort's scale. The velocity in miles per hour is given by the formula $V = (5F) + 3$, where F is the force in Beaufort's scale. (See bulletin, Instructions to Marine Meteorological Observers of the U. S. Weather Bureau, by James Page for a table showing the relation of (F)

Station No. 2, Lat. 35°, Long. 121°, $\sqrt{\sin \Phi} = 0{,}758$, $t_1 = 9^0$, $\left(x_1 + \frac{1}{2} x_2\right) = 615$ Kilometers, $t = 0{,}022\, V_W (t_2 - t_1)$

Month	V_w	t_2	$(t_2 - t_1)$	calculated t	observed t	calculated T	observed T	Differen-ces
1	9,70	15,80	6,80	1,45	2,10	14,35	13,70	0,65
2	11,30	15,20	6,20	1,55	1,10	13,65	14,10	—0,45
3	12,50	14,00	5,00	1,35	0,60	12,65	13,40	—0,75
4	12,55	13,40	4,40	1,20	0,80	12,20	12,60	—0,40
5	14,35	16,70	7,70	2,45	4,50	14,25	12,20	2,05
6	16,05	16,40	7,40	2,60	1,30	13,80	15,10	—1.30
7	18,25	20,50	11,50	4,60	4,90	15,90	15,60	0,30
8	18,20	21,20	12,50	5,00	5,10	16,20	16,10	0,10
9	17,55	21,00	12.00	4,65	5,00	16,35	16.00	0,35
10	15,80	19,40	10,40	3,60	3,40	15,80	16.00	—0.20
11	13,30	18,00	9,00	2,65	2,80	15,35	15,20	0,15
12	11,50	16,50	7,50	1,90	1,90	14,60	14,60	0,00

Average value of all of the observed values of (t), 2,79.
" " " the three greatest " " (t), 5,00.

Station No. 3, Lat. 37°, 45′ Long. 123°, $\sqrt{\sin \Phi} = 0{,}792$, $t_1 = 8^0$, $\left(x_1 + \frac{1}{2} x_2\right) = 440$ Kilometers, $t = 0{,}030\, V_W (t_2 - t_1)$.

Month	V_w	t_2	$(t_2 - t_1)$	calculated t	observed t	calculated T	observed T	Differen-ces
1	7,70	14,20	6,20	1,40	1,70	12,80	12,50	0,30
2	6,87	13,80	5.80	1,20	2,20	12,60	11,60	1,00
3	9,25	12,60	4,60	1,30	1,10	11,30	11,50	—0,20
4	11,40	12,00	4,00	1,40	0,70	10,60	11,30	—0,70
5	12,50	14.50	6,50	2,40	3,20	12,10	11,30	0,80
6	14,40	15,20	7,20	3,10	1,40	12,10	13,80	—1,70
7	17,50	20,00	12,00	6,30	6,50	13,70	13,50	0,20
8	18,60	19,90	11,90	6,65	6,90	13,25	13,00	0,25
9	17,60	19,90	11,90	6,30	6,10	13,60	13,80	—0,20
10	13,60	18,60	10,60	4,30	4,00	14,30	14,60	—0,30
11	8,80	16,60	8,60	2,30	2,80	14,30	13,90	0,50
12	6.65	15,20	7,20	1,40	1,80	13,80	13,40	0,40

Average of all the observed values of t, 3,20.
" " the three greatest " " t, 6,50.

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to (V)). The average monthly value of the component (V_w) parallel to the coast was found by resolving each velocity into directions perpendicular and parallel to the coast and adding the products of all the latter components by their corresponding frequencies and dividing this sum by the sum of the frequencies.

[2] The values of (V_w) were obtained only from the observations in the first square west of the corresponding station, the centers of the squares used being about 400 or 500 kilometers from land. Only values of (V_w) during the time from 1908 to 1911 were used.

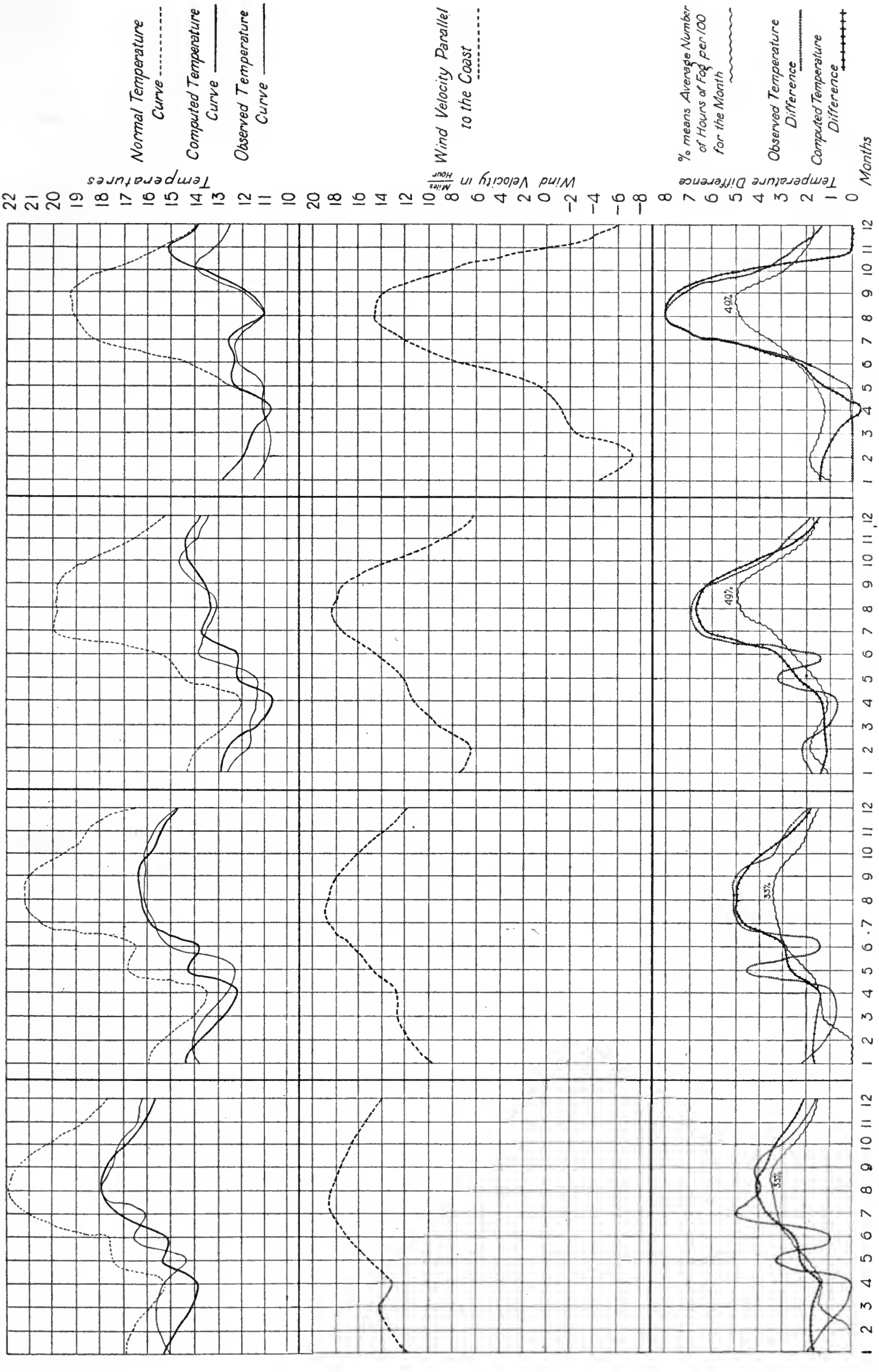

Fig. 8.

Fig. 9.

Fig. 10.

Fig. 11.

Station No. 4, Lat. 40°, Long. 124°, $\sqrt{\sin \Phi} = 0{,}802$ $t_1 = 8^0$, $\left(x_1 + \frac{1}{2} x_2\right) = 263$ Kilometers, $t = 0{,}0497 (V_W)(t_2 - t_1)$.

Month[1]	V_W	t_2	$(t_2 - t_1)$	calculated t	observed t	calculated T	observed T	Differences
1	—4,9	12,60	4,60	0,00	1,30	12,60	11.30	1,30
2	—6,40	11,80	3,80	0,00	1,10	11,80	10,70	1,10
3	—2,40	11,40	3.40	0,00	0,70	11,40	10,70	0,70
4	0,15	10,60	2,60	0,00	—0,40	10,60	11,00	—0,40
5	1,30	12,40	4,40	0,20	1,40	12,20	11,00	1.20
6	6,95	14,30	6,30	2,15	2,30	12,15	12,00	0,15
7	11.60	18,40	10,40	6,00	6,40	12,40	12,00	0,40
8	14,50	19,10	11,10	8,00	8,00	11,10	11,10	0,00
9	13,55	19,20	11,20	7,55	7.20	11,65	12,00	—0,35
10	8,35	17,70	9,70	4,10	3 70	13,60	14,00	—0,40
11	—1,85	15,40	7,40	0.00	2,40	15.40	13,00	2,40
12	—6,30	13,80	5,80	0,00	1,40	13.80	12,40	1,40

Average of all of the observed values of t, 2,98.
" " the three greatest " " t, 7,20.

VII. Discussion of the methods and results, and additional tests of the theory.

The assumptions (page 257) on which the computation of the surface temperatures were based are not in accord with the actual conditions, first, because the direction of the coast is not north and south except at station (4), second, because the flow is not normal to the coast as assumed, but, as will be shown, is directed toward the south west at angles varying from (0°) to (90°) with the coast. To justify the method used in computing the temperatures, and to explain various peculiarities of temperature distribution, a more detailed consideration of the distribution and magnitude of the currents deduced from Ekman's theory will be required. But the necessary mathematics for such a quantitative study can be avoided and a qualitative explanation of some of the results can be worked out by assuming an ideal set of conditions and then estimating the change in the result due to such a modification of these assumed conditions as will secure an agreement with the actual circumstances observed at the different regions.

If a steady wind uniform in direction and velocity blows parallel to a long straight coast, and the depth of the water is constant and

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[1] The sign (—) means that the component (V_W) is directed to the north, in all other cases it is directed to the south.

greater than (2 D), the actual velocity at any point of the surface current is the resultant of the velocity (U_0) (problem 2, page 253) and the velocity in a surface layer which the corresponding local wind would produce in an ocean extending to a great distance from the point in all directions, according to problem (1), page (251).

But along the Pacific coast, the wind direction, except during a few weeks in the winter is toward the south east, and makes an angle of about (25^0) with the coast within about (300) kilometers from the shore. Therefore the angle that the surface velocity (V) would make with the coast would be (45^0) — (25^0) = (20^0) if (U_0) were zero, and it can be shown the actual value of (U_0) next to the coast reduces this to (10^0). But the wind direction shifts to the west at points to the south and west of Cape Mendocino, and makes an angle of about (45^0) to the west of a north and south line at the latitude of (32^0) and longitude (140^0) in the summer. Therefore, we would expect (U_0) to diminish and the surface velocity to be directed more and more to the west as the distance from the coast increases and the latitude diminishes.

Therefere a stream-line (a line having at each point the direction of the motion at that point) starting from the vicinity of Cape Mendocino would be about parallel to the coast there, but would continually turn to the right as the latitude diminished. A series of such stream-lines starting from points equidistant from each other along the coast would divide the water area into a series of sections, such that the water upwelling into each one would come from the ocean bottom underneath the part next to the coast, and no surface water could intrude from other regions. Observation shows that the cooling effect of the upwelling water diminishes as the distance from the coast along a stream-line increases, and in the following discussion, we will assume for the outer boundary of any section a line parallel to that part of the coast bounding the opposite end, and at such a distance that the temperature is practically normal. Now the cooling effect in an east and west area such as was considered on page (257) could be determined by computing the amount of water upwelling into each section that crosses it, and then calculating the temperature of the part of each section included in this area.

From the appearance of the isotherms, the actual as well as the normal temperature of any section increases in proportion to the distance from the coast along the corresponding stream lines, therefore the assumptions stated on page (257) except as to direction of flow, correspond to the conditions in a section included between two stream-lines, and the method described on page (259) will give the value of (T), the actual temperature near the coast, if the mean value of the normal temperature were used for (t_2).

Thus it would be possible to obtain the value of (T) from the section that started where (T) was desired, and the temperature of an area running east and west could be found from the temperatures of the inclined sections that cross it.

Now consider two of these inclined sections, one terminating at the outer boundary of the east and west area, and the other beginning at its inner boundary. Assume the outer boundaries of the three areas to be portions of a line parallel to the coast, then if the breadths of these areas measured parallel to this line are equal, the areas will be equal, the mean temperature of the middle area will be approximately the average of the temperatures of the inclined sections, the mean wind velocity over the middle area will be approximately the average of the wind velocities over the inclined sections. Therefere if the normal temperature of the east and west area is assumed to be reduced only by the intrusion of bottom water underneath it whose amount is computed as on page (259), where the wind velocity corresponding to that latitude is used, the result would be approximately the same as the more complicated but pertinent method just explained.

Also the distance ($x_1 + x_2$) is multiplied by the perpendicular distance between the bounding lines of the east and west surface to obtain the area, but the distance (x) should be multiplied by the length of the coast included by these bounding lines, since it is the quantity flowing normal to the coast that is assumed to enter the first area. Therefore the numerator of equation (13) and also of (22) should be divided by ($\cos \alpha$) where (α) is the angle the coast makes with a north and south line, and consequently the denominator $\left(x_1 + \frac{1}{2} x_2\right)$ determined as on page (259) would be increased in the same proportion. Thus the use of formula (22) for computing temperatures is justified, but we must divide the quantity $\left(x_1 + \frac{1}{2} x_2\right)$ determined as on page (259), by ($\cos \alpha$) in order to obtain the correct distance out to where the temperature is the average of (T) and (t_2) for the latitude. And if another relation between (x_1) and (x_2) can be found, then we can calculate ($x_1 + x_2$) the distance west to where the cooling effect is practically negligible.

Under the uniform conditions assumed on page (262), if the coast is vertical the upwelling would occur in a narrow belt as shown by (Figs. 7 and 12). And the motion except within a distance from the coast of the same order of magnitude as (D), would be the same as that described on page (262), problem (3). If, instead of being level, the bottom slopes gradually upward toward the coast, the motion would be but little changed till the depth diminished from the value (2 D) (see note, page 255). From there toward the shore, the flow normal to the coast in the surface layer would continually diminish, and consequently the pressure gradient corresponding to the slope of the water surface and the flow toward the coast in the bottom layer would decrease as the coast was approached. If the depth is less than 0,5 (D), the flow is nearly parallel to the coast, and the surface is horizontal. Therefore (Fig. 13) the upwelling would begin where the depth is 2 (D) and would continue thruout the region from there till the depth had decreased to 0,5 (D), instead of being confined to a narrow region adjacent to the shore.

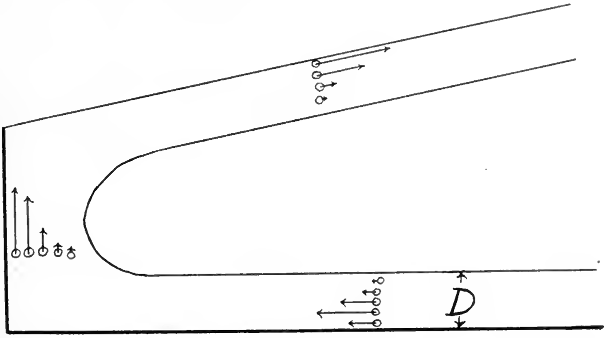

Fig. 12.

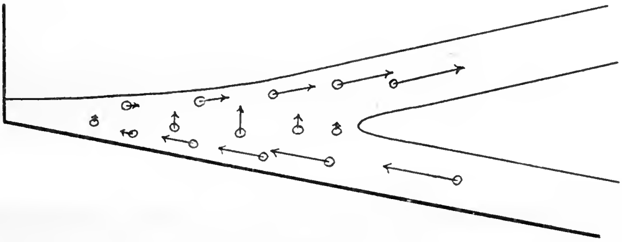

Fig. 13.

The arrows represent the average flow at the points ⊙.
Cross sections normal to the coast.

That is, the greater the inclination of the bottom, in the neighborhood of a coast, or the more rapidly the depth increases with the distance from the coast, the narrower will be the belt near the coast into which a given volume of cold water intrudes, and therefore the greater will be the reduction of the surface temperature.

For example, at the head of a submarine valley terminated at the coast, a lower surface temperature than that on either side would result from a wind so directed as to produce upwelling.

Another factor affecting the upwelling is the steadiness of the winds. The effect of the winds over the broad and deep regions, requiring several months to become fully established, can not follow the seasonal change in the wind velocity, but tends to agree with that due to the average velocity[1]), and approaches more nearly to that average amount the more steady the wind. But the effect in narrower belts, and in the shallower water of the continental shelves, requiring but a few hours or days to become practically fully devoloped would change more nearly in accordance with the monthly variation of the wind velocity.

We can now test the theory further by comparing the results of the general conclusions just reached with the corresponding observations. An estimate based upon the U. S. Coast Pilot Chart records of the breadth of the wind belt was made for each station and these distances were plotted as ordinates against distances from station (1) as abscissas (Fig. 14, curve 2). That is, this line which practically touches the coast at latitude (50°), (500) miles north of Cape Mendocino marks the approximate western limit of winds whose velocities have a north and south component. Between the coast and this line the winds tend more and more to the west as the distance from the coast increases. Therefore the system of streamlines due to the observed wind velocities that would be expected from the theory would be limited to a narrow area corresponding to the narrow wind belt off Cape Mendocino but would spread out to the west and south, as the latitude diminished and the wind direction shifted to the west. Thus a general surface drift toward the south and bearing more and more to the west would

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[1]) Owing to the large value of the effective internal friction of the water which tends to prevent the circulation, the motion would decrease nearly in accordance with the decrease of wind velocity, but would not increase to the maximum value which the wind velocity would generate if it continued for a sufficient length of time. Therefore, the actual effect of a varying wind would be less than that due to a steady wind of the same value as the average of the changing winds.

result. This is in good agreement with observations[16] (See also Figs. 20 and 21).

Also the upwelling of cold water, which we have seen is confined to a narrow belt next to the coast and depends on the component parallel to the coast of the local wind velocities would commence south of latitude (50°). And, as this cold water is carried out from the coast along the stream lines, we would expect the cold-water area to increase in width as the latitude diminishes.

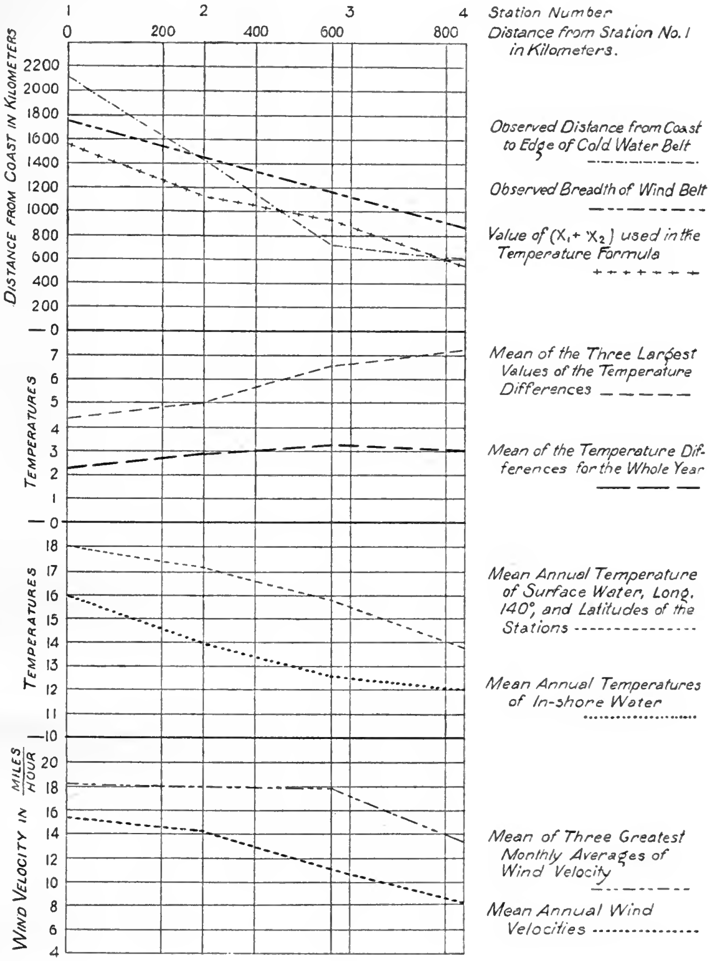

Fig. 14.

The estimated deviations of the directions of the coast from a north and south line at stations 1, 2, 3, and 4 are respectively (35^0), (30^0), (25^0), and (0^0) to the north west, and for stations 1 and 2, $(x_2) = (2x_1)$, but for stations 3 and 4, $(x_1) = 0$. From this data, and the values of $\left(x_1 + \frac{1}{2} x_2\right)$ obtained from formula (22) the corresponding distances $(x_1 + x_2)$ were found to be 1670, 1120, 910, and 525 kilometers. An estimate based upon the isothermes of Thorade's charts, of the distances out from the coast to where the cooling effect was negligible gave the following series of values: 2150, 1450, 710, and 600 kilometers. In (Fig. 14, curves 3 and 1) these results are shown graphically. These curves 1 and 3 would coincide if the theory and the observations were correct. And the observed increase in width of the cold-water area agrees well with the results deduced from the observed winds.

In (Fig. 15) are shown the contours of the ocean bottom perpendicular to the coast at each station. Recalling the effect on temperature distribution of the slope of the ocean bottom, and of the depth adjacent to the coast, we would expect the variation in the character of these contours to give rise to an increasing temperature reduction, as the distance north of station (1) increases, as is shown by (Fig. 14, curve 4).

Finally, from the general system of winds over the North Pacific Ocean we would expect from Ekman's theory, a "midwater-current" bordering the coasts of Japan, Asia, and North America, having a clockwise motion corresponding to that of the wind, the surface flow being continually directed toward mid-ocean. Thus a movement of water in a belt several hundred kilometers in width from the equatorial regions north along the western borders of the Pacific then to the east and south along the North American coast is in harmony with Ekman's theory. But a cold arctic surface-current, if started southward would be deflected to the right of its path and would therefore flow away from the coast of North America. Hence, such a current could not give rise to a cooling effect in-shore.

Considering the complexity of the circulation and temperature distribution and the many factors influencing them, it is rather surprising that the computed temperatures should agree so well with the observed ones at each of the stations where the conditions differed so widely, even tho the method used has been shown to give approximately the same results as tho the actual character of the flow was considered. Also the general agreement of the deductions of the theory with the peculiarities of the temperature distribution, such as the breadth of the cold-water belt, the magnitude of the reduction of the temperature below the normal, the location of the coldest area etc. is very satisfactory. So it seems, on the whole that the rather severe test of the theory has proved it to be at least approximately correct and to be well worthy of a detailed quantitative investigation.

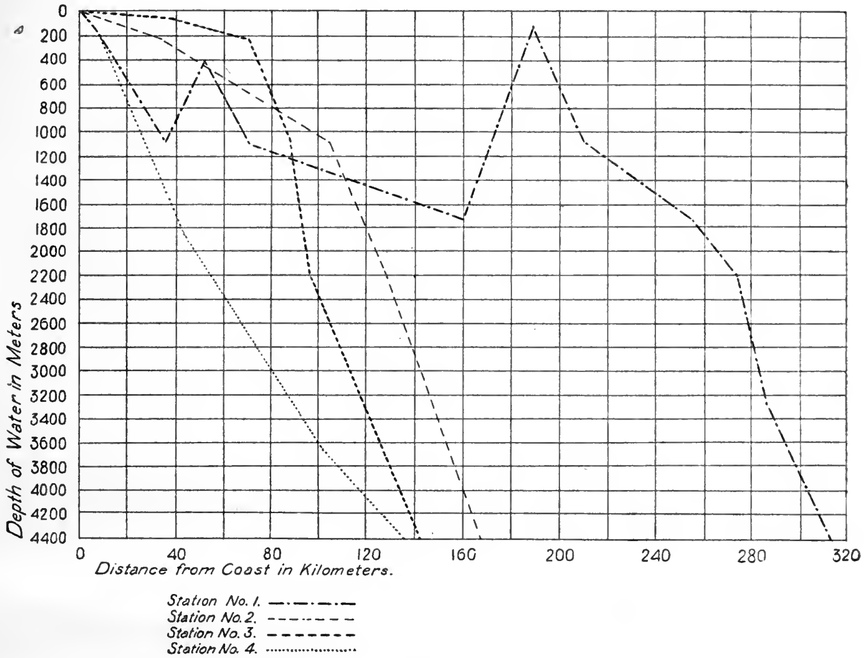

Fig. 15.

Leaving the consideration of the temperature distribution on a large scale, some results of observations confined to a limited area will now be examined in the light of the new theory. During the summer months for several years numerous temperature observations (17) close to the coast from latitude 28° to latitude 34° have indicated abnormally cold water, several temperatures being from 13° to 14°. A rapid fall of 2° or 3° has been occasionally noted. For example (18) in the summer of 1903, of the Coronado pier repeated observations taken at 10 A. M. indicated a temperature of 20°, but on July 27 the value 16,2° was observed, and continued for three days, after which the former value was resumed. Also, regions of alternately warm and cold water on a line parallel to the coast have been repeatedly observed (17) similar to those much farther north mentioned by Holway (9). And the lowest temperatures were found just in the regions where the maximum upwelling would be expected. The irregularities in temperature distribution may be accounted for by recalling the factors that we have found influenced the distribution on a large scale. The fluctuations with respect to time would result from a change in the wind, and the variations with respect to location would result from differences in the wind, irregularities in the depth and slope of the ocean bottom, and in the presence of submarine valleys.

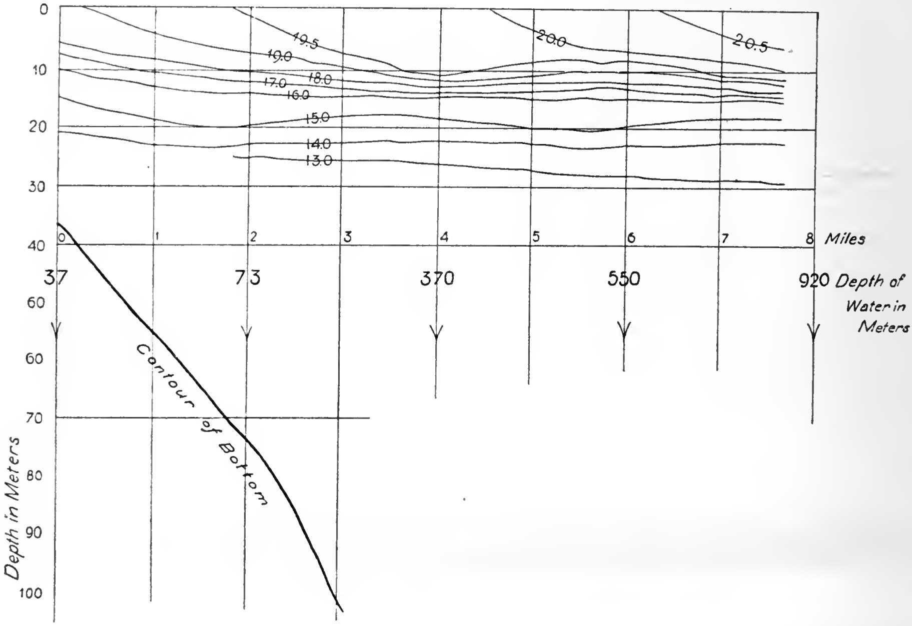

Fig. 16.

Isotherms for a Course from South Coronado Island eight miles west August, 1910.

Several lines of serial temperature measurements[1] have been made running from the coast out the distances of 10 or 20 miles. The results of two of these sets of observations shown by (Figs. 16 and 17) are typical of the others. During these observations the direction of

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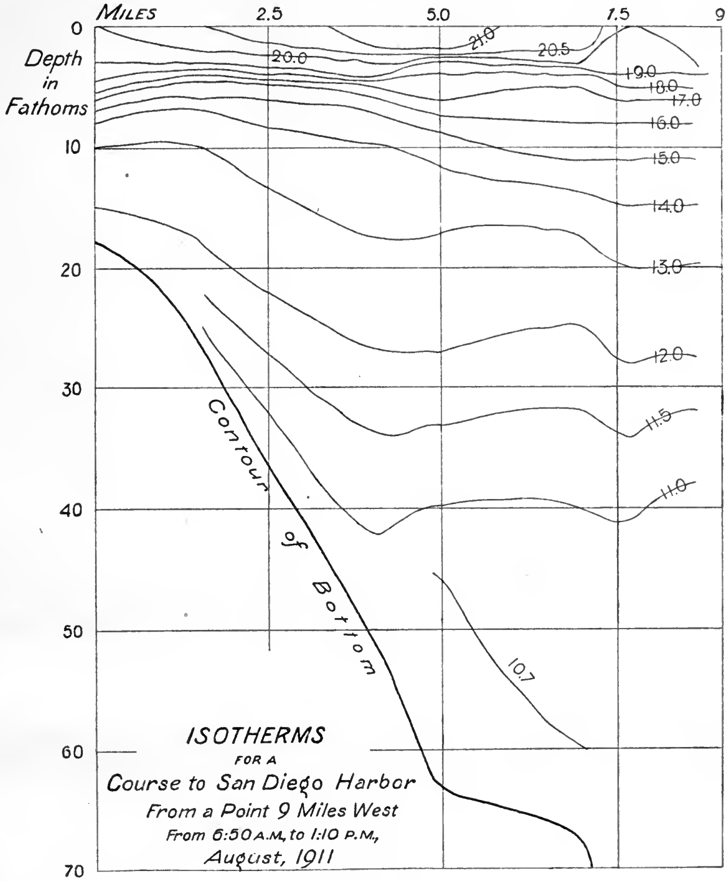
[1] The following data on ocean temperatures was copied from unpublished records on file at the laboratory of the Marine Biological Association of San Diego at La Jolla, California.

the wind was so related to the coast that upwelling would be expected. Now if the bottom water rises, and at the same time flows toward the coast, we would expect to find colder water near the coast than that farther out at the same level, as cold water is rising to a region of a normally higher temperature. Therefore, the upward deflection of the isotherms shown by the diagrams agrees with what would be expected from the theory.



ISOTHERMS
FOR A
Course to San Diego Harbor
From a Point 9 Miles West
From 6:50 A.M. to 1:10 P.M.,
August, 1911

Fig. 17.

The peculiar shape of the isotherms near the surface (Fig. 17) is explained as follows. The observations were commenced at 6:50 A.M. when the surface temperature is normally below the daily average. A rise in temperature at a given place would continue till about 2 P.M., because of the increase in the sun's heating effect. But along this course the rise continued only till about 10 A. M. when the temperature began to fall as the coast was approached till the same temperature 20° was found at 1 P. M., nearly the warmest part of the day as was normal for early morning 8 miles out to sea. Thus as the day advanced and the heating effect of the sun increased, the positions of the points of observation were nearer the coast where the maximum upwelling occurs, so the temperature distribution shown is the resultant of two effects, while the actual distribution at any given instant would be more like that shown by (Fig. 16).

An interesting and peculiar distribution of surface temperatures was observed during August 1911, adjacent to the coast and islands in the neighborhood of Point Conception. From Santa Barbara to Point Conception the coast runs nearly due west, but trends slightly to the north. On August 18, 6 A. M. the surface temperature near the coast 12 miles east of Point Conception was 18,1°, but fell to 15,8° at 9 A. M. within 3 miles of the Point. This low temperature would result from an intrusion of cold surface water driven by the local winds to the southeast from the region west of the Point where the necessary conditions for considerable upwelling prevail.

On a line running 25 miles southeast of Santa Rosa Island an average surface temperature of 16° was observed, the value farthest out being 15,1° at 7 P. M., Aug. 21. And at the latter position a strong wind was blowing toward the southeast about parallel to the contours of the ocean bottom, while the depth was about 1500 meters, so the low temperature must have been due to upwelling which would evidently result under such conditions, if Ekman's theory is correct.

Near the coast (within 1 or 2 miles) between Point Dume and Point Mugh, latitude 34° the low temperature of 14,2° was observed on Aug. 17, 6:30 A. M., the temperatures being higher both to the north and to the south. The rise of temperature per mile of distance north or south was a maximum near the place where the actual temperature was a minimum. A temperature of 20° was noted both at Santa Barbara and at San Pedro each city being about 55 miles away.

Now this region of minimum temperatures opens freely to the southwest, and is at the head of a submarine valley, the depth being 900 meters only 5 miles out, and increasing to 1800 meters 40 miles out. The deepest portion runs to the southwest of the coast.

Thus we would expect a concentration of the upwelling water, and the correspondingly greater temperature reduction, observed. (See page 264.)

Another peculiar temperature phenomenon was observed on Aug. 19, 1911, in the Santa Barbara Channel between Santa Rosa Island and the coast, latitude 34°, longitude 120°. The mean temperature of the channel water was about 18°, and values as high as 20,25° were observed near the northern coast of Santa Cruz Island. But on entering Beecher's Bay on the northeast side of Santa Rosa Island a strong off-shore wind was encountered, and the temperature began to fall, reaching the low value of 14,8° close to shore. Now as the bay was shallow about 35 meters deep on the average, the wind would tend to drive the surface water in its own direction[1]) (away from the shore in this case) and thus cause a diminution of pressure which would give rise to an upwelling of the cold bottom water. Similar phenomena have been observed in shallow lakes (19). Thus Ekman's theory is in harmony with shallow water circulation, as well as that in the deep ocean.

VIII. The influence of ocean temperatures on the coast climate.

In general, the most important factor controlling climate is the latitude, since it is the factor that has most to do with the geographical distribution of the heat received from the sun, upon which directly or indirectly, all meteorological processes depend. But the actual climate at any given place is the combined result of various factors, and may be quite different from the normal for the latitude.

On the west coast of North America the prevailing land wind blows from the northwest except for a few weeks in the winter. We would expect from this fact that between the ocean and the mountains, the air temperature would be largely controlled by that of the adjacent surface water of the ocean. An examination of the temperature diagrams for San Diego and San Francisco (Figs. 18 and 19), shows a close agreement between the air and the local surface temperatures of the ocean, resulting in the remarkably uniform air temperature, entirely different from the normal for the same latitude (Figs. 1 and 2).

An examination of the air isotherms (15) corresponding to the

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[1]) See page 255.

summer months for the coast region shows the average temperature to depend upon the distance east of the coast rather than on the latitude,

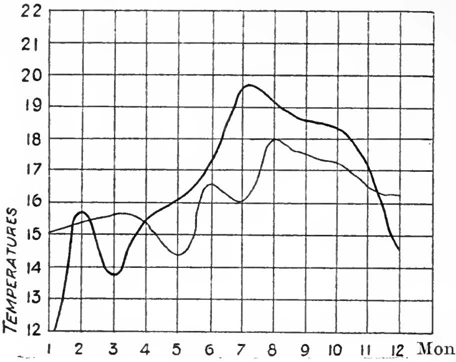

Fig. 18.

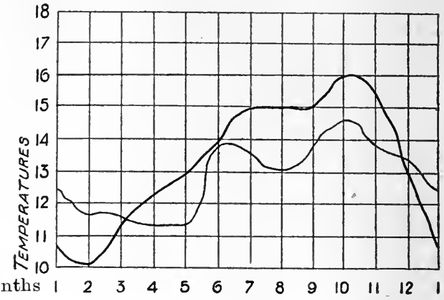

Fig. 19.

— Ocean Temperatures at San Diego.
— Air Temperature at San Diego.

— Ocean Temperatures at San Francisco.
— Air Temperatures at San Francisco.

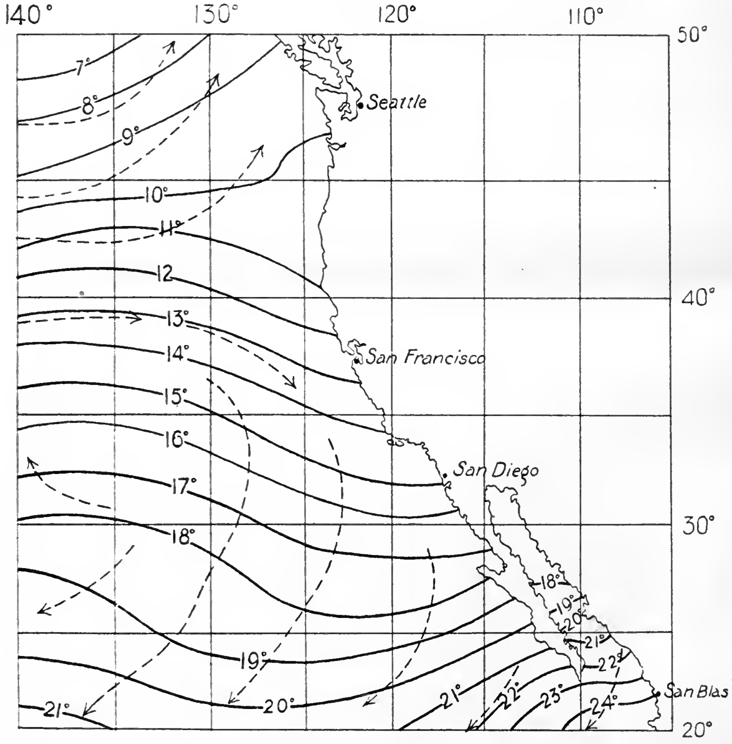

Fig. 20. Surface Isotherms for January. ←--- Prevailing Surface Drift.

between latitude 32° and latitude 50°. Therefore, the presence of cold coast water, which we have seen must be due to an upwelling from the ocean depths, practically governs the coast climate.

Since, except for a few weeks in the winter the wind blows toward the land as well as parallel to the coast from over the ocean area, and consequently passes over water areas of a successively lower temperature as the coast is approached, there will be a condensation of the water vapor of the saturated air. Thus the formation of fogs, so prevalent along the Pacific Coast may be due in part to the cold water bordering the coast. If we assume the percentage of fogs for any given month, that is, the average number of hours of fog per hundred in that time, to be due entirely to the above cause, we would expect this percentage to change in proportion to the temperature difference (t).

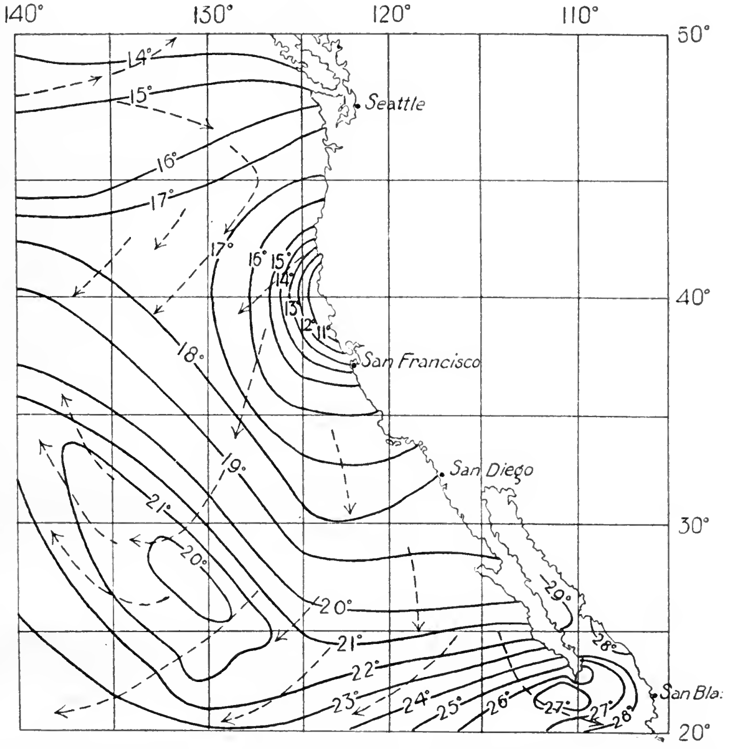

Fig. 21. Surface Isotherms for August. ←--- Prevailling Surface Drift.

From the results of a large number of observations on the frequency of fogs McAdie (20) has made a table showing the percentage of fogs in the region south of San Francisco and also in the region north of there, so we have the necessary data for testing the above conclusion. The fog curves (Figs. 8, 9, 10, and 11) representing this data have been drawn in connection with the corresponding curves showing the temperature differences. The resulting similarity of the fog curves and the corresponding temperature curves is just what would be expected from the above assumption regarding the way in which the fogs are formed.

An extensive investigation of the influence of ocean temperatures and local winds on climate has been carried on by many meteorologists since the publication in 1817 of Humboldt's chart of The Isothermal Lines of the Earth, and it was suggested by Richter (5) in 1887, that a knowledge of the isothermal lines of the ocean along the coast would probably explain the formation of our barometric maxima and minima, and afford a basis for the calculations of meteorologists, and for more accurate weather predictions than would otherwise be possible.

IX. Summary and conclusions.

Numerous observations extending over a long period have established the presence of abnormally cold surface water contiguous to the west coast of North America, but a diversity of conflicting theories have been proposed by various writers to account for the phenomenon.

The conclusions reached by different investigators may be summarised as follows.

1. A cold arctic current flows south along the coast from the polar regions.

2. The Japan Current, because of its passage thru high latitudes, becomes cooled, and as it flows south along the coast of the United States, appears as a cold stream because its temperature corresponds to the normal value prevailing in higher latitudes

3. The accumulation of water in the south polar region causes an excess of pressure which drives the cold bottom water northward with an increasing velocity owing to the diminishing distance across the Pacific, till when it reaches the latitude of Sitka, Alaska, owing to the deflecting force due to the earth's rotation it is driven up the continental slope and flows south as a cold current, since it has no other outlet.

4. The coldest water is located about 800 miles south of Sitka in the summer time, and areas of alternately warm and cold water are distributed in an irregular manner all along the coast. But from each of the previous theories, owing to the continual increase in the heating effect of the sun toward the south, a continuous rise in temperature would accompany a decrease of latitude. Therefore the low temperature must result from an upwelling of cold bottom water from the adjacent ocean depths. A general eastward drift of the ocean water extending to the bottom is assumed to result from the continued action of the winds, consequently the cold bottom water is driven up the continental slope, most of it reaching the surface at Cape Mendocino (the coldest region). The irregularities in temperature distribution are due to the effects of submarine valleys and differences in the slope of the ocean bottom.

The above theories were based on hypothetical causes, which in some cases were not verified except by the general qualitative agreement of the deductions with the particular observations considered, and the theory of oceanic circulation proposed in 1878 by Zöppritz was widely used. No attempt was made to explain the seasonal fluctuation.

Before going on with the conclusions regarding the Pacific coast region it will be necessary to consider general theories of oceanic circulation. A recent one due to Ekman differs from that of Zöppritz, in that no assumption as to regular flow in plane layers is used as a basis, but a virtual value of the coefficient of viscosity, allowing for the actual turbulent motion of the water is used, and the deflecting force due to earth's rotation is also introduced. Many results of Zöppritz's theory are inconsistent with observations, while those of Ekman's theory are in harmony with experience. Most of the results of the two theories are entirely different.

From Ekman's theory it follows that there must be an upwelling of the cold bottom water along most of the coast of North America owing to the action of the observed winds, and in the present paper, assuming the low temperature to be due entirely to cold bottom water upwelling and mixing with the surface water a theoretical formula was derived by which the abnormally low temperatures of any region could be computed for each month of the year. A very satisfactory agreement with observations was obtained, tho the temperature reduction below the normal varied from 0° to 8°.

In general the theory shows that the area affected and the magnitude of the temperature reduction and its distribution vary with the depth of the water, the slope of the bottom, the velocity of the winds, the portion of the surface over which they extend and the steadiness of the winds.

To give an idea of the peculiarities of temperature distribution that have been accounted for by means of these principles the following results of observation are enumerated.

The cooling effect of the upwelling water extends to a distance of 600 kilometers from the coast off Cape Mendocino, lat. 40° and increases to a distance of 2100 kilometers from the shore off San Diego, lat. 32° 45′.

The temperature reduction in the summer is a minimum off San Diego and a maximum off Cape Mendocino where the coldest surface water is found.

Temperatures as low as 14° in August have been found in certain limited areas near the coast south of lat. 35°, while the value 18° prevailed in the surrounding water a few miles away, both north and south.

Considering the complexity of the phenomena, the agreement between the theory and the observations has been very satisfactory, and judging from the results already obtained it would be profitable to carry on a more detailed and quantitative investigation following the lines suggested in the present paper.

The University of Illinois, U. S. A.
Feb. 8, 1912.

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Kurze Zusammenfassung der Arbeit von G. C. McEwen über: Ozeanische Temperaturen an der Westküste von Nordamerika (Dr. Wendicke, Berlin).

Durch zahlreiche langjährige Beobachtungen an der Westküste Nordamerikas hat sich ergeben, daß in diesen Gegenden in der Regel abnorm niedrige Temperaturen des Oberflächenwassers angetroffen werden. Von verschiedenen Seiten ist der Versuch gemacht worden, die Erscheinung dieses kalten Küstenwassers zu erklären. Aber so viel Autoren es gibt, die sich an diese Arbeit machten, so viel verschiedene und sich widersprechende Hypothesen wurden aufgestellt. Die ganzen Ergebnisse der verschiedenen Untersuchungen lassen sich ungefähr folgendermaßen zusammenfassen.

1. Eine kalte arktische Strömung fließt aus Polarregionen an der nordamerikanischen Küste entlang nach Süden.

2. Die japanische Strömung, die sich auf ihrer Reise durch hohe Breiten abkühlt, wendet sich an der Küste der Vereinigten Staaten nach Süden. Da nun ihre Temperaturen sich den mittleren Werten höherer Breiten angepaßt haben, so wird sie bei ihrem Vordringen nach Süden zu kalt erscheinen.

3. In den Südpolargebieten erhalten die Wassermassen, die durch den großen antarktischen Stromring dort angestaut werden, durch starke Abkühlung eine größere Dichte. Es entsteht hier also aus doppelter Ursache ein beträchtlicher Überdruck, der kaltes Bodenwasser nordwärts treibt. Je weiter diese Bodenströmung über den Äquator nach Norden dringt, um so schmaler wird ihr Querschnitt und daher ihre

Geschwindigkeit um so größer. Auf der geographischen Breite von Sitka, Alaska, wird sie schließlich durch die Wirkung der Erdrotation bestimmt werden, nach rechts auf den Kontinentalshelf hinaufzutreten und muß dann, da kein anderer Ausweg vorhanden ist, an der amerikanischen Küste südwärts strömen.

4. Das kälteste Wasser befindet sich im Sommer ungefähr 800 Meilen südlich von Sitka, und auf der ganzen Küstenstrecke befinden sich bald warme, bald kalte Wassergebiete in unregelmäßiger Verteilung. Nach allen bisherigen Theorien mußte nun aber entsprechend der durch die Kontinentnähe verstärkten Wirkung der Sonnenstrahlung eine nach Süden gerichtete Strömung mit abnehmender Breite eine ständige Temperaturzunahme erfahren. Die niedrigen Temperaturen lassen sich also nur durch das Aufquellen kalten Bodenwassers aus benachbarten ozeanischen Tiefen erklären. Zu diesem Zweck wird angenommen, daß durch ständige Windwirkung eine allgemeine Bodentrift ozeanischen Wassers nach Osten erzeugt wird. Am Kontinentalabfall wird daher kaltes Bodenwasser hinaufgedrängt werden und hauptsächlich bei Kap Mendocino, dem kältesten Gebiete, die Oberfläche erreichen. Die beobachteten Unregelmäßigkeiten in der Temperaturverteilung werden offenbar durch submarine Täler und Ungleichförmigkeit des Kontinentalabfalls verursacht.

Alle soeben angeführten Theorien zur Erklärung der ozeanischen Temperaturen an der Westküste von Nordamerika stützten sich auf hypothetische Annahmen, für die fast nur die allgemeine qualitative Übereinstimmung der Folgerungen mit den seltsamen Erscheinungen sprach. Auch die Theorie über ozeanische Zirkulationen, die 1878 von Zöppritz vorgeschlagen wurden, genügten nicht, und es wurde kein Versuch gemacht, die jährlichen Schwankungen der Isothermen dieses Gebietes zu erklären.

Erst die neuen Theorien von Ekman, der der inneren Reibung und der ablenkenden Kraft der Erdrotation Rechnung trägt, stimmen mit vielen Beobachtungstatsachen überein, bei denen die Zöppritzsche Theorie vollkommen versagt, und erst auf Grund dieser Ekmanschen Theorie ist es möglich die Temperaturerscheinungen des pazifischen Küstengebietes zu erklären.

Nach der Ekmanschen Theorie muß durch die Wirkung der beobachteten Winde an der Westküste von Nordamerika kaltes Bodenwasser aufsteigen. McEwen leitet nun unter der Annahme, daß die beobachteten niedrigen Temperaturen ausschließlich durch kaltes aufsteigendes Bodenwasser, das sich mit dem Oberflächenwasser mischt, verursacht werden, eine theoretische Formel für die sich hierbei ergebenden Temperaturen ab. Das Ergebnis ist sehr günstig, denn obwohl die beobachteten Temperaturen der verschiedenen Monate des Jahres zwischen 0° und 8° unter den Normalwerten dieser Breiten lagen, so stimmten trotzdem die errechneten Werte mit den beobachteten recht gut überein.

Allgemein ergab sich, daß die Temperaturerniedrigung und ihre Verteilung zu folgenden Faktoren in enger Beziehung stehen: der Wassertiefe, der Bodenneigung, der Windgeschwindigkeit, seiner Beständigkeit und seiner Verbreitung.

Um von den Besonderheiten der Temperaturverteilung, um die es sich bei den obigen Berechnungen handelte, eine Vorstellung zu geben, werden folgende Beobachtungstatsachen aufgezählt.

Die abkühlende Wirkung des aufsteigenden Wassers macht sich bei Kap Mendocino (40° nördl. Breite) noch in 600 Kilometer Abstand von der Küste bemerkbar, bei San Diego (32° 45′ nördl. Breite) aber sogar bis auf 2100 Kilometer Entfernung.

Die Temperaturerniedrigung hat im Sommer ihr Minimum in der Höhe von San Diego und ihr Maximum in der Höhe von Kap Mendocino, wo das kälteste Oberflächenwasser gefunden wird.

Temperaturen von nur 14° wurden im August in bestimmten begrenzten Gebieten nahe der Küste südlich von 35° nördl. Breite gefunden, während nördlich und südlich davon in wenigen Meilen Entfernung schon Temperaturen von 18° vorherrschten.

In Anbetracht der verwickelten Erscheinungen ist jedenfalls die Übereinstimmung von Theorie und Beobachtung sehr zufriedenstellend, und ein weiteres Vordringen in der eingeschlagenen Richtung verspricht neue nutzbringende Ergebnisse.

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