ROSSBY WAVES IN THE OCEAN

by

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In the atmosphere, it is clear that in the middle and upper troposphere, the synoptic charts for any large region of zonal motion systems show relatively great simplicity with smooth, wave-shaped patterns rather than the irregular closed systems which are seen in the lower layer.

This theoretical consideration was given by Rossby (1939)[1], who expressed it as a kind of inertial wave with vorticity due to the variation of the latitude of the Coriolis parameter. These waves are well known as Rossby waves or long waves.

According to this theory, in the ocean, too, it is conceivable that such waves exist, though examples of them have not been shown, because of many difficulties to ascertain from the present observation data.

It seems, however, possible to generate Rossby waves in the Antarctic circumpolar ocean where there are the same hydrodynamic conditions as in the westerly zonal wind systems.

Also, there are some other conditions which interest us such as ocean currents which do not circulate around a pole such as the North Pacific or the Alaska current systems.

The author obtained wave-shaped patterns of the iso-σ_t lines and isotherms by making synoptic charts near the Alaska current with vorticity in the upper layer from observations made in the summer of 1955.

At present, the law or the empiric rule of the formation of wave patterns on such a large scale cannot be defined, because there are many factors in the ocean. The author, therefore, studies long wave patterns in the ocean, in relation to the Rossby waves. The results are insufficient because of a lack of data, but the wave patterns are full of suggestions just as the Rossby waves. Considering the important role of Rossby waves in present weather forecasting, the author studied Rossby waves in the ocean and prepared charts of long wave patterns in the Alaska current.

Characteristics of Rossby Waves in the Ocean

In order to define the characteristics of the Rossby waves in the ocean, it is necessary to examine the derivation of the formula in the atmosphere. The fundamental formula and the conditions of deriving the result are described briefly.

Taking the rectangular co-ordinate, with the x-axis directed towards the east, and the y-axis towards the north, let u and v denote the component of the current in the directions of axes of x and y, respectively. Using the general current symbols, we assume the vertical motion is much smaller than the horizontal, and the fundamental equations of the motion are as follows:

$$\left.\begin{aligned}\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}-fv&=-\frac{1}{\rho}\frac{\partial p}{\partial x}\\ \frac{\partial v}{\partial t}+u\frac{\partial v}{\partial y}+v\frac{\partial v}{\partial y}+fu&=-\frac{1}{\rho}\frac{\partial p}{\partial y}\end{aligned}\right\} \quad \cdots\cdots \quad (1)$$

where $f=2\omega\sin\varphi=$ Coriolis parameter

$$\text{let}\ \zeta=\frac{\partial v}{\partial x}-\frac{\partial u}{\partial y} \quad \cdots\cdots \quad (2)$$

So that from (1), (2)

$$\frac{d(\zeta+f)}{dt}+(\zeta+f)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)=0 \quad \cdots\cdots \quad (3)$$

This is the vorticity equation for horizontal motion. If the effect of divergence is neglected, we have

$$\frac{d(\zeta+f)}{dt}=0 \qquad \text{or}$$

$$\frac{\partial \zeta}{\partial t}+u\frac{\partial \zeta}{\partial x}+v\frac{\partial \zeta}{\partial y}+v\frac{\partial f}{\partial y}=0 \quad \cdots\cdots \quad (4)$$

Assume that a zonal current of uniform constant velocity U upon which is superimposed a perturbation with the velocity component u' and v'. That is

$$u=U+u' \qquad v=v' \quad \cdots\cdots \quad (5)$$

Neglect $u'\frac{\partial \zeta}{\partial x}+v'\frac{\partial \zeta}{\partial y}$, as they are small terms of the second order, and from (4), (5),

$$\frac{\partial \zeta}{\partial t}+U\frac{\partial \zeta}{\partial x}+\beta v'=0 \qquad (6)$$

where $\beta=\frac{\partial f}{\partial y}$ = Rossby parameter

Now take the stream function ψ, and u', v', ζ are as follows:

$$u'=-\frac{\partial \psi}{\partial y},\quad v'=\frac{\partial \psi}{\partial x},\quad \zeta=\nabla^2\psi \qquad (7)$$

Next assume that the eastward current of the uniform constant velocity U upon which is superimposed sinusoidal waves. Then a solution is assumed in the form

$$\psi=A\cos\frac{2\pi}{D}y\sin\frac{2\pi}{L}(x-Ct) \qquad (8)$$

where A is an arbitrary constant, D represents the width of the disturbance, L, its wave length and C, the velocity of disturbance. From (6), (7), (8), it follows that

$$U-C=\frac{\beta L^2}{4\pi^2}\frac{D^2}{D^2+L^2} \qquad (9)$$

This is the formula derived by Haurwitz (1940)[2], but if D is much larger than L, from (9)

$$C=U-\frac{\beta L^2}{4\pi^2} \qquad (10)$$

This is the well-known formula derived by Rossby (1939)[1]. Now let L_s denote the stationary wave length, and from (10),

$$L_s=2\pi\sqrt{\frac{U}{\beta}} \qquad (11)$$

and then, from (10) and (11)

$$C=\frac{\beta}{4\pi^2}(L_s^2-L^2) \qquad (12)$$

Accordingly in large-scale ocean currents as in air currents, we can use equation (9) or (10) if the divergence is neglected. This means that if the divergence is negligible, the absolute vorticity is conserved from equation (3). In the Antarctic circumpolar ocean current, the trajectories of the current may show the same patterns as those of westerly winds so that in such regions Rossby's formula may be applicable.

But the general flow of an ocean current is much smaller than that of currents in the atmosphere, and its maximum velocity is about 1 m/sec. The magnitudes of the wave lengths in the ocean and in the atmosphere, therefore, are very different.

In the ocean the stationary wave length, L_s, defined by the velocity of the general current from (11) in comparison with that[1] of the atmosphere is shown in Table 1.

Table 1. Stationary wave length (km) as functions of zonal velocity (U) and latutide (φ)

U(m/sec) / φ(°)	Ocean				Atmosphere		
	0.05	0.1	0.5	1	4	12	20
30	315	446	998	1411	2822	4888	6310
45	349	494	1105	1561	3120	5405	6978
60	415	587	1313	1857	3713	7432	8304

According to *The Oceans* (Sverdrup et al)[3], the surface current velocity is about 15 cm/sec in the region of the north of the Antarctic Convergence, and 4.4 cm/sec within the arctic region. So that, now take latitude $\varphi=55°$, and $U=10$ cm/sec, the stationary wave length is about 550 km.

Now if n is the number of waves around the hemisphere, and R is the radius of the earth, then $2\pi R\cos\varphi=nL$, so that from (10)

$$U-C=\frac{2\omega R\cos^3\varphi}{n^2} \qquad (13)$$

Then under the condition of the above mentioned Antarctic circumpolar current, the number of waves, n, is 42, while in the atmosphers, if the wind velocity is 5 m/sec, n is only 6. From the results of the Discovery Expedition (Deacon, 1937 a), we find that the number of the trough of the Antarctic Convergence line is about 13. This may not show the number of Rossby waves because the charts are not made by simultaneous observations as are those of the atmosphere, and the distances between observation stations are too large for accurate verification. But if we assume that they are Rossby waves, and $\varphi=55°$, $n=13$, then from (11) the value of U—C is 103 cm/sec.

In order to see the movement of Rossby waves, the relation between the wave length of a trajectory and its current velocity is derived from considering the constant absolute vorticity by Rossby[4]. The essential point to apply in the ocean is described as follows.

If V and θ are the speed and the direction of the current, respectively, we have for the velocity components $u=V\cos\theta$, $v=V\sin\theta$, where θ is counted from the x-axis. If we now choose the x-axis tangent to a streamline, $\theta=0$, and $\frac{\partial\theta}{\partial x}=K_s$ the curvature of the streamline. Furthermore, if s measures length along the steamline and n measures length along the normal to the left of the streamline, then we obtain the following expression for the vorticity.

$$\zeta=\frac{\partial v}{\partial x}-\frac{\partial u}{\partial y}=VK_s-\frac{\partial V}{\partial n} \quad\cdots\cdots\cdots\cdots\cdots\cdots (14)$$

Next assume that absolute vorticity is conserved, and rewrite (4) using the natural coordinate system, then

$$\frac{\partial(\zeta+f)}{\partial t}+V\frac{\partial(\zeta+f)}{\partial s}=0 \quad\cdots\cdots\cdots\cdots (15)$$

In this equation, the first term is usually much smaller than the second, so that the former is neglected. Then

$$\frac{\partial(\zeta+f)}{\partial s}=0, \quad \text{or} \quad \zeta+f=\text{const.} \quad\cdots\cdots (16)$$

from (14), (16)

$$f+VK_s-\frac{\partial V}{\partial n}=\text{const.} \quad\cdots\cdots\cdots\cdots\cdots\cdots (17)$$

Now we choose such a point that the origin of the natural co-ordinate is the inflexion point of the streamline that goes through the core of the current. And then at this point $K_s=0$, and $\frac{\partial V}{\partial n}=0$, so that the absolute vorticity is equal to the local Coriolis parameter. Next assume that the shear remains negligibly small while the parcel moves through the wave pattern. Then, from (17)

$$K_sV+f=f_0 \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots (18)$$

where f_0 is the Coriolis parameter at the origin.

Now θ is the current directin, so that $\frac{d\theta}{dt}=VK_t$=angular speed. If $\frac{\partial\theta}{\partial t}$ is much smaller than $V\frac{\partial\theta}{\partial s}=VK_s$, then

$$VK_s=VK_t \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots (19)$$

from (18), (19)

$$K_tV=f_0-f=-\beta y \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots (20)$$

where β is Rossby parameter and may be regarded as constant.

Next assume the motion to be balanced, so that V is constant along the trajectory. Now the general expression for the curvature K_s is

$$K_s=\frac{d^2y}{dx^2}\left[1+\left(\frac{dy}{dx}\right)^2\right]^{-\frac{3}{2}} \quad\cdots\cdots\cdots\cdots\cdots\cdots (21)$$

If the amplitude is small, $\frac{dy}{dx}$ may be neglected in comparison with unity, so that from (19), (20), (21), the solution may be written

$$y=A_t\sin\sqrt{\frac{\beta}{V}} \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots (22)$$

The trajectory is therefore a sinusoidal curve, and its wave length L_t

$$L_t=2\pi\sqrt{\frac{V}{\beta}} \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots (23)$$

So that the wave length of the trajectory is the same as that of the stationary Rossby wave. Also the amplitude A_t is obtained as follows:

$$A_t=\tan\theta_0\sqrt{\frac{V}{\beta}} \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots (24)$$

where θ_0 is the angle from the x-axis to the current direction at the inflexion point ($x=0$).

Next let us consider the Rossby waves in the oceanic circular current which is not around the pole but bounded on both east and west sides by a continent as the North Pacific or the Alaska current systems.

In the atmosphere the solution[5] under nearly all the conditions mentioned is given, so we may summarize its principal consideration to apply it to solve the problems in the ocean.

The starting point is equation (3). Now the solution is assumed as follows:

$$y=y_0+A_s\sin 2\pi\frac{x}{L} \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots (25)$$

where A_s and L may be functions of time. Furthermore, the waves need not be zonal. So that the angle γ defines the tilt of the wave: it is counted positive from the north toward the west. Also assume that the motion is symetrical in the vicinity of the trough and wedge lines. Now, if C is the speed of the vorticity wedge (or though), then

$$C=-\frac{\partial^2\zeta/\partial x\partial t}{\partial^2\zeta/\partial x^2} \quad\cdots\cdots\cdots\cdots\cdots\cdots\cdots\cdots (26)$$

And we consider only the conditions at the part of trough and wedge lines where the ocean current is a maximum. If U is this maximum (i. e. the core speed), then

$$\frac{\partial u}{\partial y}=0 \quad \text{and} \quad u=U \qquad (27)$$

If B is the half-width of the current,

$$\frac{\partial^2 u}{\partial y^2}=-\frac{U}{B^2} \qquad (28)$$

and also if the current is not divergent, from (3), (25)—(28)

$$C=U-\frac{\beta\cos\gamma+U/B^2}{(2\pi/L)^2+1/B^2} \qquad (29)$$

or rewrite (29), then

$$C=\frac{U-\beta\cos\gamma(L/2\pi)^2}{1+(L/2\pi B)^2} \qquad (30)$$

So that from (30) stationary wave length L_s is

$$L_s=2\pi\sqrt{\frac{U}{\beta\cos\gamma}} \qquad (31)$$

which, apart from the influence of the tilt, is the same as Rossby's formula (11).

When the current runs from east to west, it is convenient to choose the x-axis toward west. With this choice of axis, $\beta<0$, and from (30) all waves are progressive.

According to the above consideration, the Rossby wave may exist in general circular ocean currents as in atmospheric ones. Next the charts which seem to show Rossby waves in the ocean are shown as follows.

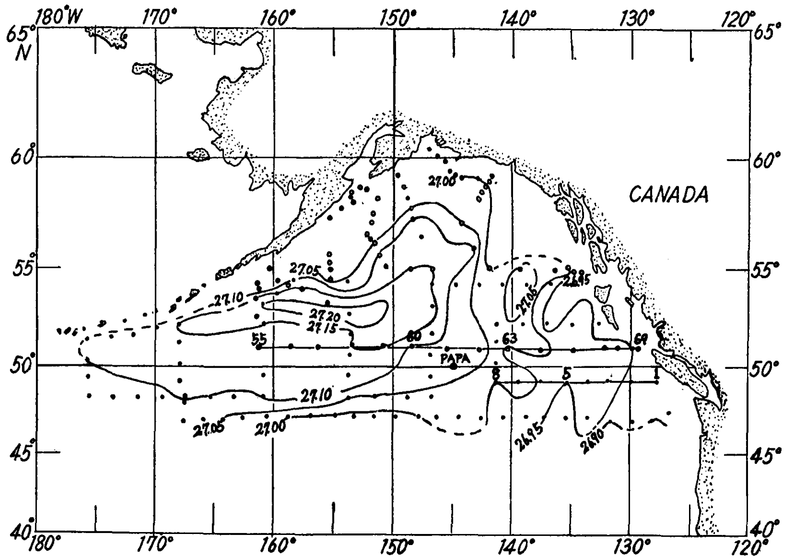

Fig. 1 Location of hydrogrphic stations: black spots show the Canadian stations by Ste. Therese, white spots show the U.S.A. stations by Brown Bear: and contour lines show the horizontal distribution of σ_t at a depth of 400m in summer 1955.

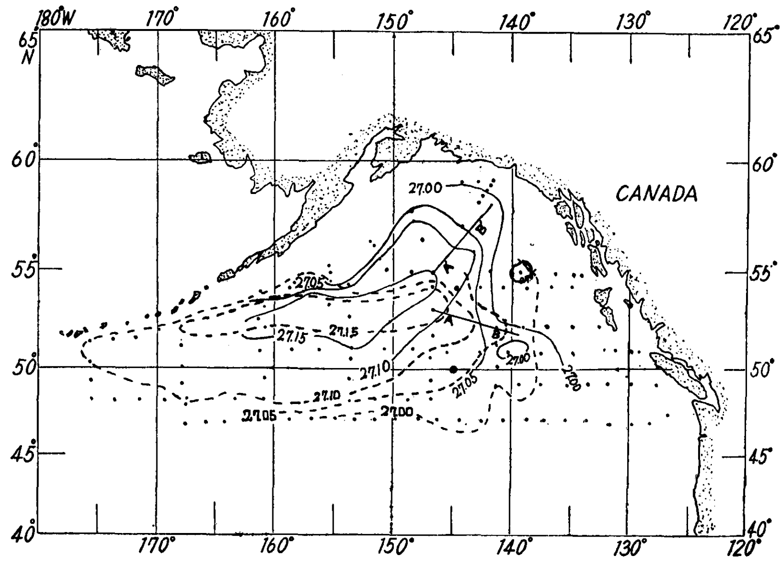

Fig. 2 Horizontal distribution of σ_t at a depth of 400m: broken lines show those of Canadian observations, solid lines show those of U. S. A. obsereva-tions.

Wave Patterns in the Alaska Current

The data used to make the charts are as follows (1) observations of the Northern Pacific Ocean (NORPAC) by the Canadian Joint Committee on Oceanography[6] (abbreviated: Canada), from July 26 to Sept. 1, 1955, and (2) those of the Department of Oceanography, University of Washington[7] (abbreviated: U. S. A.) from Aug. 1 to Sept. 19, 1955. The locations of the hydrographic stations are shown in Fig. 1, the black spots denoting the former and the white spots the latter. Now in this area the general ocean current is contra solem, and the pattern shown in the published surface current charts[3], [8] is roughly the same as that of Fig. 1, which shows the iso-σ_t curves at a depth of 400 m.

Fig. 1 is made from the total data of both Canada and U. S. A. mentioned above during the entire observation period, while Fig. 2 shows the σ_t-curves of Fig. 1 separately, that is, the broken lines denotes the former and the solid line, the latter.

Fig. 3 (a)

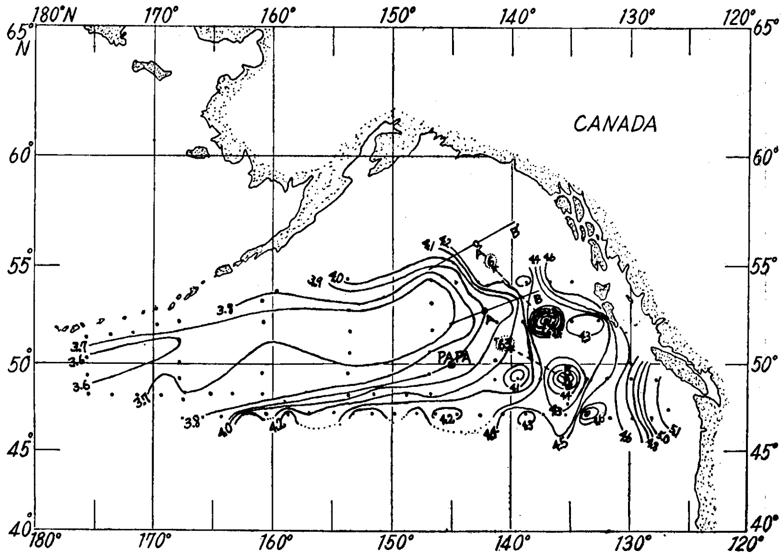

Fig. 3 (b)

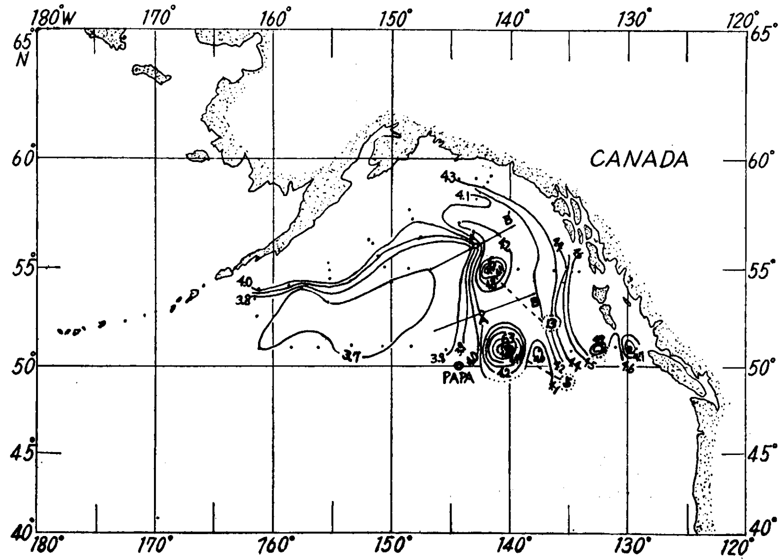

Fig. 3 Horizontal distribution of temperature (C°) at a depth of 400 m: (a) Canadian observations, (b) U. S. A. observations.

In regard to hydrodynamics in a stationary statuation, the current flows in such a direction that the water of low σ_t is on the right-hand side of the current, and the water of high σ_t is on the left hand side in the Northern Hemisphere.

Strictly speaking, a stationary situation does not exist, but assuming that a stationary condition is hold for a short period, the σ_t-curves of Canada and U. S. A. in Fig. 2 shows this stationary situation. Then in Fig. 2, the solid σ_t-curves are different from the broken σ_t-curves. That is, the seize marked difference in the pattern of σ_t-curves between Long. 140°–150°W, and Lat. 52°—57°N, and the position of the trough shows the phase difference shown by the straight lines AB and A′B′. In this region the observation period of the broken line is July 31 and Aug. 3—5, that of the solid line is Aug. 8—22, so they may show general characteristics, though the latter curves may show some distortion. The horizontal distribution of the temperature at the 400 m depth is shown in Fig. 3, and in this figure (a) and (b) are those of Canada and U. S. A., respectively. The positions and the phase difference of the trough of these wave-shaped isotherms are the same as those of σ_t-curves. But the isotherms show here and there closed curves, most of which have a higher temperature in the center. These closed curves may show the vortex of water. If these closed curves denote vortex, the magnitude of the core temperature value and the shape of the vortex of station No. 5 in Fig. 3 (a) may show the same as that of station No. 63 in Fig. 3 (b).

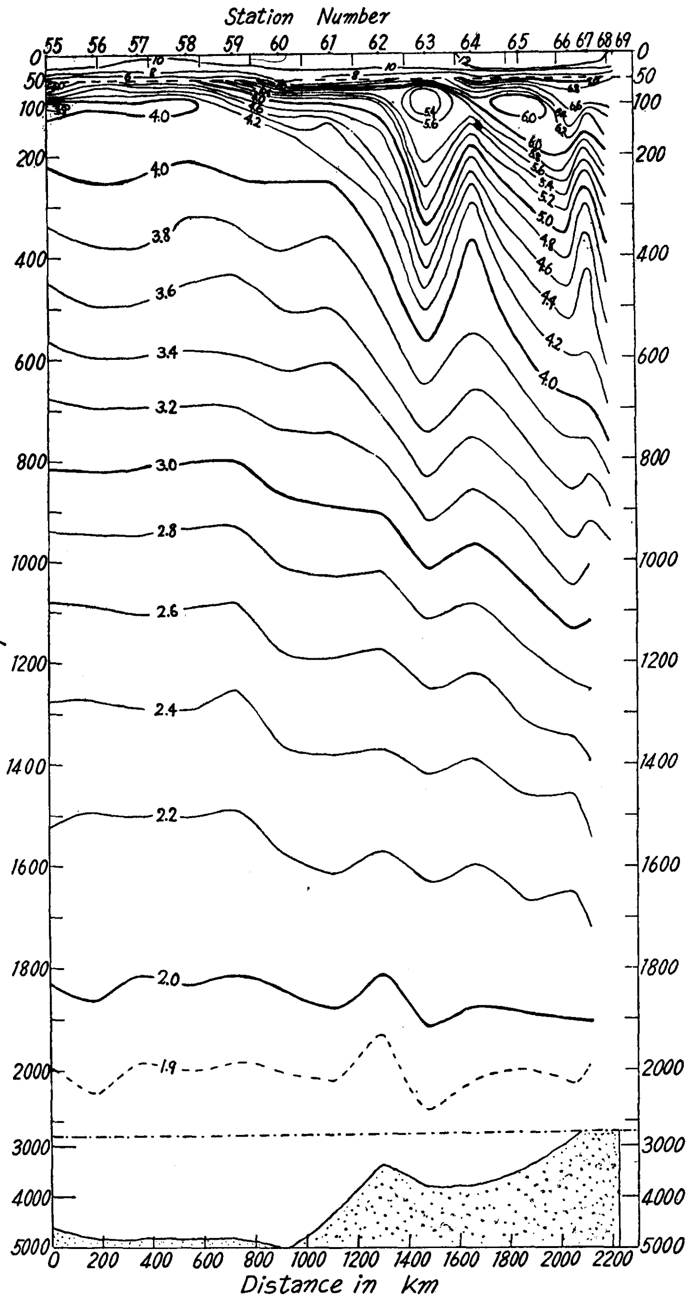

Fig. 4 Distribution of temperature (C°) in a vertical section shown by line 55-63-69 in Fig. 1.

From the data of the observations, the former is on July 29 and the latter is on Sept. 9, so the time difference is 42 days, and the current may be directed from No. 5 to No. 63. So that if these vortices are the same, since the distance between these two positions is about 370 km, the average speed is about 10 cm/sec. Next, if No. 13 in Fig. 3 (a) is the same as No. 6 in Fig. 3 (b), from the difference in the observation dates and the distance between the positions, the computed average speed is about 70 cm/sec.

The vertical distribution of temperature along the observation line, No. 55—No. 69 of U. S. A. (Fig. 1) is shown in Fig. 4. In this figure the closed isotherms are found at a depth of nearly 100 m at No. 63, and the temperature gradient of its upper and lower sides is shown in reverse. The characteristics of this figure show wave-shaped patterns in both the horizontal and vertical directions, especially marked near No. 63. This shows that such phenomena are caused by some large disturbance.

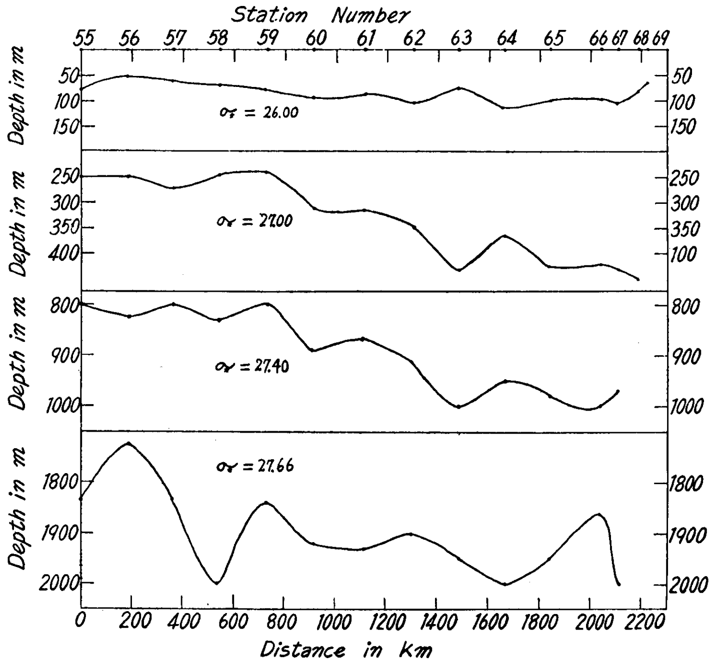

Fig. 5. Variation of the depth of σ_t=constant in a vertical section shown by the line 55-63-69 in Fig. 1.

The variation of the depth of σ_t=constant is shown in Fig. 5, and it represents also the wave-shaped characteristics mentioned above. In Fig. 4 the horizontal distances between the adjacent two troughs drawn by the 3.8°C isotherm (denoting the wave length of the 3.8°C isotherm) are from the left, 560, 580 and 550 km, those of the 2.0°C isotherm, 370, 560 and 380 km, and those of the $\sigma_t=27.40$ curves, 360, 370, 570 and 560 km. So that the wave length of the wave-shaped patterns is of two kinds, one about 370 km, the other 560 km.

If these waves denote the stationary wave length, the corresponding zonal velocity, from (11) are 5 cm/sec, and 11 cm/sec respectively. Also if these waves denote the wavelength of the trajectory, these represnt the core velocity of the trajectories from (23).

The surface current velocity by dynamical calculation (relative to the 1000 db surface) from the Canadian data is about 10 cm/sec in the strongest region, and decreases with depth.

So that if the zonal velocity is 10 cm/sec, and the wave length L=370 km and L=560 km, from (12) the wave Speed, C is 5 cm/sec and -1 cm/sec respectively.

In Fig. 3 (a), the above defined angle of the tilt of wave, at the position of the trough of the 3.9°C isotherm is 60°, (with due consideration for the current direction and co-ordinate system) so that from (30), if U is 10 cm/sec, the stationary wave length is about 760 km.

From the difference in observation dates and the distance AA′ between the two trought AB and A′B′ in Fig. 3 (a), the average moving speed of the trough is about 37 cm/sec.

Discussion

In the previous section, the Rossby waves are only considired, but such wave patterns may be generated by atmospheric disturbance on a large scale. It is necessary, therefore, to consider the atmospheric conditions from the surface weather charts in the Alaska area. But they could not be obtained.

The data of the atmospheric conditions used are as follows: (1) and (2) are the shipboard meteorological observation records from Canada and U. S. A. mentioned above. (3) are the meteorological records from another Canadian observation ship at the fixed station "PAPA" (Lat. 50°,00 N, Long. 145°,00 W.) : the observation period and times are Aug. 1—Oct. 31 twice a day.

In regard to the atmospheric conditions, it could not be found that the generation of the vortex mentioned above is related to the extrimum of the atmospheric pressure or wind. Then we considered the wind effect.

The observation period near the trough of σ_t-curves show that A′B′ in Fig. 2 was Aug. 8—22 and the wind velocity was 8 m/sec on the first day and its direction was south-southeasterly, while that of another trough denotes that AB was July 31 and Aug. 3—5 and the wind velocity was sometimes about 9 m/sec and its direction was north-north-westerly after being south-southwesterly. According to these facts, it may be said that the wind acts on the formation of the phase difference of the trough. If the atmospheric conditions are represented by those of "PAPA", the prevailing wind is westerly or west-southwesterly. Assuming that the vortex of No. 5 is the same as that of No. 63, the moving speed may be reduced by wind, while the vortex of No. 13 may be accelerated by wind because the wind of position No. 6 on Aug. 8 was south-southeasterly against "PAPA".

The movement of the vortex mentioned above may be conducted by the general current, as the circular vortex in the atmosphere is conducted (for instance cyclone). The relation between the drift current velocity V (cm/sec) and wind velocity W (cm/sec) is $V=\frac{0.013\,W}{\sqrt{\sin\varphi}}$, where φ is the geographic latitude.

If we take $\varphi=50°$, $W=6$ m/sec from this formula, V is 9 cm/sec. Furthermore if the wind velocity is 9 m/sec, the drift current velocity reaches 12 cm/sec. So that it is larger than the general ocean current velocity, 10 cm/sec as previously assumed. But in the actual state the wind direction is not constant, though the individual wind strength is not strong enough to disarrange the general ocean current which is decided by the average value of wind velocity and direction for a long time and the boundary conditions.

Next there is no lower limit of the velocity of the general current to generate the Rossby waves, as seen in the previous derivation. So that the general circular ocean current has always the Rossby waves, but if its magnitude is small, it is difficult to ascertain the waves, as they are usually affected by small disturbances.

Next the vortex may be generated from the Rossby wave patterns. As seen in Fig. 3, there are many vortices near the parts of the northward flow of the Alaska current. In such a region, the variation of the latitude for the unit distance along the trajectory is larger than other parts.

Therefore in this region, if the current velocity, V is almost constant, as seen in (18), (19) the curvature K_t must be larger than other parts. While in the bending portion where the current is directed from east to north-north-west, the curvature of the trajectory must undergo much change. As the action of the above two factors is independent, in compliance with the combination of these facters, the bending of the trajectory may be often intensified. Consequently the bending of the trajectory becomes large, till a part of it may be cut off, and a vortex may be generated.

The observation period of the line from No. 55 to No. 69 is 9 days, and if the velocity of the general current is 10cm/sec, the distance moved during that period is less than 80km. So that the distortion due to nonsimultaneous observations is not large enough to be significant when compared with the stationary wave length. Accordingly, the ocean observation period for the confirmation of the Rossby waves may not be simultaneous, but must depend on the velocity of the general current.

In the region of the Alaska current, as above mentioned, the computed surface current velocity is at most 10 cm/sec. It is weak, and difficult to obtain exatctly the core velocity from the data of the above observation interval shown in Fig. 1.

As mentioned above, from the observation data of the isotherms near trough A in Fig. 3 (a), $C=37$ cm/sec, $\gamma=60°$ and $B=150$ km are deduced. Using them and $L=370$ km, and 560 km, from (30), U is 40 cm/sec and 47 cm/sec respectively. These values are 4 or 5 times larger than the general current velocity (10 cm/sec) used before. This does not mean that there are no Rossby waves in the ocean, but the actual ocean curent velocity in any layer may be much larger than 10cm/sec, for we find the difference between the theory and its practical application in the air (for instance, L and C are almost constant in the 700—200 mb layer, while the value of U often changes 2 or 4 times in this layer).

From theoretical considerations, it is possible that Rossby waves are possible to exist in the ocean, but evidence is still inadequate, not only because the observation data are insufficient, but also because the waves are affected by other factors.

Finally in the ocean, if Rossby waves exist, it may be a good indicator of the prediction of the state of the ocean current, and affect the physical characteristics of deep water, especially its movements.

Summary

1. In the ocean it is difficult to prove the presence of Rossby waves from the present data.

2. On theoretical considerations, it is possible for Rossby waves to exist in the ocean with the following characteristics.

(a) Rossby waves are disturbances on a large scale caused by circular ocean currents.

(b) Considering the distribution of the ocean on the earth and the velocity of the general current, the wave length of stationary Rossby waves is from several hundred km, to one thousand and several hundred km.

(c) In the ocean, areas capable of generating Rossby waves seem to be in the Antarctic circumpolar ocean.

(d) The oceanic circular current on a large scale which does not run round the pole may also generate Rossby waves.

3. In the regions of the Alaska current, the horizontal and vertical long wave patterns showing the iso-σ_t curves and isotherms may suggest Rossby waves, although data are still insufficient for proof.

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