A Servomechanism in the Ocean/Atmosphere System of the Mid-Latitude North Pacific [1]

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ABSTRACT

During the autumn and winter seasons, large amounts of heat are given up to the atmosphere at the subarctic frontal zone off the east coast of Asia. According to Fisher, the Laplacian of this heat flux ($\nabla^2 Q$) is related to increases in the intensity of the relative vorticity in the westerly wind regime. This increase is related to a similar increase in strength of the wind stress curl, which thereby increases the Sverdrup transport of the subarctic and subtropical gyres. The increase in transport in turn intensifies $\nabla^2 Q$ at the subarctic frontal zone via geostrophic adjustment. This coupling of atmospheric relative vorticity, Sverdrup transport, and $\nabla^2 Q$ results in the intensification of the relative vorticity of both fluid media that can be checked only by an instability in either one or the other media. This mutual interaction of the ocean and atmosphere is termed a servomechanism, the natural time scales of which are determined by a mathematical development wherein the vertically integrated vorticity equations of the ocean and atmosphere are coupled by their interaction at the naviface. This coupling leads to a single wave equation for the ocean/atmosphere system, the solutions of which are Rossby waves modulated by $\exp[(1+i)\alpha t]$, where α depends upon the coupling parameters. Normal values of α are found to produce an e-folding increase in the vorticity of the ocean/atmosphere system in less than two months. For anomalously high values of α, the increase in vorticity can be extreme, possibly leading to the formation of a barotropic instability in the atmospheric medium. These theoretical results are illustrated using geophysical data from 1950–60 and are used to explain the events that triggered the unusual ocean/atmospheric vorticity state that existed in the North Pacific between 1956 and 1958.

1. Introduction

Dramatic and unusual events occurred in the oceanic and atmospheric fields of the mid-latitude North Pacific between 1956 and 1958 (Namias, 1959). Sea surface temperatures were anomalously high over much of the region. The westerly wind system was characterized by particularly low monthly index states during the winters of 1956 and 1957. These and other events were so marked that a special symposium was convened at the California Cooperative Fisheries Investigations Conference in 1958 to discuss the anomalous period. In a critical review of that symposium, Isaacs and Sette (1960) remarked that no theory then in existence could account for the anomalous ocean/atmosphere events that had been observed.

In the intervening years, Namias (1959, 1962, 1969) and others have begun to describe the mechanics of the large-scale, mid-latitude ocean/atmosphere coupling over the North Pacific. Namias (1959) has gathered together a large amount of evidence to descriptively explain the anomalous events observed in the eastern North Pacific during 1957–58 based upon the mutual interaction of anomalous sea surface temperature fields and the sea level pressure patterns. However, he has not been able to explain the trigger mechanism that initiated this anomaly activity.

The present paper is an extension of the work of Namias (1959) through the use of the geophysical data base developed by the North Pacific Study Group at Scripps Institution. From this data base we have established a theory describing a large-scale feedback mechanism (servomechanism) between the mid-latitude ocean and atmosphere. The theory is discussed heuristically in Section 2 and mathematically developed in Section 3. In Section 4, the theory is used with geophysical data to explain the events that *triggered* the 1956–58 ocean/atmosphere anomalies about which Namias (1959) has written.

2. Heuristic development of the servomechanism theory

Over the mid-latitude North Pacific, large temporal and spatial variations (on the order of years and thousands of kilometers, respectively) are found in the monthly and annual indices describing the general circulation of the atmosphere. However, the natural meteorological time scales are short compared to these index fluctuations. Therefore, the ocean may play an important role in the maintenance of these variations, acting to influence the atmosphere through the effect

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of the ocean circulation upon heat exchange between the two fluid media. This heat exchange can significantly affect the wind systems in the atmosphere. In turn, the circulation in the upper layers of the ocean is driven by the wind systems. The resulting mutual interaction can be thought of as a servomechanism acting to couple the two fluid media.

Throughout the year, the atmosphere drives the ocean over large geographical areas by the transfer of horizontal momentum across the air-sea interface by wind stress. In contrast, the North Pacific Ocean drives the atmosphere above it principally through the transfer of heat energy (both latent and sensible) at two geographically *localized* heat sources: the intertropical convergence zone in the equatorial Pacific Ocean and the subarctic frontal zone off the east coast of Asia. Fig. 1 illustrates the geographical distributions of heat flux (latent and sensible) across the air-sea interface in the North Pacific Ocean (Jacobs, 1950) for the winter and summer seasons. In summer, the mean heat flux from the ocean to the atmosphere (positive values) is confined principally to the region south of 30N, where much of the latent heat released at the intertropical convergence zone is picked up by the northeast trade-wind regime. On the other hand, in winter a strong heat flux source can be found at the subarctic frontal zone extending from the coast of Japan eastward to 160W. This region clearly has the largest value of heat flux in the North Pacific and, therefore, would be expected to exert a significant influence upon the westerly wind regime during the winter months.

To understand how the heat flux at the subarctic frontal zone off the east coast of Asia affects the westerly wind regime, we turn our attention to the work of Fisher (1958), who reported that the rate of change of relative vorticity in the upper-level westerly wind regime over the central North Atlantic fluctuated in response to the Laplacian of the heat flux ($\nabla^2 Q$) at the subarctic frontal zone off the east coast of North America. A similar situation should exist in the North Pacific. Taking this relationship to its logical conclusion, the corresponding increase in the intensity of the relative vorticity in the westerly wind regime (in response to $\nabla^2 Q$) should lead to an increase in the strength of the wind stress curl and, through Sverdrup's (1947) theory, the wind-driven ocean circulation.

If the oceanic region at subarctic frontal zone can be considered to be in quasi-geostrophic equilibrium (via density) with the wind-driven Sverdrup transport, then the characteristics of the zone would be intensified by any increase in transport. The intensification would result in increased magnitudes of $\nabla^2 Q$ at the subarctic frontal zone. The coupling of $\nabla^2 Q$ at the subarctic frontal zone, the relative vorticity of the upper-level westerly wind regime, and the Sverdrup transport should then operate to increase monotonically the intensity of the relative vorticity in both fluid media of the coupled ocean/atmosphere system. Such an increase can only be checked by a vorticity sink or an instability in either one or the other media.

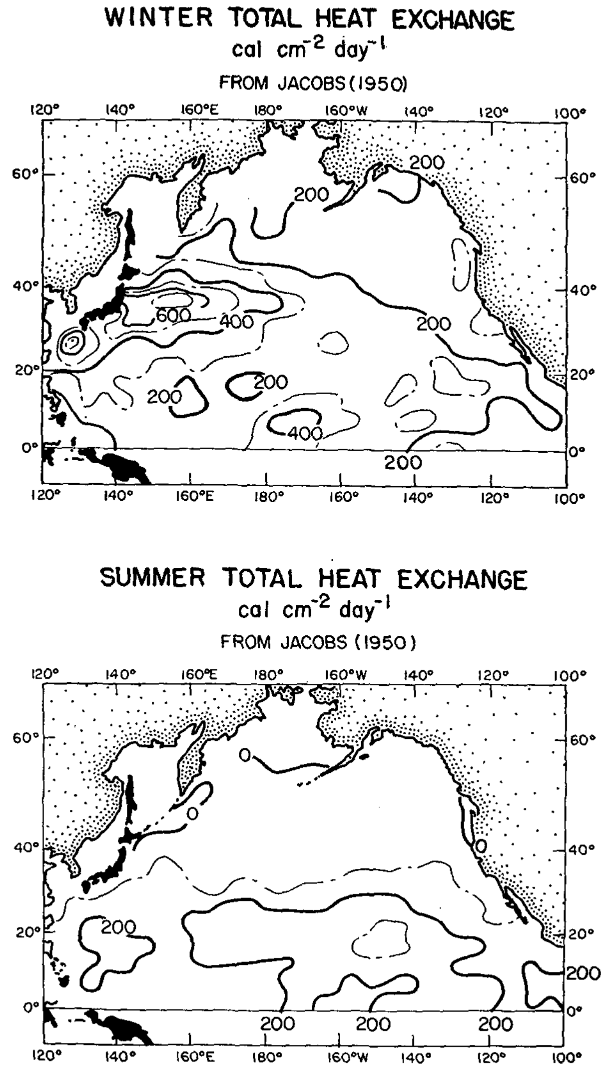

Fig. 1. Mean geographical distribution of the sensible and latent heat exchange between ocean and atmosphere in summer and winter for the North Pacific basin (Jacobs, 1950). Positive values indicate heat gain by the atmosphere.

On a climatic time scale, the ocean has a vorticity sink along the western boundary of the ocean as indicated by Munk (1950). The atmosphere has as its vorticity sink the earth, which exerts a drag upon the winds that move over it. However, in both fluid media frictional drag maintains energy and vorticity conservation—it does not limit the growth of the equilibrium state of energy and vorticity in the fluid medium. To understand this, consider the effective role of friction in the steady state model of Munk (1950). Horizontal eddy friction along the western boundary acts to provide closure for the interior Sverdrup (1947) solution, which is independent of friction. The Sverdrup interior ocean can respond to the magnitude of any reasonable wind stress curl; eddy friction simply provides for conservation of energy when the Sverdrup ocean is bounded. As such, eddy friction does not limit the response of the ocean to ever increasing strengths of wind stress curl. Similar statements can be made about the response of the atmosphere to ever increasing strengths at $\nabla^2 Q$.

Therefore, since friction will not limit the growth of energy of the coupled system envisioned here, an instability may be the result. One possible explanation of the instability mechanism in the westerly wind regime has been given by Kuo (1951). Solving the barotropic quasi-nonlinear perturbation equations of the atmosphere Kuo found that the mean westerly flow is able to increase its relative vorticity only to the extent that the meridional gradient of absolute vorticity goes to zero. When this happens, the zonal flow becomes unstable to perturbations which then amplify and thereby extract energy from the mean zonal flow. Because the absolute vorticity in the atmosphere experiences much larger changes than in the ocean, one would except the instability to occur in the former medium. The initial ridge-building observed at 700 mb over the North Pacific (see the 700-mb height chart for January 1956 in Fig. 3) may be a manifestation of this kind of instablity.

3. Mathematical development of the servomechanism theory

The servomechanism theory can explain the anomalous events observed provided the normal time scale for vorticity spin-up is some fraction of the autumn and winter time period (6 months) over which the servomechanism is operating. To establish this time scale we have developed a simple mathematical model that describes the servomechanism theory.

A first step in the development of the mathematical approach is to show that wind systems on the time scale of months can affect the ocean in a manner that can be approximated by Sverdrup's (1947) theory. Previously, Welander (1959) has calculated the total integrated transport of the world's oceans using the transport equation given by Sverdrup (1947). Upon the addition of the local time derivative of relative vorticity to Sverdrup's transport equation, Welander found that the time-scale of the oceanic response should be

$$T=(\beta L)^{-1}, \tag{1}$$

where L is the length scale involved and β the meridional derivative of the Coriolis parameter. For ocean-wide length scales, this time response is less than a day. Such a short time scale, Welander points out, is unrealistic and is due to the violation by the real ocean of one of Sverdrup's underlying assumptions that constrains the ocean to respond baroclinically to a transient wind stress curl.

The latter point has been emphasized by the work of Veronis and Stommel (1956) whose time-dependent analysis of the two-layer oceanic model clearly separated the baroclinic and barotropic geostrophic response to the time-dependent wind stress curl. The relative importance of these two geostrophic modes is dependent upon both the time and length scales involved. For time scales of months to years, the ratio of kinetic energy (E) of these two geostrophic modes can be calculated from the results of Veronis and Stommel (1956) to be

$$\frac{E_{\text{baroclinic}}}{E_{\text{barotropic}}}=\left(\frac{T\beta g' D}{f^2 L}\right)^2, \tag{2}$$

where D is the ocean depth, L the zonal length scale, g' the buoyant gravitational constant, f the Coriolis parameter, and T the time scale.

For spatial scales the size of the North Pacific Ocean, the time scale necessary to make the above ratio equal to unity is slightly over 2 years, with a predominant baroclinic response (or Sverdrup transport) taking decades. More importantly here, for spatial scales the size of the subarctic frontal zone off the east coast of Asia (1000 km), the time scale to make the above ratio unity is less than a month. On the basis of this result we might expect significant baroclinic response in the vicinity of the subarctic frontal zone during the autumn and winter seasons when the wind stress curl is much higher there than at other regions of the North Pacific. In this situation the application of Sverdrup's (1947) theory seems reasonable.

The conservation equation necessary to determine the change in oceanic absolute vorticity in response to an external source (e.g., wind stress curl) can be written

$$\left.\frac{d(\zeta+f)}{dt}\right|_w=-(\zeta+f)_w\nabla_p\cdot\mathbf{V}_w-g\frac{\partial}{\partial p}(\nabla\times\boldsymbol{\tau}), \tag{3}$$

where pressure (p) is the vertical coordinate, ζ is the isobaric relative vorticity, f the planetary vorticity, $\boldsymbol{\tau}$ the frictional stress transferring horizontal momentum across pressure surfaces, $\nabla_p\cdot\mathbf{V}_w$ the isobaric divergence, and the subscript w denotes the oceanic regime. In the ocean the pressure surfaces approximately coincide with constant depth surfaces.

The conservation equation necessary to determine the change in absolute vorticity in the atmosphere over oceanic areas of strong upward heat flux can be written (see the Appendix)

$$\left.\frac{d(\zeta+f)}{dt}\right|_a=-(\zeta+f)_a\nabla_p\cdot\mathbf{V}_a-g\frac{\partial}{\partial p}(C\nabla^2 Q), \tag{4}$$

where Q (cal cm^{-2} sec^{-1}) is the net heat flux across a pressure surface, ∇^2 the horizontal Laplacian operator, C a specified constant given in the Appendix, and the subscript a denotes the atmospheric regime.

In the vorticity equations (3) and (4), the effects of eddy frictional dissipation have been neglected to avoid unnecessary complications in the subsequent derivation. At first this may seem to be a serious omission, but it is not since, as discussed in the previous section, friction does not limit the growth of the kinetic energy or relative vorticity of the interacting system.

To couple the two fluid media mathematically, the two equations (3) and (4) are each vertically integrated from a pressure surface that most closely approximates mean sea level to another pressure surface where horizontal motion reduces to zero and in the atmospheric medium the heat flux is zero. The boundary conditions at the sea level pressure surface induce the coupling through the vertical exchange of horizontal momentum and the heat flux.

Allowing for nondivergence of volume transport in each fluid media results in the existence of the volume transport streamfunction (ψ). Expressing the vertically integrated velocities in terms of these streamfunctions results in a set of vertically integrated equations

$$\frac{\partial \nabla^2 \psi_w}{\partial t}+\beta\frac{\partial \psi_w}{\partial x}=\frac{1}{\rho_w}\nabla\times\boldsymbol{\tau}_0, \tag{5}$$

$$\frac{\partial \nabla^2 \psi_a}{\partial t}+\beta\frac{\partial \psi_a}{\partial x}=\frac{-C}{\rho_a}\nabla^2 Q_0, \tag{6}$$

where Q_0 is the heat flux from the ocean to the atmosphere and $\boldsymbol{\tau}_0$ is the horizontal wind stress applied at the sea surface. The homogeneous part of these equations is similar to that used by Welander (1959).

Before coupling these equations, we want to point out some of the assumptions that limit the credibility of the mathematical approach. First, we have made the assumption that the ocean/atmosphere system is unbounded in the horizontal direction. As such, the continental constraints upon the ocean and atmosphere are absent. Even more critical, the forcing of one fluid by the other is continuous regardless of the position of the interacting wave forms; in reality, this forcing is confined to the region of the subarctic frontal zone. We have also made the assumption that reception of heat by the atmosphere does not change the heat flux across the naviface. In light of these assumptions, the mathematical development must be considered a rather crude approximation to the actual geophysical situation.

The ocean/atmosphere coupling is completed by relating both $\boldsymbol{\tau}_0$ and Q_0 to ψ_a and ψ_w, respectively, as follows:

$$\nabla\times\boldsymbol{\tau}_0=r\nabla^2\psi_a, \tag{7a}$$

$$\nabla^2 Q_0=-\left(\frac{Q_c}{M_c}\right)\nabla^2\frac{\partial \psi_w}{\partial y}. \tag{7b}$$

In (7a) the wind stress is approximated by a linear form of the friction law with r the coefficient of friction. In (7b) the Laplacian of the heat flux is related to the Laplacian of the volume transport (M) of the ocean through the scale constants Q_c and M_c (mass transport scale of the ocean). This parameterization is possible because of the close relationship observed at the subarctic frontal zone between the field of zonal motion and the heat flux field. The latter is dominated by the distribution of the sea surface temperature in the region of the subarctic frontal zone.

Putting (7a) into (5), and (7b) into (6), the vorticity equations for each fluid media can be written

$$\frac{\partial}{\partial t}\nabla^2\psi_w+\beta\frac{\partial \psi_w}{\partial x}=\frac{r}{\rho_w}\nabla^2\psi_a, \tag{8}$$

$$\frac{\partial}{\partial t}\nabla^2\psi_a+\beta\frac{\partial \psi_a}{\partial x}=\frac{C}{\rho_a}\frac{Q_c}{M_c}\nabla^2\frac{\partial \psi_w}{\partial y}. \tag{9}$$

In the above formulations, the amount of vorticity each fluid media loses to the other is considered to be negligible compared to the amount of vorticity contained in that medium. This is tantamount to assuming a weak interaction between the fluids.

Eqs. (8) and (9) are two equations in two unknowns. Separating the variables into two equations, of one unknown each, yields a wave equation that is the *same* for both the ocean and atmosphere:

$$\frac{\partial^2}{\partial t^2}\nabla^4\psi+2\beta\frac{\partial^2}{\partial x\partial t}\nabla^2\psi-\left(\frac{CQ_c r}{\rho_a M_c \rho_w}\right)\nabla^4\frac{\partial \psi}{\partial y}+\beta^2\frac{\partial^2\psi}{\partial x^2}=0. \tag{10}$$

Assuming a wave solution of the form

$$\psi=\psi_0 e^{i(kx+ly-\sigma t)}, \tag{11}$$

the characteristic equation can be written to yield

$$\sigma=\frac{-\beta k}{(k^2+l^2)}\pm(1+i)\alpha, \tag{12}$$

where

$$\alpha\equiv\left(\frac{CQ_c r l}{2\rho_w M_c \rho_a}\right)^{\frac{1}{2}}, \tag{13}$$

which leads to the well-known Rossby (1939) solution for planetary waves, whose amplitude is modulated by $\exp(1+i)\alpha t$.

The imaginary component of σ leads to a solution of (11) whereby the amplitude of the initial vorticity state, $\psi_0(k^2+l^2)$, grows to very large amplitudes; at the same time the amplitude is modulated by the wave form $\exp(i\alpha t)$. This exponential increase in vorticity should continue until the existing absolute vorticity can no longer provide the necessary restoring forces required to make perturbations stable (Kuo, 1951). Mathematically, this occurs whenever

$$\frac{\partial}{\partial y}\left(f-\frac{\partial U}{\partial y}\right)=0, \tag{14}$$

where U is the mean flow positive to the east. The resulting barotropic instability causes the amplification of the perturbations and a subsequent extraction of energy and vorticity from the mean zonal flow.

Of primary importance is the time scale (T) of the exponential increase in the strength of the vorticity of the two fluid media. From (13)

$$T=2\pi\left(\frac{\rho_a M_c \rho_w}{CQ_c rl}\right)^{\frac{1}{2}}. \tag{15}$$

where the evaluation of T can be obtained by considering the normal values for the parameters on the right hand side of (15):

$$\begin{aligned} M_c&=3.5\times10^5\ \mathrm{cm^2\ sec^{-1}}\\ Q_c&=2.5\times10^5\ \mathrm{ergs\ cm^{-2}\ sec^{-1}}\\ C&=2.2\times10^3\ \mathrm{sec}\\ r&=1.0\times10^{-9}\ \mathrm{gm\ cm^{-3}\ sec^{-1}}\\ l&=1.0\times10^{-8}\ \mathrm{cm^{-1}} \end{aligned}$$

The time scale T can be shown to be approximately 30 days. If we consider that the baroclinic oceanic response (Veronis and Stommel, 1956) was only half that indicated in the oceanic vorticity equation (5), a more reasonable period might be closer to 60 days.

By inspection of the monthly averaged 700-mb height maps obtained from the National Weather Service, the relative vorticity in the upper-level westerly wind regime increases its strength from a summer minimum, $|\zeta|\ll f$, to a winter maximum of $|\zeta|\approx f$ by nearly an order of magnitude. In light of the present work, it seems that much of this increase can be attributed to $\nabla^2 Q$ at the subarctic frontal zone. Given an abnormally slow rate of increase of vorticity ($T>120$ days) the autumn and winter season combined might not be expected to bring on the sort of instability indicated in (14). However, from (15) one can see that a large value for l, representing abnormally large values of zonal transport and $\nabla^2 Q$ at the subarctic frontal zone, could cause the strength of the relative vorticity in the upper-level westerly wind regime to increase at such a rate ($T<30$ days) to initiate an instability sometime in the autumn season.

4. Descriptive support for the servomechanism theory

To test the feasibility of the proposed mid-latitude servomechanism theory we have applied it to the decade of 1950–60. In particular, the years 1954–56 will be emphasized, not only because they contain key events that triggered the unusual ocean/atmospheric features that dominated the latter part of the decade (1956–58), but also because they were very anomalous years and if the servomechanism theory can be demonstrated it should be possible during these years. In all cases the tests of theory will be qualitative, primarily because all the necessary data are not (and probably never will be) at hand. Hence the following discussion can only support, not substantiate, the servomechanism theory.

In Fig. 2 are three panels whose graphs characterize the fluctuations in the ocean/atmosphere system; they show how the large-scale westerly wind regime and the Sverdrup transport of the North Pacific Ocean varied over the decade 1950–60. These graphs have been computed from the work of Fofonoff (1960) and Fofonoff and Dobson (1963). The first two panels show the annual index of zonal wind stress (westerly wind regime) and wind stress curl to have fluctuated in unison. This latter parameter is proportional to the product of both the wind speed and the relative vorticity contained in the surface wind, so it is not surprising that the wind stress and the wind stress curl are in unison. Moreover, in the upper panel note that the annual values of zonal wind stress were dominated in a statistical sense by the winter values—the same can be said for the annual values of wind stress curl although it is not shown here. The annual values of Sverdrup transport of both the two main oceanic gyres, shown in lowest panel of Fig. 2, can be seen to have fluctuated in a similar fashion as the two atmospheric quantities, although the theory of Veronis and Stommel (1956) discussed in the beginning of Section 3 indicates that the actual fluctuations in oceanic transport should be about half that indicated in Fig. 2. This correspondence in the fluctuations of the bulk motions of the ocean and atmosphere is highly significant since it suggests that the two fluid media act as a closely coupled system.

Most important to this study is the characterization in Fig. 2 of the ocean/atmosphere system as being either of a low or high index state (i.e., index state here designates mean zonal kinetic energy state of the system averaged over a month or more). Because the winter months dominate the annual mean, a low index winter month corresponds to a low index year. Note that 1956 and 1957 were two relatively low index years in succession. In the atmosphere the zonal wind stress was low both years, as was the wind stress curl; in the ocean the total integrated transport was low with respect to 1955 and 1958, but was higher than low index years in the early fifties. This unusual occurrence is important to the anomalous activity that occurred in the eastern North Pacific Ocean during the 1957–58 period and is probably related to the anomalously high index state in 1955.

It is unclear what the low and high index states mean for the general circulation pattern of the ocean. However, in the atmosphere the low and high index states are characterized by two distinctive patterns

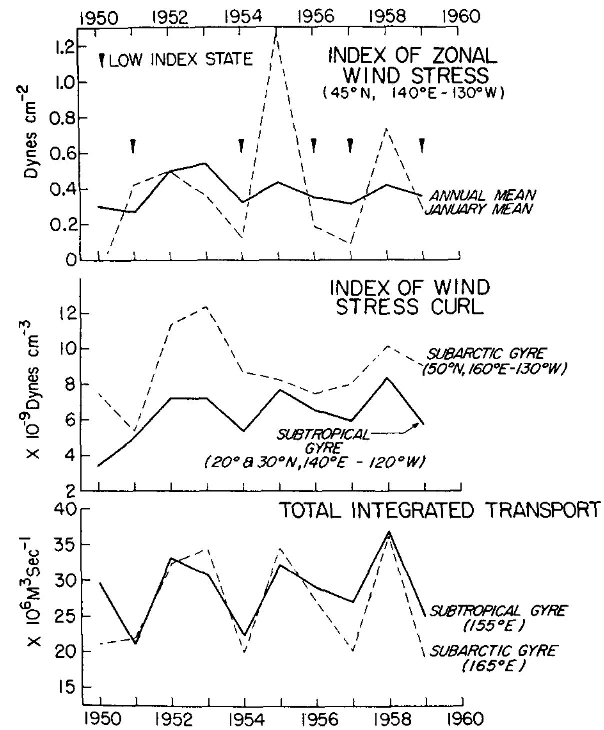

FIG. 2. Time history of the ocean/atmosphere indices of motion in the North Pacific for the decade 1950–60 (Fofonoff, 1960; Fofonoff and Dobson, 1963): annual and January index of zonal wind stress at 45N averaged between 140E and 130W; annual index of wind stress curl for respective gyres averaged from one side of ocean to the other; and annual index of total integrated transport for respective gyres measured near 160E.

of winter circulation. Two representative examples of the winter low and high index states of the atmospheric pressure field for this decade are shown in Fig. 3, obtained from the National Weather Service. In January 1955 the westerly regime at 700 mb was intensified, with the development of large values of zonal kinetic energy characterizing the high index state. At sea level the pressure distribution shows a strong low pressure area, associated with strong surface winds extending in a zonal band across the entire North Pacific Ocean. On the other hand, in January 1956 the westerly regime at 700 mb was weak, with much of the kinetic energy taken out of the zonal flow by the meridional motion of the *amplified quasi-stationary long wave:* this characterizes the low index state. The sea level pressure distribution was divided into two weak low pressure areas, associated with weak zonal surface winds.

The important question is how the atmosphere is transformed from its high index state in 1955 to its low index state in 1956. What triggered the formation of the amplified long wave associated with the low index state? The discussion of Sections 2 and 3 suggests that the formation of the amplified long wave may have been due to a barotropic instability in the upper-level westerly wind regime induced by the forcing of an anomalously high value of the Laplacian of the heat flux at the subarctic frontal zone during the winter of 1955–56. While data are not available to check the onset of this instability via Eq. (14), it is possible to verify the events that would lead to the occurrence of such an instability.

Consider the year-to-year fluctuations in the character of the subarctic frontal zone off the east coast of Asia (Fig. 4). The Kuroshio extension forms the southern boundary of this frontal zone and the upper panel in Fig. 4 presents the time history of its latitudinal position. In the middle panel the second derivative of temperature ($\partial^2 T/\partial y^2$) across the subarctic frontal zone is displayed. Comparing the latter time history with the former shows that as the southern boundary of the subarctic frontal zone was displaced northward,

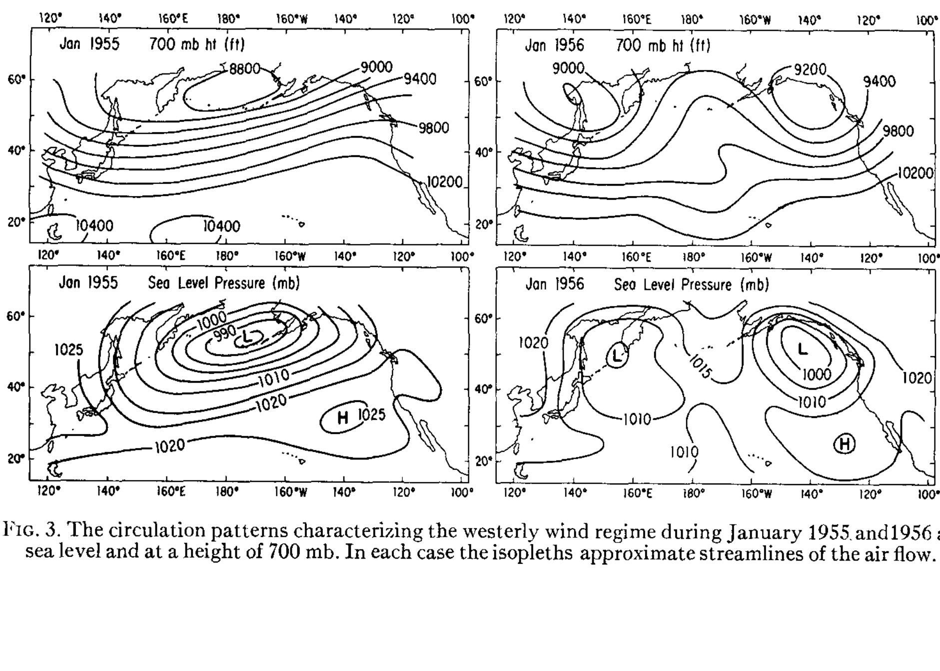

FIG. 3. The circulation patterns characterizing the westerly wind regime during January 1955 and 1956 at sea level and at a height of 700 mb. In each case the isopleths approximate streamlines of the air flow.

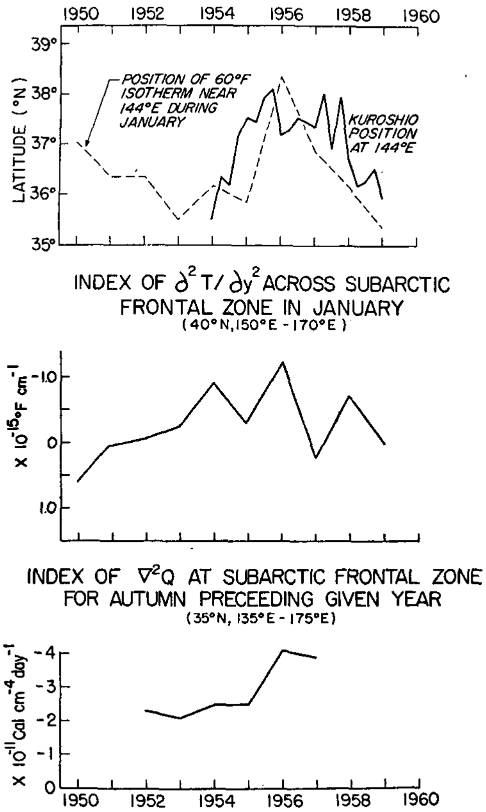

FIG. 4. Time history of subarctic frontal zone characteristics for the decade 1950–60: latitude of Kuroshio extension at 144E (Masuzawa, 1960); index of the second meridional derivative ($\partial^2T/\partial y^2 \approx T_{35}+T_{45}-2T_{40}$) of the January sea surface temperature across the subarctic frontal zone averaged between 150 and 170E; and index of the Laplacian ($\nabla^2Q \approx Q_{40}+Q_{30}-2Q_{35}$) of the autumn heat flux at the subarctic frontal zone averaged between 135 and 175E (Q values courtesy of Nate Clark, National Marine Fisheries Service (NMFS), La Jolla).

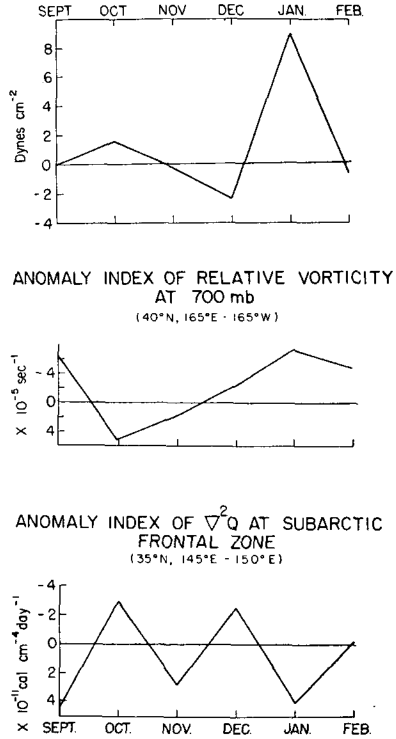

FIG. 5a. Time history of ocean/atmosphere anomaly characteristics for the autumn and winter of 1954–55: anomaly index of westerly wind stress averaged over 40 and 45N from 140E to 130W (Fofonoff, 1960); anomaly index of relative vorticity at 700 mb averaged over 40N from 165E to 165W; and anomaly of Laplacian of heat flux averaged over 35N from 145 to 150E (Q values courtesy of Nate Clark, NMFS-La Jolla). The anomaly index values are calculated by subtracting the monthly mean index values from the monthly index values.

$\partial^2T/\partial Y^2$ increased in intensity. Because the distribution of heat flux is determined principally by the sea surface temperature distribution, we expect the Laplacian of the heat flux (∇^2Q) to have varied in phase with $\partial^2T/\partial y^2$. The two lower panels of Fig. 4 show the intensity of both $\partial^2T/\partial y^2$ and ∇^2Q reaching peak values in the winter of 1956 together with the maximum northward displacement of the Kuroshio extension.

Comparing Figs. 2 and 4, one can see that the secondary peaks in ∇^2Q occurred during the low index years of 1954 and 1956, each preceded by years, of a high index state in the ocean/atmosphere system. The high index state is associated with large values of Sverdrup transport in the ocean. The evidence indicates that this larger transport causes the subarctic frontal zone to narrow in response to geostrophic adjustment. This narrowing of the frontal zone in turn increases ∇^2Q with the onset of the following autumn and winter seasons when the air-sea temperature difference becomes large. Conversely, during low index years the Sverdrup transport is low so that the subarctic frontal zone is broader, resulting in low values of ∇^2Q with the onset of the following autumn and winter seasons.

Following this reasoning, increased Sverdrup transport in one year should set the stage in the following autumn and winter seasons for an even larger increase in the Sverdrup transport as the result of an increase in the strength of the relative vorticity in the westerly wind system due to the forcing of anomalously intense values of ∇^2Q. However, observationally this does not happen because of the formation of an amplified, quasi-stationary long wave in the upper-level westerly wind regime. To understand this consider the anomalous monthly values of the westerly wind stress, relative vorticity at 700 mb, and ∇^2Q at the subarctic frontal zone for the autumn and winter seasons of 1954–55 and 1955–56 (Figs. 5a and 5b, respectively).

In Fig. 5a, the anomaly[2] index of westerly wind stress was relatively normal from September–December, associated with inconsistent values of anomalously

[2] Anomaly index refers to the monthly *mean* index values subtracted from the monthly index value.

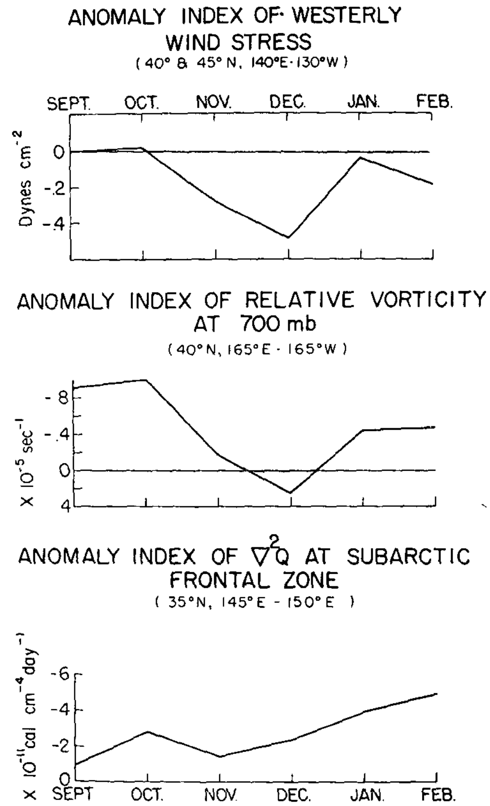

Fig. 5b. As in Fig. 5a but for 1955–56.

high and low vorticity at 700 mb. Over this time period the forcing on the atmosphere by $\nabla^2 Q$ at the subarctic frontal zone was variable. In January a very high monthly index situation is encountered in the westerly wind stress, but seemingly independent of any influence from $\nabla^2 Q$ at the subarctic frontal zone. From these data, the qualitative requirements for instability were not found, nor was the manifestation of any instability (amplified long wave) observed from an inspection of the monthly mean 700-mb height charts for the autumn and winter seasons of 1954–55.

In 1955–56, the situation was much different. Fig. 5b shows the anomaly index of westerly wind stress to be normal from September–October; however, the cyclonic vorticity at 700 mb was found to be anomalously intense. Over these two months the forcing on the atmosphere by $\nabla^2 Q$ at the subarctic frontal zone was also anomalously and continuously strong, reaching a secondary peak in October. Unlike the previous year, this situation is conducive for instability, and, indeed, during the following month of November a low index state did occur. This is reflected in the November low indices of both the westerly wind stress and the relative vorticity at 700 mb, together with the November 700-mb height chart which shows (not given here) the existence of an amplified, quasi-stationary long wave. This long wave was maintained from November–February by the continued anomalously intense values of $\nabla^2 Q$ and can be observed in the 700-mb height chart for January 1956 in Fig. 3.

There seems little doubt that the oceanographic and meteorological conditions required by the servomechanism theory of Section 3 were present to some degree in the two fluid media in 1954–56. It also seems clear that the chronology of events required by the theory was at least partially observed. The evidence reported by Namias (1959) indicates that the entire sequence of anomalous events throughout 1957–58 in the central and eastern parts of the North Pacific Ocean was initiated by the formation of the amplified, quasi-stationary long wave in the upper-level westerlies in the winter of 1955–56. The servomechanism theory presents a plausible explanation for the initial development of that feature.

5. Summary

A servomechanism theory has been formulated to explain the beginning of the anomaly activity that dominated the mid-latitude North Pacific from 1956–58. Basically, the theory entails a mutual feedback of vorticity between the ocean and atmosphere, the oceanic vorticity (and hence distribution of heat) being altered by the wind stress curl and the atmospheric vorticity being altered by the Laplacian of the heat flux at the subarctic frontal zone. These two mechanisms are coupled such that the vorticity in each fluid media experiences an e-folding increase every two months. An anomalously high increase in relative vorticity is thought at present to initiate an atmospheric barotropic instability (Kuo, 1951).

A significant amount of historical data supports this theory. During late 1955 the wind-driven Sverdrup transport was at a maximum; the strength of the second meridional derivative of sea surface temperature across the subarctic frontal zone was a maximum for the decade; and the strength of $\nabla^2 Q$, in turn, was at a decadal maximum during the autumn of 1955 and the winter of 1956. The size of the Laplacian of heat flux was such that it should have increased the intensity of the atmospheric relative vorticity in the westerly wind regime during early 1956. However, this was not observed; instead, during November 1955 the zonal wind stress in the westerly wind regime plunged to small values, associated with the low index state and diminished intensity of relative vorticity in the upper-level westerly wind regime. The evidence suggests that the increase in cyclonic vorticity during late 1955 was too much, such that the upper-level westerly wind regime could no longer maintain the proper restoring forces (through the conservation of absolute vorticity) to otherwise stable perturbations. The result was the formation of an amplified, quasi-stationary long wave in the upper-level westerly wind system.

The unstable mid-latitude servomechanism described above has important implications, for according to

Rossby (1939) it is possible for the amplified, quasi-stationary long wave to position the semi-permanent centers of action of the atmosphere over the North Pacific Ocean. In more pertinent studies of this nature, Namias (1963) has found evidence in the North Pacific which suggests that the ridges and troughs in the upper-level westerly wind regime affect the ocean/atmosphere system in the equatorial and Southern Hemisphere regions through a back interaction with the Hadley cell circulation. Therefore, this mid-latitude servomechanism may play an important role in the generation of ocean/atmosphere anomalies over the entire Pacific Basin.

In the future, we plan to make a careful study of the westerly wind regime in an attempt to more fully understand the transition from high to low index state. The servomechanism theory suggests that the transition results from an instability in the ocean/atmosphere vorticity state. In addition, other aspects of the servomechanism will be tested although this is severly hampered at the present time by the lack of necessary data in the mid-latitude zone of the North Pacific.

APPENDIX

Development of Atmospheric Vorticity Equation

The conservation of absolute vorticity in the atmosphere can be expressed by

$$\frac{d(\zeta+f)}{dt}=-(\zeta+f)\nabla_p\cdot\mathbf{V}+g\frac{\partial}{\partial p}\left(\rho_a gB\frac{\partial\zeta}{\partial p}\right), \tag{A1}$$

where pressure (p) is the vertical coordinate. Most of the parameters in (A1) have already been defined in Section 3 except for B which is the vertical exchange coefficient for the atmosphere. In (A1) the change in absolute vorticity following a parcel of air can occur in two ways: through isobaric divergence ($\nabla_p\cdot\mathbf{V}$) and by diffusion of relative vorticity $[\rho_a g(B\partial\zeta/\partial p)]$. Concerning this latter term, it is of interest to establish the diffusion of relative vorticity across the air-sea interface induced by the strong upward heat flux from the ocean to the atmosphere. To establish this an approach similar to that used by Fisher (1958) is followed.

The geostrophic relative vorticity at any atmospheric pressure surface can be written

$$\zeta_n=\frac{g}{f}\nabla^2 h_n, \tag{A2}$$

where h_n is the height of the isobaric surface p_n above sea level. Taking the partial derivative of (A2) with respect to pressure and integrating it from some pressure surface (p_n) near sea level to a reference surface ($p_0\rightarrow 0$ mb) yields

$$\zeta_n=-\frac{g}{f}\nabla^2 h_n'+\zeta_0, \tag{A3}$$

where

$$h_n'=h_0-h_n, \tag{A4}$$

h_0 being the height of the reference pressure surface. Taking the partial time derivative of (A3) yields

$$\frac{\partial\zeta_n}{\partial t}=\frac{-g}{f}\nabla^2\frac{\partial h_n'}{\partial t}, \tag{A5}$$

where $d\zeta_0/dt=0$. Eq. (A5) is now compatible with the hydrostatic relation of the atmosphere

$$\frac{\partial h_n'}{\partial t}=\frac{R}{g}\ln\left(\frac{p_n}{p_0}\right)\frac{\partial\bar{T}_n}{\partial t}, \tag{A6}$$

where $\bar{T}$ is the mean temperature of the layer and R the universal gas constant for dry air. Substituting (A6) into (A5) yields

$$\frac{\partial\zeta_n}{\partial t}=-\frac{R}{f}\ln\left(\frac{p_n}{p_0}\right)\nabla^2\frac{\partial\bar{T}_n}{\partial t}. \tag{A7}$$

From the first law of thermodynamics, $\partial\bar{T}_n/\partial t$ in (A7) can be approximated to read

$$\frac{\partial\bar{T}_n}{\partial t}\approx\frac{1}{c_p}\frac{\partial\bar{q}_n}{\partial t}, \tag{A8}$$

where $\bar{q}_n$ is the mean heat accession (cal gm^{-1} sec^{-1}) in the air column and c_p the specific heat at constant pressure. Putting (A8) into (A7) yields

$$\frac{\partial\zeta_n}{\partial t}=\frac{-R}{fc_p}\ln\left(\frac{p_n}{p_0}\right)\nabla^2\frac{\partial\bar{q}_n}{\partial t}. \tag{A9}$$

Integrating (A9) in the time domain yields

$$\zeta_n=\frac{-R}{fc_p}\ln\left(\frac{p_n}{p_0}\right)\nabla^2\bar{q}_n, \tag{A10}$$

where at some initial time both ζ_n and $\nabla^2 q_n$ are considered zero.

Independent of the above derivation, the heat flux Q_n across the pressure surface p_n can be written

$$Q_n=\rho_a gB\frac{\partial\bar{q}_n}{\partial p}\ [\text{cal cm}^{-2}\ \text{sec}^{-1}]. \tag{A11}$$

Forming the flux of vorticity from (A10) and utilizing (A11) yields the diffusion of relative vorticity due to the Laplacian of the diffusive heat flux:

$$\rho_a g B \frac{\partial \zeta_n}{\partial P} = \frac{-R}{f c_p} \ln\left(\frac{p_n}{p_0}\right) \nabla^2 \left(\rho_a g B \frac{\partial \bar{q}_n}{\partial p}\right) = -C \nabla^2 Q_n, \qquad \text{(A12)}$$

where

$$C \equiv \frac{R}{f c_p} \ln\left(\frac{p_n}{p_0}\right). \qquad \text{(A13)}$$

Substitution of (A12) into (A1) gives Eq. (4) in Section 3.

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