The Spatial and Temporal Variability of the Polar Front in the Sea of Japan (Polar Front, Eddies, Thermal Gradient).

Taebo Shim

Louisiana State University and Agricultural & Mechanical College

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THE SPATIAL AND TEMPORAL VARIABILITY OF THE POLAR FRONT IN
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THE SPATIAL AND TEMPORAL VARIABILITY OF THE POLAR FRONT IN THE SEA OF JAPAN

A Dissertation

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical college in partial fulfillment of the requirements for the degree of Doctor of Philosophy in

Marine Sciences

by

Taebo Shim
B.S., Seoul National University, Korea, 1974
M.S., Seoul National University, Korea, 1980
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ABSTRACT

The spatial and temporal variability of the Polar Front in the Sea of Japan is investigated by analyzing both the conventional hydrographic data from the files of NODC and sea surface temperature (SST) data and sea height data derived from satellites.

The root mean square variability of the annual mean frontal position, which extends from 38.5 °N near the Korean Peninsula to 41.5 °N near the Tsugaru Strait, is about 100 to 120 km in the western and central region and is 300 km in the eastern region. The Polar Frontal zone appears to be unstable in summer, suggested by the large amplitude of the frontal wave motion. Small scale eddies (50-100 km) are prevalent in the frontal zone. Entrainment associated with these eddies appears to be important in transferring momentum, heat and salt to the cold water region north of the Polar Front. Both hydrographic and satellite data show the dominance of the seasonal time scale in frontal dynamics. A monthly time scale is found to be of secondary importance. Frontal dynamics on the eastern and western extremities of the Polar Front are suggested to be different.
CHAPTER 1. INTRODUCTION

Oceanic fronts in the upper ocean are ubiquitous features. Satellite observations, in particular, have demonstrated their widespread occurrence (Legeckis, 1978). Many are found to be persistent, even semi-permanent, retaining coherence for periods of several months and longer.

'Front' is a meteorological term that has been widely adopted by oceanographers, but has no precise definition in oceanography. Cromwell and Reid (1956) define a front as a band along the sea surface across which density changes abruptly. This definition does not specify the cause of the density change, which may be due to a change in either temperature or salinity, or both. (In fact, there can exist counterbalancing salinity and temperature fronts such that there is no density front).

It is difficult to formulate a uniformly acceptable numerical definition of a front. For example, Uda (1959) specifies a temperature front in terms of a gradient of about 0.25 °C/10 km to 2.5 °C/10 km, which illustrates the lack of precision in defining a front. Knauss (1957) finds extreme temperature changes of 3.0 °C/100 m at a frontal boundary in the eastern equatorial Pacific Ocean. Voorhis and Hersey (1964) suggest that the temperature gradient across a front should be 1 °C or more over 10 km distance. Despite the ambiguous numerical definition of the temperature front, it seems generally to be accepted that a temperature gradient of 1 °C/10 km is a good criterion for a thermal front. Similar difficulties exist with other parameters. Thus, while there is no agreed upon quantitative definition, the qualitative features of a front, i.e., a band across which some property
distribution exhibits a locally increased gradient, appear to be generally understood and accepted.

In many respects, ocean fronts are to be contrasted with atmospheric fronts. Nearly all of the density gradient associated with an atmospheric front is due to the temperature gradient, whereas the density gradient in the ocean is often due to both temperature and salinity. The salinity field in the vicinity of a front may interact with the temperature gradient in such a way that the density difference between adjacent water masses can be enhanced or can disappear entirely. Thus, there is not always an associated baroclinic jet with ocean thermal fronts, as there is with atmospheric fronts (Roden, 1977). The relative time and space scales of fronts in the ocean and atmosphere also differ remarkably, i.e., there exist intrinsic differences in the respective Rossby numbers and radii of deformation (Pedlosky, 1982).

An increased interest in fronts is dictated for many reasons. Fronts are important in ocean dynamics since they are regions where vertical motions of the water and exchange of momentum and other properties are locally strong. Large scale fronts influence air-sea interaction and, thus, weather and climate. Since frontal zones usually involve one- or two-sided surface convergence, they are very effective in collecting and in concentrating floating detritus, particulate matter, and pollutants (Sick et al., 1978.)

The position of local fronts may be important to the design and positioning of sewer and power plant effluent outfalls, the positioning oil rigs, and the location of ocean dumping sites. Formation of sea fog at coastal fronts can be a hazard to navigation
and small craft piloting. The choice of fishing ground for maximum
catch often involves location of oceanic fronts, since these have been
found to be areas of high biological productivity. Underwater
acousticians know very well that fronts serve as acoustic lenses,
leading to large sound wave propagation anomalies, both positive and
negative. The knowledge of the location and intensity of fronts is
crucial to the people who operate underwater acoustic devices such as
SONAR and fish detectors. Since marine mammals and acoustically active
creatures congregate at frontal zones, these zones can be expected to
be anomalously intense in ambient noise. Therefore, the temporal and
spatial variability of fronts as well as their dynamics are important
to both the business and defense communities.

Several dynamic processes may be important, under suitable
conditions, in affecting the maintenance of strong frontal gradients
in the presence of active horizontal diffusion. These processes are 1)
Ekman transport within internal boundary layers across geostrophically
balanced fronts at shelf scales (Mooers et al., 1978), 2) cross
frontal interleaving and mixing across fronts where the horizontal
density gradient may approach zero (Horne et al., 1977), 3) cabling,
which has been suggested as a mechanism for surface convergence and
the removal of the mixed water from the frontal interface in density
balanced fronts (Garrett and Horne, 1978), and 4) turbulent
entrainment and downwelling induced by interfacial shear in small

Frontogenesis has been studied in different environments. The
scale of the resulting fronts, in general, is related to the scales of
the regions supporting them, e.g., estuaries, river mouths,
continental shelves, or open oceans.

There are a variety of proposed dynamical models for the buoyant plumes and associated fronts which are generated by the discharge of fresh or brackish water from rivers and estuaries into coastal sea water and the discharge of heated water from power plants into lakes, rivers, or the ocean. Among naturally occurring plumes, estuarine effluents have been most studied (e.g., Ryther et al., 1967; Wright and Coleman, 1971; Garvine, 1977).

At shelf scales, a variety of buoyancy flux and momentum flux mechanisms are important to frontogenesis. Garwood et al. (1981) find that a sudden air temperature drop over relatively shallow water results in cooling producing distinct temperature fronts and mild salinity fronts. They also find that increased wind speeds, under low humidity conditions, maximize this effect. The rapid response of inshore regions to such forcing can be attributed to the shallow depth and consequent low thermal inertia. According to Garwood et al. (1981), the parameter controlling occurrence of a front is the ratio d/h where d is the depth of the shoal region and h is the surface mixed layer depth in the adjacent deeper water. If d is greater than or equal to h, then there will be no front formed due to differential vertical mixing and cooling. However, if h is greater than d, a front will occur where the water depth equals h. Simpson and Pingree (1978) and Simpson et al. (1978) similarly show that, during summer months, the location of fronts in the shelf seas around the United Kingdom are essentially determined by the parameter d/u^3, where d is the water depth and u the amplitude of the tidal stream. They also present indirect evidence for vertical motions suggesting that both upwelling
and downwelling occur in the frontal zone.

Flagg and Beardsley (1978) propose that the topographic slope controls the position of shelf break fronts. They show that the New England shelf/slope water front would be quite unstable if the front were located over a nearly flat bottom. Since an increase in bottom slope generally tends to reduce the growth rate of baroclinic instabilities, the front presumably migrates offshore toward the shelf break until an equilibrium is reached between the instability mechanism, which excites frontal undulations, and local dissipation.

Frontogenesis in the open ocean has been studied by Wodan (1975, 1977). He states that frontogenesis depends strongly upon differential Ekman transport. On time scales much longer than a week or two, frontogenesis is affected not only by the wind stress field but also by the geographically non-uniform distribution of radiant and turbulent heat flux. DeSzoeyeke (1980) and Cushman-Roisin (1981) further investigate the formation of surface fronts by non-uniform wind fields, numerically and analytically, respectively.

DeRuijten (1983) investigates frontogenesis in equatorial regions, and finds that frontogenesis can occur under proper gradients in air-sea temperature differences. Nof (1983) demonstrates that a gradient in cooling, set-up by the air-sea temperature gradient in the vicinity of the Gulf Stream, is capable of producing cross stream velocities that displace the Gulf Stream System by as much as 100 km over an entire season. Wind forcing is not considered in his study. Adamec and Garwood (1985) study the effects of a gradient in the wind direction and surface buoyancy flux together. They find results consistent with observation and show that advection plays a
significant role in the horizontal displacement of the surface isopycnals. Thus, both air/sea buoyancy flux and Ekman advection appear to be important frontogenetic processes on oceanic scales.

Fronts often develop wavelike perturbations which cause large amplitude transverse displacements of the frontal zone, referred to as meanders. According to Voorhis (1969), the Sargasso Sea fronts exhibit frontal meanders with a typical wavelength of 50 km and amplitude of 22 km. Meanders of the Gulf Stream front are large, about 300 km in wavelength and 100 km in amplitude (Hansen, 1970). Roden (1981) finds that the front near 30°N in the Pacific Ocean meanders with a wavelength of 180 km and a wave amplitude of 55 km. The frontal waves generally propagate along the front in the direction of the current (Garvine, 1983). Phase speeds of about 5 cm/sec are well documented for the Gulf Stream (Halliwell and Mooers, 1979). Garvine (1983) suggests that meanders can be explained by conservation of potential vorticity along the material path. Various vorticity models dealing with sinuous meandering behavior have been put forward. Warren (1963) and Robinson and Niiler (1967) attribute the meander structure to a steering of the Gulf Stream by bottom topography, while others have sought the origin of the waves in barotropic or baroclinic instability.

The most important front in the Eastern Sea of Korea (hereafter Sea of Japan) is the Polar Front, but its characteristics are poorly known. Uda (1938) initiated the study of this front (Siome or rip current in his terminology). He depicted the wavy form of the front, and reported that the front was associated with vorticity which is usually proportional to the degree of discontinuity of the front.
According to Uda (1938), the frontal meander wavelength and amplitude at 50 m depth were about 190 km and 75-110 km, respectively, in June, 1932, and 200-260 km and 90-150 km, respectively, in October, 1932. 

Istoshin (1960) shows that thermal fronts in the Sea of Japan can be discerned all year round at 25 m depth, even in warm summer months. He also reports that the mean frontal position lies along 40 north latitude; both ends of the front, one near the Korean coast and one near the coast of Japan, show large annual variations in position, but the mid-ocean portions are relatively stable. Huh (1976) finds thermal fronts near the southeastern coast of Korea using the Defense Meteorological Satellite data. Legeckis (1978), using VHRR (Very High Resolution Radiometer) satellite imagery, shows that the Polar Front in the Sea of Japan can be detected in the sea surface temperature field.

The importance of this front is recognized by the local business and defence communities. However, present ignorance about the Polar Front will, of necessity, result in future studies. The goal of the present study is, therefore, to describe and explore the temporal and spatial variability, and structure of the Polar Front in the Sea of Japan. Such knowledge is indispensable to the efficient planning of future investigations.

The dissertation consists of five logical sections. Chapter 2 presents a description of the general oceanography of the region and the structure of the Polar Front using hydrographic data in the files of the NODC (National Oceanographic Data Center). Chapter 3 discusses results of a finer scale study of the temporal and spatial variability of the Polar Front using sea surface temperature data derived from
satellites. Chapter 4 explores the possibility of studying the
dynamics of the Polar Front using satellite altimetry. Chapter 5
summarizes the results of this study.
CHAPTER 2. The Sea of Japan and the Polar Front.

2.1 General oceanography of the Sea of Japan

In this chapter I briefly review the general oceanography of the Sea of Japan as determined from classical hydrographic surveys. Particular emphasis is placed on the structure and variability of the Polar Front. The discussion of this feature of the Sea of Japan is based largely on the National Oceanographic Data Center (NODC) data set.

General features.

The Eastern Sea of Korea (Sea of Japan) is a marginal sea of the Northwestern Pacific Ocean with an average depth of 1350 m and a maximum depth of 4049 m (Hidaka, 1966). Figure 2-1 shows the bottom topography and geography of the Sea of Japan. The middle of the Sea of Japan is traversed by the Korea Plateau and Yamato Rise, which also divides the sea into the Japan Basin on the northern part and Korea Basin and Yamato Basin to the south. A narrow continental shelf fronts along the Korean Peninsula and the USSR coast, while on the Honshu side many banks, basins and troughs develop to form ridges such as the Oki Ridge and Sato Ridge. The bottom topography in the northern half of the Sea of Japan is deeper (about 4000 m or deeper at the deepest portion) and simpler than the southern half.
Figure 2-1. The bottom topography and geography of the Sea of Japan.
Atmospheric influences.

The prevailing wind system in the Sea of Japan is the Asian Monsoon, which is northwesterly in the cold season and southeasterly in the warm season. Mean wind speeds are 6 m/s and 4 m/s in winter and in summer, respectively. The winds affect momentum transfer by shearing stresses and heat transfer by evaporation and sensible heat exchange. Roden (1977) calculated mean Ekman transport, using mean geostrophic winds, in the North Pacific Ocean including the Sea of Japan, and showed that fronts can be formed at the Ekman transport convergence zones. These Ekman transport diagrams reflect the monsoon in the Sea of Japan. Shim and Kim (1981) showed that the annual mean heat fluxes in the Sea of Japan are negative, i.e., a loss of heat to the atmosphere, because of the severe evaporation and sensible heat loss caused by the dry and cold winter monsoon.

Increases of rainfall due to the summer monsoon freshen the sea waters. The low salinity of the waters flowing into the Sea of Japan through the Korea Strait during summer is due, in part, to the Asian monsoon, which influences not only the rainfall over the Sea of Japan but also fresh water discharges into the East China Sea and Yellow Sea. Typhoons, coming usually between June and October, are another source for severe winds and rain in the area. However, these are neither of long duration nor consistent, occurring for a few days to a week two or three times a year.

Water levels.

The Sea of Japan is characterized by tides of an extremely small range. While the tidal range off the Pacific coast of Japan is 1-2 m,
it is only 0.2 m in the Sea of Japan (Lisitzin, 1967). The highest amplitudes (30-40 cm) are noted in the southern part of the Sea of Japan. Tidal amplitudes decrease gradually towards the north and are 10-20 cm near the Soya Strait. Two amphidromic points for the semidiurnal component, $M_2$, exist north of the Soya Strait and near the Korea Strait. Since tidal waves propagate at right angles to the cotidal lines, they cyclonically sweep the Sea of Japan (Hidaka, 1966).

The maximum monthly mean sea level generally occurs late in the summer (August), and the minimum occurs in winter (February-April). In summer, the sea level is at its maximum and is always higher along the eastern shore of the Korea Strait than along the Korean shore. Lisitzin (1967) claims that the monthly mean seasonal variation of the sea level is reflected in that of the baroclinic transport into the Sea of Japan through the Korea Strait. Within the Sea of Japan, seasonal sea level variability along the Japan coast is predominantly steric, but this does not appear to be the case along the Korean coast (Pattullo et al, 1955)

Mass exchange with adjacent seas.

There are four straits connecting the Sea of Japan with the surrounding ocean (Figure 2-1). They are the Korea Strait (Tsushima Strait), Tsugaru Strait, Soya Strait (La Perouse Strait), and Tatar Strait (Mamiya Strait). The Korea Strait is the main entrance to the Sea of Japan. The Strait is 280 km long, 200 km wide, and 50 - 150 m deep. It is composed of two channels, the eastern channel and western channel, which are separated by the Tsushima Islands. The western
channel narrows to the northwest with a sill depth of 200 m; the eastern channel widens to the northeast and has a nearly constant depth of about 100 m. The influx of water into the Sea of Japan is mainly through the western channel. According to Yi (1966), as much as 73% of the total volume transport through the Korea Strait flows into the Sea of Japan through the western channel. The Tsushima current, a mixture of Kuroshio, East China Sea and Yellow Sea waters, passes into the Sea of Japan via this strait and exits through the Tsugaru and Soya Straits (Sverdrup et al., 1942). Volume transport through the Korea Strait, which is the major source of advected heat and salt for the Sea of Japan (Moriyasu, 1972; Yoon, 1982), has been indirectly monitored for many years (e.g. Hidaka and Suzuki, 1950; Miyazaki, 1952; Hada, 1962; Yi, 1966; Toba et al., 1982; Shim et al., 1984). Moriyasu (1972) notes large differences in estimates of volume transport. Toba et al. (1982) suggest the volume transport to be $2.0 \pm 1.6$ Sv ($1$ Sv $= 10^6$ m$^3$/sec). Shim et al. (1984) question the validity of estimating volume transport in the Korea Strait by geostrophic calculation because this method can not resolve the strong barotropic component of the flow. Nevertheless, seasonal variations of geostrophic transport estimates, maximum in summer/fall and minimum in winter, have been detected by many scientists. Volume transport estimates differ by as much as a factor of three on a seasonal time scale. In the eastern channel, the vertical distributions of temperature and salinity are nearly homogeneous in winter and strongly stratified in summer. The lateral gradients of the temperature and salinity in the western channel are large in winter, while, in summer vertical gradients are large and lateral gradients
are small (Conlon, 1981). Water flowing through the western channel is much diluted by mixing with Yellow Sea and East China Sea waters in summer.

The Tsugaru Strait lies between Honshu and Hokkaido and is approximately 20 km wide at both ends of the strait, 50 km wide in the central part of the strait and has a sill of 125 m depth at the western end. According to Moriyasu (1972), about 75% of the total volume transport through the Korea Strait, on the average, flows out to the North Pacific through the Tsugaru Strait. The volume transport through this strait is estimated to be 1 to 4 Sv by Hada (1962) and 1.5 to 1.8 Sv by Conlon (1981). According to Conlon (1981), a lower layer return flow which is typical for most two layered straits is absent because the Tsushima Current influx into the Tsugaru Strait is so large.

Part of the Tsushima Current flows northward along the western coast of Hokkaido and into the Sea of Okhotsk through the Soya Strait. The Soya strait is about 50 km wide and less than 60 m deep. The volume transport through the strait is not large (25% or less of the volume transport through the Korea Strait (Moriyasu, 1972)).

Among the four straits, the Tatar Strait is the least important due to its shallowness and negligible exchange of water mass (Hidaka, 1966).

Circulation.

There are two current systems in the Sea of Japan (Figure 2-2): the Liman and Maritime Province Current, and the Tsushima/Tsugaru Current System (TTCS). The Liman and Maritime Province Currents (MPC)
Figure 2-2. Two current systems in the Sea of Japan. The North Korean Cold Current (NKCC) and the Liman Current compose the Liman and Maritime Province Current; the TTCS is composed of the East Korea Warm Current (EKWC) and the Tsushima Warm Current (TWC).
form a cold current and cyclonic circulation pattern. The Maritime Province Current is slow, 0.15 m/s or less. The Liman Cold Current flows at about 0.15 m/s. The extension of the Liman Cold Current, called the North Korean Cold Current (NKCC), attains speeds as fast as 0.25 m/s (ROK Hydrographic Office, 1982). These cold currents are known to be approximately 100 km wide and 50 m thick (Hidaka, 1966).

The flow associated with the warm water moving into the Sea of Japan through the Korea Strait is called the TTCS. There is no consensus as to the pattern of circulation within the TTCS. Suda and Hidaka (1932) and Uda (1934) propose that the TTCS is composed of three branches. The first branch (the near shore branch along Honshu) is the extension of the current which flows through the east channel of the Korea Strait. Often, this branch is called the Tsushima Current or Tsushima Warm Current. The second branch is an offshore branch and flows between the first and the third branch. The East Korea Warm Current (EKWC) is the third branch. Both the second and the third branches flow into the Sea of Japan through the western channel of the Korea Strait. Kawabe (1982) showed that three branches could be formed when the main thermocline within the western channel of the Korea Strait intersects the bottom slope. He proposes that the second branch (offshore branch) is attributable to the increased inflow through the western channel from June to August. However, according to other scientists (e.g., Tanioka, 1968; Moriyasu, 1972), the TTCS is composed of two branches, only. These scientists deny the existence of the second of the three branches, and consider it to be merely a meandering of the path of the EKWC. Yoon (1982) numerically predicts two branches of the TTCS. He explains the first branch as largely
topographically controlled and the second branch (EKWC) as a western boundary current.

Water masses.

Nearly 90% of the entire water body of the Sea of Japan is occupied by homogeneous water of low temperature (0-1 °C) and low salinity (34.0-34.1 ‰). This water mass is called the Proper Water of the Sea of Japan or, equivalently, the Deep Water (Yasui et al., 1967). Above this, cold Surface Water (0-25 m) and Middle Water (25-200 m) are found. Surface Water occurs in summer and is separated by a seasonal thermocline from Middle Water below. Both Surface Water and Middle Water are divided horizontally into a cold sector in the NW part and a warm sector in the SE part of the Sea of Japan by the Polar Front. The warm sector Surface Water is a mixture of high temperature and low salinity waters originating from the East China Sea and coastal water near Japan. Cold sector water is mainly due to melting of sea ice and cooling by cold winter winds. Surface temperature varies much from season to season, and from place to place. For example, it is 25 °C near the Korea Strait and 24 °C near the Tsugaru Strait in summer months, and 15 °C near the Korea Strait and 7 °C near the Tsugaru Strait in winter months (ROK Hydrographic Office, 1982). The surface salinity also varies (35 ‰ or higher in April/May, 32.5 ‰ in August/September near the Korea Strait).

Middle Water is characterized by high temperature and variable salinity. In the warm sector, water of high temperature and high salinity originates from the intermediate layer of the Kuroshio and
enters into the Sea of Japan from winter to early summer. The core of Middle Water, with a salinity which varies between 34.5 °/oo, and 34.8 °/oo is found near 50 m depth. Both sectors show a strong vertical temperature gradient through the Middle Water (e.g., 17 °C at 25 m and 2 °C at 200 m). Salinity in the warm sector has a maximum between 34.5 °/oo and 34.8 °/oo in summer, and reduces to 34.1 °/oo near 200 m depth (Figure 2-3). However, salinity in the cold sector is almost homogeneous at 34.05 to 34.1 °/oo (Hidaka, 1966). Figure 2-4 shows the horizontal salinity distribution at 50 m depth in July/August, 1973 based on hydrographic data from the National Oceanographic Data Center. The 34.1 °/oo isohaline is usually located north of the salinity frontal zone.

Dissolved oxygen contents, according to Fukuoka (1965), are 6 to 9 ml/l at the surface and vertically homogeneous near 6 ml/l in the lower layer down to 1500 m depth (observational limit in his data). These are to be compared with 7 to 10 ml/l at the surface layer, 3.5 ml/l near 400 m depth, and about 1 ml/l at 1200 m depth in the Sea of Okhotsk. The high dissolved oxygen content at depth in the Sea of Japan implies active formation of Deep Water at the surface during winter.

2.2 The Polar Front.

The complexity and variability of the upper few hundred meters of the warm water sector in the Sea of Japan is tightly linked with the variability and dynamics of the Polar Front. In particular, the
Figure 2-3. Vertical salinity profile (July, 1975) along the line AB shown in Figure 2-12-a.
Figure 2-4. Horizontal salinity distribution at 50 m (July/August, 1973) from NODC.
behavior of the EKWC, after leaving the coast, is extremely important in determining the characteristics of the Polar Front.

Systematic hydrographic surveys of the Polar Front region were started in the late 1940's even though there were a lot of hydrographic surveys by Japan before World War II. However, major features of the Polar Front such as eddies in the frontal zone and meandering of the Tsushima Current or of the EKWC have not been described by conventional hydrographic survey techniques (eddies in the Polar Frontal zone will be discussed in a later chapter.) This is due to the predetermined coarse spatial resolution of the seasonal hydrographic measurements. In spite of the poor resolution of the hydrographic data, they still provide a basis for understanding the large scale features of the circulation of the Sea of Japan.

Temperature or salinity alone can not determine circulation patterns. However, it is reasonable to use temperature and salinity as tracers because, for example the EKWC has high temperature and high salinity characteristics and loses heat and salt by mixing with ambient waters as it flows downstream. Furthermore, the Polar Front is a boundary between cold water and warm water in the Sea of Japan, and is easily detected in the temperature field. Kolpack (1982) demonstrates that salinity can also be used as a tracer for the circulation in the Sea of Japan.

After carefully inspecting almost five decades of hydrographic data (1930 -1978), 7 years, 1971 through 1977, were selected for more detailed analysis of the Polar Front region. The data used were from the files of the National Oceanographic Data Center (NODC) and received on magnetic tapes for computer analysis. The most complete
coverage for the selected period was for July/August (summer time) and February/March (winter). Even though there were scattered observations in May and November, they have been neglected because of their scanty and irregular coverage.

I choose three layers (surface, 30 m, and 50 m depth) on which to describe distributions of temperature, salinity, and sigma-t. The geostrophic velocity at the surface is calculated based on a reference level of 1000 db. Because of the volume of data, I discuss only a few typical distribution patterns.

2.3 Results

At the sea surface, the temperature distribution always shows a thermal front in February, but shows no thermal front in August except during 1971 (Figure 2-5). The thermal frontal zone (Polar Front) in February, which usually can be discerned by isotherms between 2 °C to 7 °C, is found near 38° to 41° N latitude, varying interannually in shape and position. A front associated with the Tsushima Warm Current (TWC) is sometimes found south of the Polar front. While the temperature difference between the northern and southern sides of the Polar Frontal zone is about 5 °C, that of the TWC front is usually not greater than 2 °C. The frontal zone is broader (about 100 km) than that associated with TWC front (about 20 to 30 km). The temperature gradient observed within the front is 1.0 °C/10 km, or more. This is probably an underestimate because of the coarse grid on which temperature observations are made. Typical temperatures in February are 11 °C near the Korea Strait, and 1 °C north of the frontal zone in
the northern cold water sector.

No surface salinity fronts are found in February data, during which time the horizontal surface salinity is almost homogeneous (34.0 to 34.1 °/oo) except near the Korea Strait where salinity is about 34.2 to 34.3 °/oo. In August, however, salinity fronts (Figure 2-6) define the meandering of the Tsushima Warm Current (TWC) (However, the meander wavelength of the TWC is probably undersampled). Low salinity TWC waters (33.0 to 34.0 °/oo) were observed in both 1971 and 1972. The low salinities probably originate in waters from the East China Sea and Yellow Sea, and are further lowered by fresh water discharge from land.

The patterns of surface density fronts for all 7 years mimic the thermal fronts in February because surface salinity in February is so homogeneous. On the other hand, in summer when neither the Polar Front nor the TWC thermal fronts form at the sea surface (Figure 2-7-a), density fronts in the TWC reflect the surface salinity front (compare Figure 2-6 with Figure 2-7-b). The sigma-t values along the TWC front vary from 26.75 to 27.25 in February and 22.25 to 22.5 in August.

At 30 m depth, the Polar Front can always be detected in the thermal structure near 40 °N. There is, though, large interannual variability; the mean frontal position appears to migrate northward in summer and southward in winter. The temperature distribution at 30 m depth in winter is very similar to that at the surface. The observed temperature gradients sometimes increase to 2 °C/10 km. The 5 °C isotherm is centered within the frontal zone in winter, while the 12 °C isotherm is centered within the frontal zone in summer. The shape of the thermal frontal zone at 30 m depth in summer is generally more
Figure 2-5. Sea surface temperature in August, 1971.

Figure 2-6. Sea surface salinity distribution in August, 1972.
Figure 2-7. SST (a) and sea surface sigma-t (b) distribution in August, 1972.
complex than in winter; one meander of 50 km amplitude and approximately 700 km wavelength is observed in winter, and one and half to two cycles of roughly 100 km amplitude with about 400 km wavelength in summer.

The salinity distribution at 30 m depth shows no resemblance to that at the sea surface. No salinity fronts can be seen in winter but a weak salinity front occurs near 40 °N in summer. The 34.0 °/oo isohaline always bends to the north near the mouth of the Tsugaru Strait, as part of the Tsushima Current flows out through the Soya Strait (Figure 2-8). The density pattern at 30 m depth reflects the thermal pattern in both seasons, i.e., long wavelength meanders with small amplitude in winter and relatively short wavelength meanders with large amplitude in summer. The density front is typically characterized by the 26.75 and 25.75 sigma-t isopleths in winter and in summer, respectively (Figure 2-9).

At 50 m depth, the horizontal distributions of temperature, salinity, and sigma-t are similar in pattern to those at 30 m depth in winter in the entire 7 year set. However, about a 5 °C difference in temperature occurs in summer between the two depths, the 50 m temperature being colder than that at 30 m (Figure 2-10). There are very few changes in the density patterns between 30 m and 50 m depth. Again, this is because density is mostly influenced by temperature. As can be seen on Figure 2-11, there are two thermal fronts; one near 40 °N, which is the Polar Front, and another near the Honshu coast, which is the coastal front due to the Tsushima Warm Current. The latter front shows meandering with about 150 km of amplitude and with 300 to 350 km wavelength, although these are probably undersampled. Ichiye
Figure 2-8. Salinity distribution at 30 m depth

in August, 1971.
Figure 2-9. Sigma-t distribution at 30 m depth in February (a) and August (b), 1972.
Figure 2-10. Horizontal temperature distribution at 30 m (a) and 50 m (b) in August, 1971, and 30 m (c) and 50 m (d) depth in August, 1975.
Figure 2-10. continued.
Figure 2-11. Horizontal temperature distribution at 50 m in August, 1972.
(1983) reports that the meandering of the Tsushima Current to be about 200 km amplitude with about 400 km wavelength and phase speed predicted to be 1 km per day.

Figure 2-12 shows the vertical profiles of temperature, salinity, and sigma-t in February (winter) and July (summer), 1975, along the cross section A-B shown in Figure 2-12-a. Because of mixing by the wind in winter, in the upper 100 - 200 m layer of the warm water sector of the Sea of Japan, a very strong thermocline is formed at the base of the surface mixed layer. The cold water sector is vertically well-mixed from surface to bottom, but very large lateral temperature gradients define the Polar Frontal zone, where temperature varies from 8 °C near the southern edge of the frontal zone to 3 °C at its northern edge. The mean slope of the frontal surface based on the 5 °C isotherm is very steep, i.e., 65 m/20 km to 65 m/30 km (0.0033 to 0.0022). Frontal zone width is about 100 km in winter. In summer, a strong seasonal thermocline forms near 25 m depth such that no surface thermal front occurs. Salinity fronts are weak in the vertical sections in both seasons.

The surface geostrophic velocity is calculated assuming 1000 db as a reference level. Figure 2-13-a and Figure 2-13-b show the horizontal distribution of $(\Delta \Phi / (2 \Omega)) \times 10^4 \text{ m}^2/\text{s}$ at the sea surface where $\Delta \Phi$ is the geopotential anomaly and $\Omega$ is the rate of rotation of the earth. In order to help to estimate the velocity differences between two stations from the Figures 2-13-a and 2-13-b, nomograms are presented in Figures 2-13-c and Figure 2-13-d. The relative velocity can be obtained from the diagram, when the difference of $(\Delta \Phi / (2 \Omega)) \times 10^4 \text{ m}^2/\text{s}$ between two stations $L$ km apart at latitude $\phi$ is known.
Figure 2-12. Location map (a) for the vertical profiles of temperature, (b), salinity (c) and sigma-t (d) in February, and temperature (e), salinity (f), and sigma-t (g) in August, 1975.
Figure 2-12 continued.
Figure 2-12 continue.
Figure 2-13. Horizontal distribution of the (Δϕ/Ω) x 10^4 m^2/s in August, 1973 (a) and 1975 (b). Nomogram for (Δϕ/Ω) x 10^4 m^2/s = 0.8 (c) and 1.0 (d). 7 °C isotherm depth in August 1973 (e).
Figure 2-13. Continued (different curves for different latitude $\phi$).
7 DEG. C ISOTHERM DEPTH IN AUG. 73

Figure 2-13. Continued.
Figure 2-13-e is the horizontal distribution of the 7 °C isotherm depth for the same period of time with the Figure 2-13-a (note that the 7 °C isotherm is usually an isotherm inside the thermocline in the Sea of Japan, e.g., Figures 2-12-b and 2-12-e). Comparing the distribution of the $(\Delta \phi/(2\Omega)) \times 10^4 \text{ m}^2/\text{s}$ with the horizontal distribution of the 7 °C isotherm depth, it is apparent that they mimic each other.

One can draw a few important inferences from this analysis. First, the geostrophic relative velocity is mainly attributable to the thermocline slope. Therefore, the geostrophic velocity should be maximum along the Polar Front. Second, large gradients of $(\Delta \phi/(2\Omega)) \times 10^4 \text{ m}^2/\text{s}$ near the Tsugaru Strait and the Soya Strait are always observed. This agrees well with the fact that most of the water flowing into the Sea of Japan flows out through the Tsugaru Strait or the Soya Strait. Conlon (1982) showed that geostrophic balance is a good approximation to the flow dynamics in the mouth of the Tsugaru Strait.

2.4 Discussion and summary

Density distribution in the surface layers of the Sea of Japan is largely influenced by the temperature. The thermal front is, thus, an excellent representation of the Polar Front. Not only can the Polar Front always be detected in winter by sea surface thermal signals but the marked differences of temperature between the cold and warm water sectors can be remotely monitored as will be discussed in the next chapter. A weak surface thermal front can sometimes be detected in
summer. Below 30 m depth, thermal fronts appear throughout the year (e.g., Figure 2-12). Temperature gradients observed in the frontal region are 1.0 °C per 10 km or more at the surface and about 2.0 °C per 10 km on deeper surfaces. These temperature gradient estimates must be underestimated because of the coarse hydrographic sampling grid. The Polar Frontal zone width is about 100 km or less and becomes narrower as depth increases. The position and the temperature gradients of the Polar Front show both seasonal and annual variability.

The meandering motion of the Polar Frontal zone is seasonally variable so that small amplitude (50 km), long wavelength (700 km) meanders occur in winter, and large amplitude (100 km), short wavelength (400 km) meanders occur in summer. This might suggest that the frontal wave in summer is unstable when the low salinity water flows into the Sea of Japan through the Korea Strait and baroclinicity along the front increases.

The geostrophic velocity shear calculated with respect to a reference level of 1000 db is larger near the front than any other place in the Sea of Japan. It is 20 to 25 cm/s/km along the front. The resemblance of the distribution of $\Delta \phi / (2 \Omega) x 10^4$ m²/s with the horizontal distribution of the 7 °C isotherm depth shows that thermocline topography is responsible for the large horizontal velocity shear along the front.

3.1 Introduction.

In the preceding chapter, I demonstrate that the Polar Front, particularly during winter, is well defined by its surface thermal signature. The past two decades have seen the development of a variety of satellite techniques which allow the rapid sampling of the entire sea surface temperature field. In this chapter, using satellite measurements, I discuss the measurement of scales of variability within the Polar Frontal zone which were either poorly resolved or not observed by classical hydrographic studies. In particular, I discuss eddies within the frontal region and variation of the position and shape of the Polar Front at annual and sub-annual periods.

Eddy, in physical oceanography, is a generic term. It includes all types of variable flows such as meandering and filaments of an intense current system, semi-attached and cast-off rings, advective vortices, planetary waves, topographic waves, wakes, etc. (Robinson, 1983). Eddy is often used interchangeably with the term turbulence (e.g., Harrison, 1983). Richardson (1983) claims rings to be the most energetic eddies in the ocean and Fuglister (1972) defines a ring in the Gulf Stream system as a special type of eddy formed from cut-off Gulf Stream meanders. Eddy in this study will mean an isolated and ring-like feature associated with the Polar Frontal zone turbulence.

Only suggestive evidence for the existence of eddies away from the coast was available until the 1970’s (Robinson, 1983). As pointed
out by Robinson (1983), ignorance of eddies was attributable to the lack of proper observation methods. The most promising technique is a remote sensing method such as satellite altimetry or AVHRR data, which will satisfy not only the need for synoptic data, but also that for fine spatial resolution.

Huh (1976) finds fronts associated with the coastal water along the southeastern coast of Korea using DMSP satellite. Since Legeckis (1978) showed that fronts in the ocean can be detected using VHRR data, the importance of the AVHRR of the polar orbiting satellites, such as TIROS-N and NOAA series satellites, has been stressed by many scientists (e.g., Bernstein, 1984). AVHRR can provide a synoptic sea surface thermal field with great detail by virtue of its high spatial resolution (1.1 km at satellite nadir point). Such resolution was previously impossible using only conventional in situ hydrographic observations. With proper calibration of the sensors and data, both the AVHRR and the multi-channel sea surface temperature (MCSST) techniques can give resolution as good as 0.5 °K rms (McClain, et al, 1985).

It is known that the surface thermal pattern is well reflected in subsurface layers of the Sea of Japan during the winter season due to strong wind mixing and negative heat flux at the sea surface. The wintertime monsoon also provides dry and cool air over the Sea of Japan, which is the best condition for obtaining very clear and accurate sea surface temperature estimates from AVHRR (Huh, 1982). These facts provide good reason to use AVHRR data in studying the dynamics of eddies and circulation in the Sea of Japan.

Level 1-B tapes of AVHRR data in the Sea of Japan were acquired
from the National Environmental Satellite, Data, and Information Service (NESSDIS). Historical AVHRR pictures from NOAA 6, 7 and 8 during the 1978 - 1984 periods were also available. For analytical purposes, clear images were selected by eye, and the level 1-B data analyzed on the computer using the NASA Earth Resources Laboratory Application Software (ELAS) package on a Compton 5000 image analyzer and Perkin Elmer 3220 host computer.

3.2 AVHRR Data Analysis.

Due to both poor quality of the historical AVHRR images and severe cloud cover in the area of interest, only one time series of fronts showing eddies in the frontal zone were obtained, in spite of 6 years of data accumulation. Hopefully, these problems will be overcome with the routine use of "all-weather" microwave sensors. Two methods are used in the analysis due to the nature of the data on hand. One is to qualitatively and subjectively analyze the historical photographic data. No data editing or processing has been done on the historical data. The best pictures were selected by visual inspection among the large volume of data. (Note that there are very distinctive patterns of SST and also very distinctive patterns of clouds. So, evaluation of imagery by eye is very effective. Visible channel data also allows objective separation of cloud pattern from water patterns). The other is to process level 1-B data on the computer using the ELAS image processing system with a few steps, such as data calibration, which convert sensor signal (voltage) into gray levels, image enhancement to get the best contrast between different water masses, and image
Figure 3-1. Enhanced TIROS-N image, orbit 2317, channel 4, on March 26, 1979 (a), and after applying the Sobel edge operator for better contrast (b).
expanding to focus on the area of interest.

Figure 3-1-a shows enhanced black and white thermal imagery from the Sea of Japan, dated March 26, 1979 (TIROS-N, orbit 2317). To obtain the best contrast, an arctangent transfer function is applied to the imagery. Because of the noise existing in the imagery the frontal region is still blurred. Figure 3-1-b is a better image showing the clear boundary of the fronts after applying the Sobel edge operator (Ballard and Brown, 1982), defined as

\[
\begin{bmatrix}
1 & 2 & 1 \\
0 & 0 & 0 \\
-1 & -2 & -1
\end{bmatrix}
\]

where each number is a weight in a 3 x 3 window. This is a gradient operator which reduces the effects of noise by taking a local average. One slides the operator (or window) over the picture function, each weight is multiplied by the gray level at the corresponding picture element (pixel) and the average is taken over the window. The image clearly shows, for example, that the EKWC is branching from TTCS, carrying warm water to the north and losing heat to the surrounding water (see the changing shade of the EKWC along the Korean coast in Figure 3-1-b). A sector of this image is selected for spectral analysis (Figure 3-2).
Figure 3-2. Selected area (white box) for further spectral analysis (a) and enlargement of the selected area(b).
3.2.1 Results of the level 1-B data.

Wavenumber spectra of the sea surface temperature (SST).

The SST can be calculated for each pixel using the relationship (DiRosa, 1986; private communication)

\[ T = G \times 0.25 - 19.0 \]

where \( T \) is the SST in degrees C, and \( G \) is the pixel gray level value varying from 1 to 256. (Temperatures in Figure 3-1-a are obtained by this equation). Figures 3-3-a and 3-3-b are temperature profiles along and across the front, respectively. Sampling lines in the selected area (Figure 3-2) are shown in Figure 3-3-c. The numbers of points sampled are 400 and 300 along and across the front, respectively. The sampling interval is 1.1 km in both directions, which yields a Nyquist wavelength of 2.2 km. As can be seen in the temperature profiles, there are increasing temperature trends southward and westward in the frontal zone. Trends were removed from each profile by the least squares method assuming a first order polynomial fit.

The wavenumber spectra of the SST, calculated by the Cooley-Tukey or FFT method, with degrees of freedom of 20 and 18 along and across the front, respectively, are shown in Figure 3-4. The three-dimensional inertial range theory, as originally proposed by Kolmogroff (1941) to describe the statistical small-scale behavior of locally isotropic turbulence, results in a \( k^{-5/3} \) spectrum for velocities. Since then, numerous studies have shown that a \( k^{-5/3} \) law behavior is observed out to much larger scales than can be
Figure 3-3: SST profiles along (a) and across (b) the frontal zone for the selected area shown in Figure 3-2.
Figure 3-3. Continued
Figure 3-4. Wave number spectra of SST of the along-frontal lines (a) and the across-frontal lines (b). Line numbers as shown in figure 3-3-c.
considered three-dimensionally isotropic (e.g., Ellsaesser, 1969). Kraichnan (1967) showed that two dimensional turbulent velocities also obey a \( k^{-5/3} \) law and enstrophy (half of the squared vorticity) follows a \( k^{-3} \) law. Gage (1979) provided observational evidence that two dimensional turbulence exhibits a \( k^{-5/3} \) law. In isotropic turbulence, advected thermal fluctuations have the same shape spectrum as the velocities doing the advection (Phillips, 1980). For both north-south and east-west directions, spectra of the SST fluctuations are very similar and decay at a rate between \( k^{-5/2} \) and \( k^{-5/3} \) over the whole range of wavenumbers. This suggests that turbulence in the frontal zone has characteristics of two-dimensionally isotropic turbulence (e.g., Mode Group, 1978; Bernstein and White, 1974; Fu, 1983). No significant spectrum peak is observed.

To investigate the structure of eddy activity in the Polar Frontal region, two-dimensional Fast Fourier Transform (2-D FFT) spectra are calculated from the data in the boxed area shown in Figure 3-2-a. Figure 3-2-b shows an expanded image of the sampling area. Again, the number of pixels involved in the box are 400 and 300 along and across the front, respectively. Since the general trend of the temperature field is increasing toward the south and west, trends are removed by least squares fitting a quadratic surface to the data before calculating the two dimensional wavenumber spectrum of the residual SST field. The least squares quadratic surface fit of the input data is obtained using the PROC RSREG in the SAS users guide (SAS Institute, 1982). The SST residuals are subsampled at every 9 pixels in both directions giving an effective sampling interval of 9.9 km. This gives a linear Nyquist wavenumber of 0.05050 km\(^{-1}\). 2-D FFT is
Figure 3-5. 2-D FFT spectra with 8 degrees of freedom.
calculated by calling the subroutine FFT3D in the IMSL Library (1982) Subroutine package. Figure 3-5 shows the 2-D wavenumber spectrum of the SST residuals, where wavenumber smoothing provides independent estimates with 8 degrees of freedom (note that spectra in Figure 3-5 are logarithmic values). The low energy levels near the origin are the result of the quadratic trend removal. Significant lobes of energy, though, remain along the two axes of the figure. In fact, isolated peaks occur at wavenumbers of (0.0136, 0.0) and (0.027, 0.0) and a plateau near (0, 0.0182). These wavenumbers correspond to wavelengths of approximately 45 km and 75 km in the east-west direction and 55 km in the north-south direction, respectively. There are four other lesser, but significant, peaks in the $k_1 > 0$ half-plane. The associated wavenumbers are $k' = (0.0275$ to $0.0321, 0.0189$ to $0.0253), k'' = (0.0321, -0.0189), k''' = (0.0, 0.0378)$, and $k^{iv} = (0.0413, 0.0)$. Associated wavelengths are about 25 to 30 km for these wavenumbers. The wave directions are $+30$ to $+45$ and $-30$ degrees with respect to the x-axis for $k'$ and $k''$, respectively, and 90 and zero degrees for $k'''$ and $k^{iv}$, respectively.

2-D structure of the eddies.

To determine the shape, size and intensity of the eddies, the gradient of the image was calculated for the same data as used in the 2D-FFT analysis. Because the gray levels in the object are similar, the boundary of the object, where the rate of change of the gray level is locally maximum, is more clearly delineated by application of the gradient operator. Let $f(x,y)$ be a picture function at grid point
(x,y). Then the gradient is defined as

$$\nabla f(x,y) = \frac{\partial f}{\partial x} \hat{i} + \frac{\partial f}{\partial y} \hat{j}$$

where \(\hat{i}, \hat{j}\) are the unit vectors in x, and y direction, respectively.

Because of the expense of this calculation, the magnitude of the gradient is approximated as (Castleman, 1979)

$$| \nabla f(x,y) | \approx \max \left[ |f(x,y) - f(x+1,y)|, |f(x,y) - f(x,y+1)| \right]$$

where | . | stands for absolute value, and max means take the maximum value among the values in the square brackets. Figure 3-6 shows the results after applying the gradient operator to the image. The light tone indicates a small gradient (0.0 - 0.1 °C/km), while the darkest tone indicates a gradient larger than 1.0 °C/km. The lengths of the axes of an eddy can be subjectively estimated by counting the number of pixels involved. The frontal zones, demarcated by the upper and lower boundaries, are clearly differentiated with a width of about 100-200 km. The size of the eddies is approximately 50-100 km, in agreement with the results of the two-dimensional spectrum analysis described above.
Figure 3-6. Result of application of the gradient operator to the selected area (Figure 3-2).
3.2.2 Results of the historical AVHRR data.

Eddies in the frontal zone.

Figures 3-7-a to 3-7-g show a time series of eddies at the western end of the frontal zone. Figures 3-7-a to 3-7-f are line drawings from the historical photographs. Figure 3-7-g is the image from which Figure 3-7-f is drawn. The cold water region is dotted. On February 11, 1982, there is no eddy observed near the Korean coast (Figure 3-7-a). However, a streak of warm water (the streak S1) intrudes into the cold water region, and begins to bend anticyclonically. About three weeks later on March 3, 1982, a pair of eddies (the eddies E1 and E2) appear fully evolved at the same location with a diameter of about 20-30 km (Figure 3-7-b). These two eddies are circular, and appear to be rotating anticyclonically. A week later on March 10, 1982, this pair of eddies (the eddies E1 and E2) no longer appears to be active (Figure 3-7-c). The remnant eddies E1 and E2 are no longer circular and seem to have stopped rotating even though they have not completely disappeared. They (the eddies E1 and E2) again begin to evolve on March 28, 1982, and take on a much more active circular shape (Figure 3-7-d). A week later, this pair of eddies has developed so as to be distinguishable from the ambient water (Figure 3-7-e), but then they retain the same SST signal intensity and same shape without further evolution until April 21, 1982 (Figure 3-7-f). No similar time-series of eddies is found after this date in our historical photographic data. A few other isolated eddies can be identified in Figures 3-7-d, e, and f. These relate to the entrainment of warm water into the cold water region and
Figure 3-7. Hand drawings (a - f) of the AVHRR image by tracing the boundaries of the warm water showing eddies and streaks. Date, time and orbit numbers are shown in the left upper corner of each figure. NOAA-6 image on 21, April, 1982 (g). (Note that Figure 3-7-f is a hand drawing of Figure 3-7-g).
Figure 3-7. Continued
Figure 3-7. Continued
Figure 3-7. Continued.
are associated with streaks S2, S3, S4, and S5. It is impossible to
decide whether these eddies are migrating or not because the temporal
resolution of the successive images is not fine enough to track the
eddies' evolution in space. Whether or not they are migrating, they
must be losing heat to the cold water. (Note that all pictures are
based on channel-4 of the NOAA-7, except Figure 3-7-b, which is from
channel 4 of the NOAA-6. Orbit number is on the upper left corner of
each figures). This series of pictures of eddies is not good enough to
definitively characterize the life cycle of the eddies in the frontal
zone. However, it is very clear from these pictures that the eddies
are not migrating much in the area involved, and seemed to have a
lifespan of a month or so.

An interesting feature from the historical data is the character
of the mixing of warm water and cold water. For example, Figures 3-7-e
and 3-7-f show streaks of warm waters (S2, S3, S4, and S5) in the cold
water region, suggesting entrainment of the warm waters into the cold
water region. It is inferred that transport of momentum, heat and salt
into the cold water region is by entrainment rather than by migrating
warm core eddies. Figure 3-7 shows that the polar frontal zone is full
of eddies of different sizes. Clearly, further study at different
scales of motion is necessary.

3.3. Guam 7-day SST analysis.

Satellite derived SST data are useful in detecting synoptic ocean
features, especially the thermal signatures of fronts and eddies.
(Halliwell and Mooers, 1979; Olson et al, 1983). Since satellites
provide synoptic thermal fields with very accurate temperature resolution, satellite derived SST data can provide very important information concerning the kinematics and dynamics of ocean features reflected in the ocean surface thermal fields. The high repetition rate at which such data is collected makes it ideal for the study of temporal evolution on scales previously unobtainable.

The Naval Oceanographic Command Center (NOCC)- Joint Typhoon Warning Center, Guam issues weekly sea surface temperature (SST) analysis charts, which present contoured SST based on the DMSP (Defense Meteorological Satellite Program) temperature data supplemented with ground truth data provided by surface ships and aircraft. The Guam 7-day SST analysis are dated on Tuesday of each week. The Guam 7-day SST analyses is preferred over the 10-day Marine Report, which is issued every 10 days by the Japan Meteorological Office, for three reasons. The Sea of Japan in the 10-day Marine Report is of such small scale (about 1/5 of the NOCC scale) that spatial resolution is poor and temperature fields are too smooth to allow detection of any detail in the Polar Front. The Guam 7-day SST analyses present a more detailed thermal pattern because they are mainly derived from satellite IR images. The Guam 7-day SST analyses have better temporal resolution than the 10-day Marine Report, i.e. there is a 50 % decrease in the Nyquist frequency.

125 weeks of Guam 7-day SST analyses from January 1983 to August 1985, are subjected to spectrum analysis. Daily mean wind observations at Kangreung and Ulleungdo Islands, and daily sea level records at Pusan and Pohang observed during the same period as the Guam 7-day SST analyses, are also analyzed.
3.3.1 Data processing

Digitization.

Figure 3-8 shows the portion of an original Guam 7-day SST analysis chart covering the Sea of Japan from which I digitized the thermal front associated with the Polar Front. This figure is reduced 71 percent from its original size. Both compilation date and the initials of the analyst are marked on the chart. The "frontal position" is quasi-objectively identified on each chart. When an even number of isotherms is subjectively deemed to define the frontal zone, the center line drawn between the two most central isotherms is considered as the "frontal position". When an odd number of isotherms is involved, the middle isotherm is chosen as the "frontal position".

A few further assumptions are made for the analysis. First, the frontal positions are defined only in the area between 130 E to 140 E. Second, the frontal position is assumed to be continuous even though there are occasionally small segments along the frontal zone in which the temperature gradient is not large enough to define a frontal position, presumably due to thermal diffusion or mixing with the surrounding waters. In the summer time, however, the frontal zone near Hokkaido is so diffuse that the frontal position can not be defined. Once defined, the frontal position is digitized on the Calcomp 9000 digitizer. Figure 3-9-a shows the coordinate system for the digitization. The x'-axis is aligned in such a manner that it is positive to the east along 35°N, and the y'-axis is positive to the north. I set the origin of the coordinate system roughly in the center.
Figure 3-8. Guam 7-day SST analyses chart (71% reduced). Analyst and dates are shown in the upper left corner.
Figure 3-9. Coordinate system for analysis of the frontal position (a) and frontal position sampling strategy (b).
of the Korea Strait.

**Rotation of the axis.**

The digitization process actually records the longitude and latitude of the frontal position. The displacement of the front from the $x'$-axis is obtained by converting the longitude and latitude coordinates into distance. It is known that the Tsushima Tsugaru Current System (TTCS) flows into the Sea of Japan through the Korea Strait and, principally, out through the Tsugaru Strait. It is therefore reasonable to assume that a natural coordinate system for the Polar Front in the Sea of Japan involves an axis connecting the center of the Korea Strait to that of the Tsugaru Strait and the perpendicular to this line.

Let $D(t,x)$ be the frontal position displacement with respect to the OXY coordinate system, where $t$ is time and $x$ is distance in the $x$-direction, and let $D'(t,x')$ be the frontal position displacement in the OX'Y' coordinate system as shown in Figure 3-9-a. Then $D(t,x)$ can be calculated from $D'(t,x')$ using the rotation matrix operator

$$
\begin{bmatrix}
D(t,x), x
\end{bmatrix} =
\begin{bmatrix}
D'(t,x'), x'
\end{bmatrix}
\begin{bmatrix}
\cos(\theta) & \sin(\theta) \\
-sin(\theta) & \cos(\theta)
\end{bmatrix}
$$

where $\theta = \Delta XOX'$. 

Sampling of the frontal amplitude.

$D(t,x)$ was subsampled every 20 km along the x-axis. Pseudo-sampling stations are shown in Figure 3-9-a. Figure 3-9-b shows the sampling strategy. Occasionally $D(t,x)$ was multiple-valued due to the meandering motion of the front. The minimum value of $D(t,x)$ was always selected for further analysis in all multiple-valued cases. Because of mixing, it is occasionally impossible to define $D(t,x)$ for stations southwest of station 13 or northeast of station 35. These stations are then not used in the spectrum analysis.

Error sources in data processing.

Because of the many steps involved in getting final frontal estimates of $D(t,x)$, a large number of error sources are incurred. These include Guam 7-day SST analysis chart errors, frontal position estimate errors, and digitizing errors. All involve human errors. The NOCC SST analyst changes once or twice each month. However, it is not possible to attribute any abrupt changes in the contoured SST field to the change in analyst. Other possible sources of error are the different input data sources used when preparing the SST chart. I have not attempted to estimate these. When the frontal position is drawn, errors may occur because it is not possible to exactly follow the midline between two isotherms. However, such errors are far less than the spacing of the Guam isotherm estimates in the frontal zone, which is about 10 km. According to Bendat and Piersol (1971), the error due to digitization is usually unimportant relative to other sources of error in data processing procedures. The rms value of the quantization
error due to the digitizer is approximately 0.29 \( \delta x \) (Bendat and Piersol, 1971), where \( \delta x \) is the scale unit of the quantization. Hence, the rms quantization error is about 0.3 km based on \( \delta x = 1 \) km. This is quite negligible. Human error introduced in tracing the frontal position during digitization has a rather larger error than this. We assume that the error in tracing the frontal position is the same as that in drawing it, i.e., far less than 10 km. I estimate the total error as less than 5 km which will not affect the analysis since typical amplitudes of the front displacement are an order of magnitude larger than the estimated total errors due to data processing.

3.3.2 Methodology

Two analyses of the data are presented. One gives the mean and rms variability of the frontal position. Another analyzes the autospectra and cross spectra to investigate the spatial and temporal scales of variability of the front. The FFT (Fast Fourier Transform) method is used to estimate wavenumber spectra, and Blackman-Tukey methods are used for other spectrum analyses.

The overall mean frontal position is estimated from the whole data set of 125 weeks at 200 subsampling points. The seasonal mean frontal position is estimated from the data available for each three month season (e.g., winter is based on December, January, and February).
3.3.3 Results.

**Mean frontal position and its variability.**

Figure 3-10 shows the overall mean frontal position. In the west, the mean front begins at 38.5 N and has a sinuous shape with its axis tilted east-northeastward, ending near 41.5 °N. The rms variability of the mean frontal position is 100 - 120 km over the western two-thirds of the front, and increases to 300 km near the Tsugaru Strait (Figure 3-11). This does not agree with the existing wisdom, which suggests high variability at both ends of the front and little variation at the center (Istoshin, 1960). The high variability near the Tsugaru Strait agrees well with the fact that the TTCS flows out through the Tsugaru Strait in winter, and has significant outflow through both the Tsugaru Strait and the Soya Strait in summer (Moriyasu, 1972).

Seasonal mean frontal positions are shown in Figure 3-12. The mean frontal position in summer is about 100 km north of that in winter except near the Tsugaru Strait, where it is about 350 km further north. The mean frontal positions in spring and fall are almost collinear. This suggests that the fluctuation of the mean front is seasonal, moving north-south with respect to the annual mean frontal position.

**Spectral analyses.**

Figure 3-13 shows wavenumber spectra with 12 degrees of freedom for selected months. Wave number dependency of the spectra is
Figure 3-10. Mean frontal position from all data.

Figure 3-11. Rms variability of the mean frontal position. Distances are calculated from the western most point of the front in Figure 3-10.
Figure 3-12. Seasonal mean frontal position.  
Sp ; (March, April, May), Su ; (June, July, August),  
F ; (Sept., Oct., Nov.), W ; (Dec., Jan., Feb.).
approximately $k^{-1}$ for all years. The energy spectra show interannual variation; energy levels in 1983 and 1984 are one order of magnitude larger than those in 1985 at the longest wavelengths. Seasonal variability in 1984 is so large that the energy levels for August are almost one order of magnitude larger than those for February (Figure 3-13-b), while records from 1983 and 1985 do not show large seasonal variability.

Autospectra with 21 degrees of freedom are estimated for each station. The spectra are red, being dominated by frequencies smaller than 0.01 cpd (cycle per day). Secondary peaks near $0.03 \pm 0.005$ cpd are present at almost all stations. Thus, the energy of the frontal motion is predominantly associated with seasonal or longer periods with a secondary periodicity of approximately one month. Since seasonal and annual variability of the fronts are already known to be important, the one month periodicity draws attention. Cross spectra of all possible combinations of stations have been obtained. For example, cross spectra between stations 18 (X) and 14 (Y), and between stations 18 (X) and 22 (Y) are shown in Figures 3-14-a and 3-14-c. As can be seen in Figures 3-14-a and 3-14-c, both autospectra and cross spectra at low frequencies are one order of magnitude larger than those at frequencies near 0.03 cpd. Figures 3-14-b and 3-14-d show the coherence squared and phase of the corresponding cross spectra. The 95% significance level for coherence squared based on 21 degrees of
Figure 3-13. Wave number spectra for selected months F (Feb.), M (March), A (August), and N (Nov.) of 1983 (a), 1984 (b), and 1985 (c).
Figure 3-14. Auto- and cross- spectra (a and c), and coherence squared and phase (b and d) between stations 18 (X) and 14 (Y), and between stations 18 (X) and 22 (Y), respectively.
Figure 3-14, continued
freedom is 0.27. The confidence interval, CI, for the phase is calculated, when the coherence squared is significant, using the equation of Jenkins and Watts (1968),

$$CI = \Phi_{xy}(f) \pm \sqrt{\frac{2 F_{(1-\alpha)^2}, v (1-\alpha)(1-G_{xy}(f))}{(v-2) G_{xy}(f)}}$$

where $\alpha$ is the probability of a Type I error, $\Phi_{xy}(f)$ the phase at the frequency $f$, $v$ the equivalent degrees of freedom, $G_{xy}(f)$ the coherence squared at a frequency $f$, and $F_{(1-\alpha)^2}, v$ is the F-statistic with degrees of freedom $(2, v)$ for a confidence level of $(1- \alpha)$. CI for a 95% confidence level is marked on the phase diagram by a vertical bar. Figure 3-15-a shows the distribution of coherence squared and phase of the cross spectra between station 18 and other stations at a frequency of 0.03125 cpd. A coherence length scale of about 220 km is estimated from the figure. The phase between stations is almost zero over this length scale, suggesting that the waves are stationary since the time lag, $\tau$, between two stations at frequency $f$, as given by

$$\tau(f) = \Phi_{xy}(f) / (2 \pi f)$$

becomes zero as $\Phi_{xy}(f)$ goes to zero. Figure 3-15-b is a similar plot based on station 34. Cross spectra at station 34 are not coherent with stations other than those between 30 and 34. This suggests that the dynamics, at the monthly time scale, are different in the eastern and western portions.
Figure 3-15. Coherence squared and phase estimated from the cross spectra between station 18 and all other stations (a) and between station 34 and all other stations (b) at frequency 0.03125. The solid lines are for coherence squared and dotted lines are for phase.
of the Sea of Japan.

If there were no disturbances by winds and ocean currents, the isotherms in the ocean would be basically zonal because the solar insolation decreases from the equator polewards. A given isotherm, for instance the 10 °C isotherm in the Sea of Japan, would move northward or southward depending on the seasonal variation of the solar insolation.

Winds modify this temperature distribution and influence frontal dynamics in many ways, both directly and indirectly. Convergence induced by the wind-driven Ekman transport can intensify a front. Winds supply kinetic energy to the frontal circulation by shearing stresses, or cause the ocean surface layer to become turbulent, resulting in deepening the surface mixed layer and hence dropping the sea surface temperature. Winds also contribute to air/sea heat flux by affecting evaporation and sensible heat transfer. Both heat flux and winds in the Sea of Japan are in phase, i.e., they cause the isotherms to seasonally move in the same direction. Yoon (1982) states that seasonal variation of the Polar Frontal position in the Sea of Japan is due to seasonal variation of the atmospheric conditions.

Both autospectra and cross spectra of the frontal position show that secondary peaks occur at a period of about one month (Figure 3-14). Wind stress and sea level records have been analyzed to determine if they are significantly related to frontal position on a monthly time scale. Daily mean winds observed at Kangreung and Ulleungdo Island and sea level observations at Pusan and Pohang are available from the period of January, 1983 to February, 1985,
overlapping the Guam 7-day SST time series for 112 weeks. Since the sampling interval of the Guam 7-day SST analysis is one week, weekly averaged wind stress and sea level data are computed. Wind stress is computed from

\[ \tau = \rho_a C_d |U| U \]

where \( \rho_a \) is the air density (\( \approx 1.2 \times 10^{-3} \text{g/cm}^3 \)), \( C_d \) is a constant drag coefficient of \( 1.5 \times 10^{-3} \), and \( U \) is the wind vector. Wind stress is estimated from weekly mean winds. Adjusted sea levels are obtained by correcting for the inverse barometer effect. To avoid leakage from the strong secular signal, data with periods longer than the seasonal scale are filtered from the record before calculating cross spectra.

Cross spectra between wind stress and frontal positions at each station have been estimated. \( \tau_x \) and \( \tau_y \) at Ulleungdo Island are incoherent, at 95% significance level, with frontal position at all stations in the frequency band of interest. The same is true for the wind stress measured at Kangreung.

Adjusted weekly mean sea level, at both sites, is also incoherent, at the monthly time scale, with frontal position with one exception. The unadjusted weekly mean sea level at Pusan shows significant coherence with frontal position at station 34, only (Figure 3-16). The phase in this band is not significantly different from zero. This means that the frontal position at station 34 moves northwestward as sea level at Pusan increases. According to Lisitzin (1967), when sea level is maximum at Pusan in summer, sea level at the Japan coast in the Korea Strait is higher than the sea level at Pusan. This induces the maximum
Figure 3-16. Auto- and cross- spectra (a) between station 34 (X) and unadjusted sea level at Pusan (Y) and the coherence and phase (b). CI based on 95% significance level and 21 degrees of freedom is marked on the phase diagram by a vertical bar when it is significant.
geostrophic mass transport into the Sea of Japan through the Korea Strait (Yi, 1966). If similar dynamics hold at the monthly scale, then the northward movement of the frontal position at station 34 should reasonably be a response to increased flow through the Korea Strait.

3.4. Discussion.

It is clear that the use of modern satellite technology for the remote sensing of SST fields has greatly improved our understanding of the variability associated with the Polar Front in the Sea of Japan. A broad eddy field associated with the Polar Front has been identified and time and space scales of the eddy structure determined. The annual cycle of the position of the Polar Front has been characterized and a fundamental difference in, at least, the intensity of processes at the eastern and western ends of the Polar Front suggested. Finally, variability of the frontal position at a time scale of one month is observed, although I am unable to attribute the source of this variability to wind or water level forcing.
CHAPTER 4. SEASAT Altimeter data.

4.1 Introduction

In the preceding chapter I discuss the utility of satellite infra-red imagery in defining the sea surface temperature fields. This data allows one to locate features, such as the Polar Front in the Sea of Japan, and to track their evolution and migration. Many of these features are associated with strong currents. It would greatly enhance the understanding of the oceans, if one were able to remotely monitor these currents. In this chapter I discuss one technique which is potentially capable of satisfying this need with emphasis on data from the Sea of Japan.

Geostrophic equilibrium is a first order solution of the momentum equations (Pedlosky, 1982). The principal geostrophic components of the ocean circulation are a quasistationary large scale component reflecting the mean flow, and a mesoscale eddy component which varies from a few tens of kilometers to a thousand kilometers in scale (Wunsch, 1981). In the upper layers of the ocean, the non-geostrophic wind-driven Ekman drift is superimposed on the geostrophic ocean currents. The ocean surface height that is associated with the total current can be measured, in principle, by radar altimetry. The radar altimeter measures the range to the sea surface near the satellite ground track. The altimeter transmits a radar pulse towards the subsatellite point and measures the time for the pulse to be reflected and returned to the satellite. The total range error includes, among
others, variation in the atmospheric path length of the pulse returned to the satellite due to variation in the speed of electromagnetic waves.

Precise radar altimetric measurements have been made over the last decade from the Skylab, GEOS-3, SEASAT, and GEOSAT satellites. Precision achieved by the altimeter aboard the GEOS-3 satellite was about 50 cm, which was the first sufficiently precise measurement to show the height differences across the Gulf Stream (Leitao et al., 1979). In 1978, a significantly more precise (about 10 cm) altimeter operated aboard the SEASAT satellite (potentially greater precision was proposed by Born et al., (1984) as 5 cm for SEASAT, 30 cm for GEOS-3, and 1 m for Skylab).

The SEASAT altimeter successfully began to function on June 27, 1978 and collected useful data until it experienced a short circuit in the power system on October 10, 1978. The altimeter was turned off twice during the mission due to hardware problems (Townsend, 1980). The SEASAT satellite was originally equipped with 5 sensors: a radar altimeter, SEASAT-A Satellite Scatterometer (SASS), a Synthetic Aperture Radar (SAR), a Visible and Intra-Red Radiometer (VIRR), and a Scanning Multichannel Microwave Radiometer (SMMR). It flew in a nearly circular orbit with an inclination angle of 108° at an altitude of 800 km. SEASAT circled the earth approximately 14 times a day (a period of about 101 minutes) covering 95% of the global oceans every 36 hours with two of its wide swath sensors. Data transmitted to earth by SEASAT included information on sea surface winds, wave height, sea surface temperature, atmospheric water, sea ice, ocean topography,
and the marine geoid.

A short pulse (3 ns), nadir-viewing altimeter operating at 13.5 GHz was the first instrument selected to be incorporated onto the satellite. The unique features of the satellite altimetry data were two: the high accuracy to which surface topography, wave height, and wind speed could be measured, and the ability to collect these measurements globally over the time interval of a few days. The instrument field of view (FOV) varied from 2.4 km to 12 km, depending on the sea state. The precision of sea height measurements was expected to be at least 10 cm for a sea state less than 20 m (Tapley et al., 1982).

Bernstein et al. (1982) have compared sea surface topography over the Kuroshio extension using SEASAT altimeter data with the surface dynamic height inferred from Air-Expendable Bathythermograph (AXBT) data. They show that the variation in the altimeter-determined topography agree to within 10 cm of those derived from the AXBT data. Cheney and Marsh (1981a) and Cheney (1982) show that the location of the Gulf Stream as well as the sea surface topography associated with cyclonic or anticyclonic rings are detectable in the altimeter data.

It is difficult to obtain a synoptic picture of the ocean circulation by classic in situ measurements only. Previous results (Cheney and Marsh, 1981a; Douglas and Gaborsky, 1979; Bernstein et al. 1982; Rizos, 1981; Menard, 1983) show that it is feasible to determine dynamic heights of the sea surface with respect to a reference geoid using altimetry. The geoid (equipotential surface) in the western North Atlantic Ocean is known to be sufficiently accurate.
to allow systematic estimates of geostrophic current velocities from satellite data (Cheney, 1982). Watts and Leeds (1977) present a gravimetric geoid in the North Pacific Ocean including the Sea of Japan based on surface ship and pendulum measurements. Measurements are, however, not available from the Northwestern half of the Sea of Japan, where bottom topography is complex along the continental slope, especially from near the Korea Strait to Wonsan Bay, Korea. According to Watts and Leeds, the geoid over the ocean, in this area, does not agree with the satellite derived model from Skylab and GEOS-3 altimeter data, while that over the continents agrees well. Figure 4-1 shows the bottom topography (the continuous lines) and the gravimetric geoid (the shackled lines) in the Sea of Japan, where the gravimetric geoid is digitized from Watts and Leeds (1977).

One of the difficulties in applying altimeter data to dynamical oceanography lies in the fact that the spatial geoid variability whose amplitude is on the order of tens of meters is superimposed on the ocean topography spatial variability of which the amplitude is on the order of only 1 meter (Wunsch and Gaposchkin, 1980). There are two methods to remove time invariant geoid signals from altimetry data; the crossing arc method (e.g., Huang et al., 1978; Cheney and Marsh, 1981 b) and the repeat track method. (Douglas and Gaborski, 1979; Gordon and Baker, 1980; Cheney and Marsh, 1981 a; Douglas and Cheney, 1981; Bernstein et al, 1982; Cheney et al., 1983; Fu, 1983; Thompson et al., 1983).

The initial SEASAT orbit produced altimetric coverage with a grid size of approximately 1.5°, and the repeating tracks were not closely
Figure 4-1. Bottom topography (solid lines), geoid distribution (shackled lines) from Watts and Leeds (1977), and the repeat tracks of the SEASAT (labelled AA', BB', and CC' for group 1, group 2, and group 3, respectively).
collinear (a deviation of 18 km may occur between the nearest passes; Menard, 1983). The orbit was later modified so that from September 13 to October 10, the grid size was 900 km at the equator and 700 km at midlatitude. These passes were repeated within 3 km every 3 days (Bernstein et al., 1982), producing sets of 8 or 9 nearly collinear passes, depending on the region.

4.2 Data correction

The precision of the altimeter is 10 cm rms or better. Besides the mean sea level signal, though, numerous contaminating signals are included in these measurements. Comparable accuracy could be achieved for the mean sea surface after applying appropriate corrections for errors. These corrections involve (1) instrument corrections, (2) a propagation medium correction, (3) a geoid model, (4) effects of temporal variation in the ocean surface including solid earth tides, oceanic tides, and barometric pressure effects, and (5) the satellite orbit. The altimeter data used for this study are obtained from the Jet Propulsion Laboratory (JPL). The SEASAT geophysical data record includes sea height with respect to the reference ellipsoid (corrected for instrument effect and errors due to ionospheric effects and solid earth tide) and significant wave height estimated from the shape of the return signal (private communication, Dr. J. Mitchell of NORDA, NSTL, MS, 1985).

Figure 4-2 is a schematic representation of the altimeter geometry. The equation for further error correction used in this study
Figure 4-2. Schematic representation of altimeter geometry.

$H$ : Ocean surface topography.
$H_{al}$ : Altimeter height from sea surface.
$H_{g}$ : Geoid height with respect to the reference ellipsoid.
$H_{sat}$ : Satellite height with respect to the reference ellipsoid.
where $H_{cr}$ is the corrected sea surface height relative to the reference ellipsoid, $H_s$ is sea surface height measured by the altimeter after correcting for instrument errors, ionospheric errors, and solid earth tide effects, $SWH$ is significant wave height, $TIDE$ is a tidal model correction, $FNOC$ is wet tropospheric correction based on the Fleet Numerical Oceanographic Center (FNOC) meteorological data, and $P_a$ is a barometric pressure correction. 3% of $SWH$ is used to account for effects due to surface waves (J. Mitchell, 1985, private communication). The inverse barometer correction is made assuming a direct proportionality of 1 cm/mb in the change in the sea surface height due to changes in the sea surface atmospheric pressure. The SEASAT SMMR wet tropospheric correction, due to its higher spatial resolution, is known to be better than FNOC wet tropospheric correction (J. Mitchell, 1986, private communication). However, due to numerous missing values in the SMMR data, the FNOC correction was chosen for this study. Tapley et al. (1982) find that the rms difference between SMMR and radiosonde corrections is 2.7 cm, while the rms differences between the corrections derived from the FNOC and radiosonde data is 5.5 cm. Further corrections should be made for the marine geoid and satellite ephemeris error. The largest single signal (or error source) is due to the geoid. The radial orbit error is predominantly long wavelength, being characterized by a once per
resolution signature.

Further data correction are applied to $H_{cr}$ (corrected height) separately for the repeat orbit data and the earlier data. For the repeat orbit data, the average height of each group (track), which contains all long wavelength features such as geoid and mean sea surface topography, is removed from the individual profiles of each group. A final step is necessary for removing the residual orbit error. Radial uncertainties of approximately 1 meter are known to exist in the computed SEASAT orbit (Cheney and Marshy, 1981a). Since the arc length scale in this study area is at most 1000 km, removing the linear trend can eliminate most of the uncertainties in orbit determination (J. Mitchell, 1986, private communication). Now, this final error free data set is suitable for the analysis of mesoscale temporal variability. (Note that information about the mean flow was lost with the geoid.)

Cross over differences, defined as the discrepancies in sea height at the intersection of the descending and ascending tracks, are primarily attributable to the orbit error since the geoid is time invariant. The rms cross over difference based on the 156 cross over points (Figure 4-3) is 169 cm. (The repeat tracks were excluded from this analysis to eliminate a possible bias due to their high density). SEASAT radial orbit accuracy is therefore assumed to be 169 cm for our analysis. The cross over differences at all cross over points were used to calculate the best fit plane by the least squares method. Orbit error was then removed by subtracting this plane from $H_{cr}$. 
Figure 4-3. Detrend sea height of cross-over tracks at cross over points. Study area is divided into 6 subregions to estimate the associated rms variability.
Figure 4-3 shows the results of this calculation. The contoured data include both the geoid and mean sea level topography due to currents.

4.3 Data analysis.

Figure 4-4 shows the SEASAT tracks over the Sea of Japan during the SEASAT mission. Three groups of repeat tracks, which are labelled A-A', B-B', and C-C', are composed of 8 tracks each. Group 1 (A-A', where the unprimed label is for the starting position and primed one is for the ending position) and group 3 (C-C') are ascending (north bounding) tracks and group 2 (B-B') is descending (south bounding) tracks. (The rest of the tracks are earlier SEASAT passes, which were not repeated.)

After geographically aligning each track within a group, starting points of each group are not separated by more than 1.5 km. The distances along the tracks with respect to the origin or common starting point of the group are calculated based on the US Hydrographic Office standard distances of longitude and latitude for a given latitude. Figure 4-5-a shows the corrected height ($H_{cr}$) of group 1, where each track was plotted by regularly moving the origin 3 m upward. Other groups are plotted in the same manner. The increased slope at the beginning of the group 1 tracks is due to the geoid and possibly the Tsushima Warm Current, and the slope at the end also reflects the geoid and the EKWC signal. The bump at an along-track distance of approximately 260 km, with lateral scale of about 50 to 60
Figure 4-4. SEASAT altimeter tracks. The repeat tracks are labelled in Figure 4-1.
Figure 4-5. $H_2$ origin translated regularly upward for group 1 (a), group 2 (b), and group 3 (c), and no translating the origin from group 1 (d), group 2 (e) and group 3 (f).
Figure 4-5. Continued (numbers indicate track number)
km, is either due to the geoid or due to an eddy with a time scale long compared to one month. It is most likely geoid alongtrack variability due to bathymetric change (compare with the 1000 m isobath on Figure 4-1). The length of the track of group 2 (Figure 4-5-b) is short (300 km). Group 3 is shown in Figure 4-5-c. The Liman Cold Current signal is superimposed on the geoid near the end of the track. A bump of 100 km length appears on each track near 200 km, which also appears to reflect geoid variation. Figures 4-5-d, e, and f show the corrected height of group 1, 2, and 3, respectively without moving the origin. Relatively high variability is noticeable at the end of the track of group 1 (Figure 4-5-d). The average height along the tracks of each group is shown in Figure 4-6. The average height along the track varies from 32.0 m to 24.2 m for group 1, from 27.0 m to 29.0 m for group 2, and from 26.78 m to 36.3 m for group 3. The large ranges of the average height imply that the average height mainly reflects the geoid variation. Figure 4-7 shows the "geoid corrected" sea heights after removing the ensemble mean for each group. Because of the orbit error, each profile, in which the origin is moved 1 m upwards, is not equally spaced. The final error-free sea height ($H_f$) of each group after removing the tilt and bias of orbit assuming the orbit error to be linear is shown in Figure 4-8. As before, the origin of each profile is translated 1 m upward with time increasing upward (each profile 3 days apart) in Figure 4-8-a, b, and c (compare with Figures 4-8-d, e, and f, for which the origin of the each profile is not translated.)
Figure 4-6. Averaged $H_{cr}$ from group 1 (a), group 2 (b), and group 3 (c).
Figure 4-7. Geoid corrected sea height from group 1 (a), group 2 (b), and group 3 (c).
Figure 4-8. Error free sea height \( (H_f) \) origin translated regularly upward for group 1 (a), group 2 (b), and group 3 (c), and no translation of the origin for group 1 (d), group 2 (e), and group 3 (f).
Figure 4-8. Continued
Figure 4-9. The rms variability of sea heoght for group 1 (a), group 2 (b), and group 3 (c).
Figure 4-10. Wave number spectra estimates of tracks 1, 3, 5 and 7 of group 1 (a), group 2 (b), and group 3 (c).
The rms variability of the sea height (Figure 4-9) along the track of each group is calculated from $H_f$. The rms variability of group 1 is as large as 42 cm near Honshu, and as large as 25 cm at mid-track and 32 cm at the end of the track. The rms variability of group 2 varies from a few cm to 10 cm at the middle of the track and about 18 cm at both ends. The rms variability of group 3 is as low as 10 cm for the whole track. Figure 4-10 shows wavenumber spectra of selected $H_f$ spatial series for each group. At wavelengths of 100 km or longer, the power spectrum levels are one or two orders of magnitude larger than at the shorter wavelengths for each group. The wave number ($k$) dependencies of the power spectra are $k^{-2}$ to $k^{-5}$, $k^{-1}$, and $k^{-1}$ for group 1, 2, and 3, respectively. Fu (1983) observed that wave number spectra follow a $k^{-5}$ dependency in high energy areas, defined as areas of major currents, and a $k^{-1}$ dependency in the low energy areas remote from major currents.
4.4 Discussion

The average height (Figure 4-6) of each group of collinear tracks is supposed to be mainly the geoid signal. When I reconstruct the geoid along the track of each group by resampling from Watts and Leeds (1977), (Figure 4-11) it is not difficult to see that the average sea height (A in Figure 4-11) profile of each group does not reflect the geoid of Watts and Leeds (W in Figure 4-11). For comparison purposes, let us estimate the sea surface height difference across the TTCS front. Assuming geostrophic balance,

$$ u = \frac{g}{f} \frac{\partial h}{\partial y} \approx \frac{g}{f} \frac{\partial h}{\partial y} $$

where $f = 2 \Omega \sin \varphi$, the Coriolis parameter, $\Omega$ = angular velocity of the earth, and $\varphi$ is the latitude. The sea height difference across the front calculated using this equation is 47.8 cm if we assume $\Delta y = 100$ km, and $u = 50$ cm/sec. Except for the frontal regions associated with strong currents, the sea height variation is less than 50 cm, because velocities in other places are smaller than in the frontal region. It is therefore natural to ask what causes the large sea height differences which can not be explained by the Watts and Leeds' geoid. The answer will be either that the geoid along the track, as estimated by Watts and Leeds (1977), is in error or the orbit error is locally nonlinear. The latter is highly unlikely (Tapley et al, 1982). The geoid profile along the track of each group is reconstructed from altimeter data with orbit error removed by an analysis of crossover points (C in Figure 4-11). Except for a few
Figure 4-11. Geoid from Watts and Leeds, 1977 (W), estimated geoid from cross-over tracks (C), and the average sea height of the repeat tracks (A) of group 1 (a), group 2 (b), and group 3 (c).
points within each group, the average sea heights are not much different from the geoid profile obtained from the cross over tracks. It is inferred that the average sea heights are nothing but geoid signals. One must ask why the average sea height can be explained better by my geoid model than by the geoid constructed by Watts and Leeds. Possible explanations are 1) Watts and Leeds' measurements do not cover the northwest half in the Sea of Japan, and 2) the spatial resolution of their geoid measurements are not fine enough to detect local high variability of the geoid associated with complex bottom topography along the continental slope near the Korean coast. In fact, Watts and Leeds (1977) point out that the geoid model obtained based on earlier satellite altimetry data disagrees with their geoid measurements over the ocean. Note that the contours of mean sea surface height in Figure 4-3 mostly follow the isobaths along the continental slope near Honshu, Japan, while the isopleths in the geoid (Figure 4-1) of Watts and Leeds (1977) cross the isobaths. This is in contradiction to the experience that there is a very strong correlation between bathymetry and short wavelengths (40 - 400 km) features in ocean gravimetric geoids (White et al., 1983).

The rms variability of sea level along collinear tracks is relatively low except for group 1 (Figure 4-9). Considering that the estimated sea height is about 25-50 cm across the polar front and main currents, the low variability of group 3 suggests that the polar frontal region and associated currents were stable during the month of SEASAT repeat orbit data. Most mesoscale features with length scales of 100 km or longer have time scales of a month or longer (Wunsch,
SEASAT altimetry might not properly have detected the variability of these features because of its short life time. Also, because of the possible uncertainties in the geoid and the absence of correct orbit tracking information, the effective rms sea height resolution may be larger than 10 cm in this study. Real sea height variation may be masked by noise. Variability within group 1 (30-42 cm) is much larger than within group 2 (about 10 cm), although the two lines cross. One possible explanation is that the track of group 1 follows isobaths, while the track of group 2 crosses them (Figure 4-1). When the satellite passes over the "same" track, it may deviate from the previous actual path by as much as 3 km (Bernstein, et al., 1982). Variability due to the change of bottom topographic gradient, associated with the lack of collinearity of the repeat track, is maximum if a satellite follows isobaths, and is minimum if it crosses the isobaths. The high variability of group 1 is possibly attributable to the bottom topographic gradient change associated with the satellite's crosstrack motion. It is certain that a precise geoid with a finer spatial resolution than presently available will be essential for studying mesoscale dynamics in the low energy Sea of Japan.

I use the mean sea surface derived from the crossover point analysis to estimate the spatial distribution of rms variability of the residual orbit error plus sea height. The Sea of Japan is divided into 6 sections as shown in Figure 4-3. Most of the rms variability is due to section 1 (190 cm) and section 2 (176 cm), while the variability of the other sections falls below the rms variability of the overall area (169 cm). This coincides with the high rms
variability of group 1 residual sea level based on repeat tracks. Although time differences for cross over points (about 2 months) are longer than that of collinear tracks (less than 1 month), the high rms variability in sections 1 and 2 is still likely due to complexity of topography in those sections. Note, also, that the rms variability along the continental slope in the northwest sections is higher than that in the southeastern sections (see table 4-1).

<table>
<thead>
<tr>
<th>sections</th>
<th>2</th>
<th>4</th>
<th>6</th>
</tr>
</thead>
<tbody>
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<td>rms variability</td>
<td>176</td>
<td>145</td>
<td>169</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>sections</th>
<th>1</th>
<th>3</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td>rms variability</td>
<td>190</td>
<td>60</td>
<td>100</td>
</tr>
</tbody>
</table>

Table 4-1. The rms variability in the sections in cm.
4.5 Summary

The rms variability of residual (time varying) sea level for group 1 is as high as 42 cm near the main currents flowing into the Sea of Japan through the Korea Strait, and that for group 2 and group 3 is as low as 20 cm and 10 cm, respectively. The Polar Frontal region and associated currents appear stable on time scales shorter than a month. The high variability in group 1 could be attributable to topographic effect associated with the lack of exact collinearity of the repeat tracks.

The residual orbit plus sea height uncertainty in the Sea of Japan is 169 cm rms, based on the cross over differences, and occurs, mostly, near the continents in the northwest section of the Sea of Japan. This rms value is higher than the 1 m proposed by Cheney and Marsh (1981a). Nevertheless, the geoid estimate (Figure 4-3) obtained by the cross arc method, explain the repeat track data better than that of Watts and Leeds (1977).

Wavenumber spectra show that at wavelengths of 100 km or longer, the spectral level is higher by one (group 2 and 3) or two (group 1) orders of magnitude than at shorter wave lengths. How much of this variability is due to currents, as opposed to the geoid, is not presently determinable.

The precise orbit error and detailed geoid distribution are crucial in studying the mesoscale dynamics in the Sea of Japan by satellite altimetry, because of the low energy level of currents and the high variability of the bottom topography there. In spite of the low precision of GEOS-3 (about 30-40 cm; Douglas and Cheney, 1981),
it may be very useful in studying mesoscale ocean dynamics in the Sea of Japan because it has been operational over a longer period of time (3.5 years) than SEASAT. This data should be further analyzed.
CHAPTER 5. SUMMARY AND DISCUSSIONS.

In order to investigate the spatial and temporal variability of the Polar Front in the Sea of Japan, I analyze both the conventional hydrographic data from the files of the NODC and SST data and sea height data derived from satellites.

I identify important features of the Polar Front in the Sea of Japan. These include the spatial and temporal scales of the Polar Front, frontal structure (mostly horizontal), and eddies associated with the Polar Front.

The annual mean position of the Polar Front extends from 38.5 °N near the Korean Peninsula to 41.5 °N near the Tsugaru Strait. The basic surface shape of the Polar Front inferred from this annual mean frontal position is sinuous, tilted toward the north. The rms variability of the annual mean frontal position is about 100 to 120 km in the western and central region, while the rms variability in the eastern part of the sea is about 300 km (Figure 3-11). Temporal variability on the annual time scale is reflected in the seasonal mean frontal positions of the Polar Front. The mean frontal positions in both spring and fall are nearly the same as the annual mean position. The mean position in summer is 50 km north of the annual mean position except at the eastern end near the Tsugaru Strait where it is about 200 km north. The winter mean position is south of the annual mean up to 50 km. This seasonal variability is also apparent in the cross-spectrum analyses of the frontal position, which suggest that the dynamics of the Polar Front on the eastern end, at a month time scale are different from the dynamics on the western end. This is in
contradiction to the conventional wisdom which states that both ends of the Polar Front are unstable, while the center is relatively stable (Istoshin, 1960).

The Polar Front, being the boundary between warm and cold water sections in the Sea of Japan, is well defined by the thermal fields. A strong local increase in the horizontal temperature gradient defines the Polar Frontal zone. The width of the Polar Frontal zone estimated from the hydrographic data (100 km) is smaller than that estimated from the AVHRR data (about 200 km). This may reflect both the coarse sampling grid and non-synoptic nature of the hydrographic data. Wave motion of the Polar Front is characterized by small amplitude (50 km) and long wavelength (700 km) in winter, and large amplitude (100 km) and short wavelength (400 km) in summer. This large amplitude and short wavelength of the summer frontal motion suggests that the frontal regions may be unstable. This instability can be due to the increase of baroclinicity as a result of the low salinity water inflowing through the Korea Strait and positive heat flux during summer.

Eddies in the Polar Frontal zone vary in size and intensity, and seem not to be migratory. It is, in general, difficult to determine the size of an eddy using conventional hydrographic data, especially when the eddy is small. I am not able to measure the vertical and horizontal structure of eddies from the NODC hydrographic data, except to determine that they do not penetrate below the permanent thermocline as shown in Figure 2-12-b and Figure 2-12-e. The lifespan of eddies in the Polar Frontal zone is estimated to be about a month or so. The transport of momentum, heat, and salt into the cold water
north of the Polar Front is therefore by entrainment rather than by migration of eddies. The usual size of individual eddies discerned from AVHRR imagery is generally 50 to 100 km in diameter, either circular or elliptic, though size varies from place to place. A two-dimensional Fast Fourier Transform spectrum shows that the eddy field of the frontal region can be described as waves, with wavelength from 45 to 75 km in the principal directions (north-south and east-west) and less energetic waves of wavelength 25 to 30 km in ± 30 to 45° with respect to the positive x-axis (to the east).

The Polar Frontal zone is a region of locally high geostrophic velocity shear. The thermocline topography, estimated by the 7 °C isotherm depth, is responsible for this local maximum. The sea height change associated with the Polar Front must be greater than at any other place in the Sea of Japan as long as the local dynamics are geostrophic.

With the intention of detecting this sea height change, SEASAT altimeter data was analyzed. I failed to detect the anticipated signal. This may be either because the Polar Frontal zone is invariant on time scales as long as three weeks (length of the SEASAT repeat track data), or the precision of the SEASAT altimeter (about 10 cm rms) is not good enough to detect the real variability of the Polar Front. The lower precision GEOS-3 altimeter data (about 30 to 40 cm; Douglas and Cheney, 1981) may provide useful information concerning the Polar Front dynamics because of its long operational time span (3.5 years), if one utilizes the geoid estimated from SEASAT data.

Time-series analyses of the polar Frontal position at 20 km
intervals along an axis from the Korea Strait to the Tsugaru Strait show that time scales longer than seasonal are dominant, and a one month time scale is of secondary importance. The seasonal signal is clear in the hydrographic data. To determine possible forcing(s) responsible for the dynamics of the Polar Front on the monthly scale, cross spectra of the frontal position with wind and sea level were calculated. Neither the weekly mean wind stress at Kangreung and Ulleungdo Island nor the weekly mean adjusted sea level at Pusan and Pohang are coherent with the frontal position. Unadjusted sea level at Pusan, however, shows a significant coherence at one station, 34. This suggests that the northward movement of the frontal position is a response to increased flow through the Korea Strait which is reflected in increased water level at Pusan.
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VITA

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I was born on April 26, 1951 in Cheongcheon, Choongbook-do, Korea. Since my home town is in the central part of the Korean Peninsula, I always dreamed of the sea when I was young. I attended Seoul National University and received B.S. in oceanography in February, 1984. I didn't realize that physical oceanography would be fun to study until I joined the Navy of the Republic of Korea as an officer. I decided, then, to continue my studies in physical oceanography. I received a Master of Science degree in oceanography from Seoul National University in 1980, and came to study for the doctorate in the Marine Sciences department of Louisiana State University in 1981. I will receive the degree of Doctor of Philosophy at the summer commencement, 1986 and begin working at the Chinhae Research Laboratory as a senior researcher shortly thereafter.
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Major Field: Marine Sciences

Title of Dissertation: THE SPATIAL AND TEMPORAL VARIABILITY OF THE POLAR FRONT IN THE SEA OF JAPAN

Approved:

Co-Major Professor and Chairman

Dean of the Graduate School

EXAMINING COMMITTEE:

Co-Major Professor and Chairman

Date of Examination:
14 July 1986