The association of frontal formation, development, and movement with large scale baroclinity is well documented. This investigation deals with the mean large-scale baroclinic zones near the Asian East Coast for the last two weeks of February 1975. Two deep, large-scale baroclinic zones are found to be situated along the axis of the zones of high frontal frequencies depicted by previous investigations of the region's frontal climatology. The "southern baroclinic zone" lies along the path of the warm Kuroshio ocean current and beneath the climatological location of the upper level jet stream.

A frontogenesis equation is developed to assess the role of mean fields and perturbation fields upon maintaining the mean baroclinity. Analyses contained within demonstrate the various effects of those fields. It is shown that to the north and to the south of the southern baroclinic zone the mean diabatic heating and mean
horizontal advection (of mean potential temperature) are the dominant terms in the equation, but that they tend to cancel each other. Within the southern baroclinic zone the frontogenetic effect of the mean diabatic heating term is negligible as is the effect of the perturbation vertical advection term. While the frontogenetic effect of the mean horizontal advection term is smaller within the zone than outside, it is important for the maintenance of the baroclinity there. The effects of mean vertical advection and perturbation horizontal advection were the other important terms within the zone and oppose the effects of the mean horizontal advection term.
Mean Fronts and Frontogenesis
Near the Asian East Coast

by

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The earth-sun orientation, as well as the earth's axial rotation, is such that low latitudes receive more solar energy than high latitudes. The outgoing terrestrial radiation does not balance the incoming solar radiation everywhere on the globe and thus there is a high-latitude deficit and a low-latitude excess of energy accumulation. A meridional temperature gradient exists due to this imbalance and atmospheric motions are induced to redistribute the energy. As a result of these motions, the time-averaged temperature distribution varies little from year to year.

It is interesting to note that the atmosphere chooses to confine much of the meridional temperature gradient to relatively narrow zones called "fronts." The interrelationships of fronts with cyclones, jet streams, and cloud and precipitation systems continues to be an interestingly complex topic which investigators are continually trying to better understand.

While the literature is full of studies attempting to explain the features of fronts and the processes causing their formation and desolution, few studies have been completed which pinpoint the areas on the globe in which fronts tend to be located and the areas where they form and dissolve. The purposes of this presentation are:
1. Review the previous concepts of fronts and frontogenesis.

2. Study the mean frontogenetic processes in an active region.

3. Study quantitatively frontal and frontogenesis frequencies.

4. Relate these results to those of other investigators.
II. HISTORICAL REVIEW OF FRONTAL AND FRONTGENESIS CONCEPTS

In order to study quantitatively the climatology of fronts and frontogenesis it is necessary to know what constitutes a front and what processes lead to their development and movement. This problem has been the subject of many investigations even prior to this century. The first major breakthrough on this theme came when J. Bjerknes and his collaborators at the Bergen School of Meteorology presented the Norwegian Cyclone Model and formulated the Polar Front theory.

Bjerknes (1918) was the first to accurately depict the structure of the surface cyclone (see Figure 1). Having been instrumental in setting up a dense network of observing stations, he discovered the two zones of confluence associated with the warm and cold fronts and identified the warm sector air as being of tropical origin. He related the areas of precipitation to the vertical motion associated with the warm air sliding up over cold air at the warm front and being forced up by cold air at the cold front. By pointing out the movement, development, and energy transformation of cyclones, he dispelled the belief that cyclones were cold core phenomena.

Further developments, relating to the cyclone structure were imminent and followed in relatively rapid succession. In 1919 Bergeron discovered the occlusion process. Bjerknes and Solberg (1922) presented their model of the life cycle of cyclones; from the formation of a wave at the boundary of two opposite flowing currents
FIGURE 1. Idealized cyclone, from J. Bjerknes and Solberg (1921). In middle diagram, dash-dotted arrow shows direction of motion of cyclone; other arrows are streamlines of air flow at earth’s surface. Top and bottom diagrams show cloud systems and air motions in vertical sections along direction of cyclone movement north of its center and across the warm sector south of its center.
of different temperatures to the stage of the dissipating vortex. They illustrated the two possible occluded fronts which form when the cold air behind the cold front overtakes the cold air ahead of the warm front. They also discussed the formation of secondary cyclones south of the primary cyclone at the point of occlusion.

With the knowledge of the formation and evolution of surface cyclones, together with their associated fronts, theory and observations about the regions of cyclogenesis and frontogenesis were also investigated. Bjerknes in his paper of (1918) states "Cyclones are most frequent and develop to greatest intensity in the zones and during the seasons of great horizontal temperature gradients, the surface of discontinuity preeminently evolving under such conditions." Bjerknes and Solberg (1922), along with their model of the life cycle of the wave cyclone, also presented their famous Polar Front Theory. "The polar front is generally a wavy line in continual motion through all latitudes of the temperate zone, bordering large tongues of polar and tropical air."

The polar front is thus the zone of horizontal temperature gradient on which Bjerknes (1918) stated cyclones form. In describing the polar front Bjerknes and Solberg (1922) discussed the formation of cyclone families. They showed that cyclones forming on the polar front transport heat northward. They envisioned an earth devoid of landmass in which there are continually a longitudinal series of cyclone families each member of which follow a spiral path to the pole. Each cyclone family is separated by an anticyclone associated
with tongues containing air of tropical origin. Described in this manner, fronts and the cyclones with which they are associated become an integral part of the general circulation. For a recent review of the role of the cyclone in the general circulation, see Newton (1970).

The development of frontal theory continued through kinematic studies showing that a scalar field (such as temperature) can be concentrated along the axis of dilation by a motion field possessing deformation. Bergeron (1928) showed that a linear field of potential temperature can be concentrated in this manner when acted on by a stationary field of deformation. Petterssen (1936) illustrated many examples of frontogenesis possible by this process while examining non-linear conservative scalar fields with linear velocity fields. Defining frontogenesis as the total time derivative of the gradient of a conservative scalar, he showed that only the deformative and divergent parts of linear velocity fields contribute to frontogenesis and that frontogenesis will occur only when the tangents to the isopleths of the scalar lie within about 45 degrees of the axis of dilation.

Palmén and Newton (1969) show that there must be cyclonic wind shear across fronts. Petterssen and Austin (1942) emphasized this important aspect of frontal dynamics. They found that to maintain the cyclonic vorticity at a non-dissipating front, horizontal convergence is required.

Major advances on the structure of atmospheric circulation systems began to occur in the late 1940's. Concentrated efforts
brought about by World War II resulted in rapid development of many disciplines and the field of meteorology kept pace. The increased use of airplanes brought about not only a need for better weather forecasts, but also a need for knowledge of the flow patterns and other essential air mass characteristics above the lowest layers of the atmosphere. This and the development of more efficient methods of communication and data processing led to the establishment of a global network of sounding stations which twice daily record variables throughout deep layers of the atmosphere. The increasing availability of observational material also stimulated the more rigorous application of thermodynamics and fluid mechanics to meteorological problems.

Prior to World War II quantitative investigations of fronts and frontogenesis were mainly confined to near the surface for practically all observations were made at that level. With the greatly increased amount of upper air data came the discoveries of new atmospheric phenomena demanding study and explanation. Of these the most noteworthy is the jet stream, a narrow band of high-speed flow (frequently greater than 50 meters per second). Such winds were encountered by bombers flying missions over Japan near the close of World War II. While evidence of the jet stream had been inferred on pressure maps prior to its discovery, the data from which the maps were based were believed to be of too low a quality to verify its existence (Reiter, 1963).

Palmén (1948) and Palmén and Newton (1948) related the jet stream to the polar front. They illustrate the three-dimensional
structure of the polar front on cross-sections and the 500 mb surface (see Figure 2). By analyzing meridional cross-sections along 80°W they described the three-dimensional features of the polar front. The front is shown as a sloping stable layer throughout the troposphere separating polar and tropical air. The frontal layer is characterized by a large magnitude of horizontal temperature gradient which far exceeds that found in the air masses on either side. The jet stream is pictured as a narrow band of high zonal geostrophic wind located in the warm air approximately above the intersection of the polar front and the 500 mb surface. The tropopause is analyzed higher in the tropical air than in the polar air such that a reverse horizontal temperature gradient exists in the lower stratosphere. Thus the height of the maximum wind speed is located below the tropopause in accord with thermal wind considerations. They point out that the lack of a well-defined tropopause above the front could be due to bad resolution due to station spacing or a multiple structure of the tropopause in that region.

With the development of reasonably good three-dimensional structural models of the atmosphere in the region of the polar front, research was concentrated on the explanation of the time evolution of cyclones and the role that these "eddies" play in the general circulation. The importance of the dynamics of the upper-level waves in the westerlies to the development of cyclones as explained and modeled by many different researchers was recognized. However, these topics in themselves are not within the scope of this work.
FIGURE 2. From Palmén and Newton (1948). Mean temperature and zonal component of geostrophic wind, computed from 12 cases in December 1946. The cross section lies along the meridian 80°W. Heavy lines indicate mean positions of frontal boundaries. Thin dashed lines are isotherms (degrees Centigrade, slanting numbers) and solid lines are isolines of westerly component of wind (meters per second, upright numbers). Means were computed with respect to the polar front.
The remainder of this study deals with some of the investigations which have a direct bearing on the location of fronts and frontogenesis in the atmosphere.

The classic paper by Miller (1948) on the subject of fronts and frontogenesis in the atmosphere serves as a basis for many recent studies on the subject. The important point of this paper is that it is the first to assess the role of vertical motions on frontogenesis. By using the definition of Pettersen (1936) Miller shows that the frontogenetic effect of the terms involving the vertical velocity are the same order of magnitude as those involving the horizontal deformation. At this point in time, observations were sufficient in both quality and quantity throughout the troposphere that reasonable values of vertical velocities could be calculated. This allowed the study of the three-dimensional aspects of frontogenesis.

Miller's equation reduces in component form to

\[ F_z = \frac{\partial}{\partial z} \frac{\partial S}{\partial t} - \left[ \frac{\partial S}{\partial x} \frac{\partial u}{\partial z} + \frac{\partial S}{\partial y} \frac{\partial v}{\partial z} + \frac{\partial S}{\partial z} \frac{\partial w}{\partial z} \right] \]

\[ F_y = \frac{\partial}{\partial y} \frac{\partial S}{\partial t} - \left[ \frac{\partial S}{\partial x} \frac{\partial u}{\partial y} + \frac{\partial S}{\partial y} \frac{\partial v}{\partial y} + \frac{\partial S}{\partial z} \frac{\partial w}{\partial y} \right] \]

with a similar component for the \( x \)-direction. The terms in the equation can be interpreted as follows:

\( A \& E \) represent superimposed non-conservative influences; such as gradients of diabatic heating or cooling if \( \theta \) is replaced for \( S \).
B & F represent effects due to shear of the flow across the front.
D & G represent effects due to vertical and lateral confluence fields.
C & H represent effects due to variations, of the horizontal and vertical winds normal to the front, in the vertical and horizontal directions respectively.

These processes are illustrated schematically in Figure 3.

Reed and Sanders (1953) were among the first to apply Miller's equation and did so to a case of intense upper-level frontogenesis. They treated both the thermal and vorticity aspects of frontogenesis. Reed (1955) studied another case of intense upper-level frontogenesis by considering the conservation of potential vorticity on isentropic surfaces. A number of interesting observations are brought out by these studies.

For example, the classical picture of the polar front as a surface or zone of separation of air of polar and of tropical origin is not valid in the case of upper tropospheric fronts. Instead, at least some upper-level fronts (in particular, the intense kata fronts) form when stratospheric air intrudes into the troposphere.

In addition, the vertical velocity is seen to be a dominant factor leading to the frontogenesis. A thermally indirect (kinetic energy consuming) circulation results in the sinking of warm air and the lifting of cold air as frontogenesis proceeds. The surfaces of potential temperature are tilted by the effects of the vertical motion (see Figure 3). Since the tilting increases the horizontal gradient of potential temperature, there must be a compensating
Illustration of frontogenetic effect of terms D and G of Miller's equation.

Illustration of frontogenetic effect of terms C and H of Miller's equation.

Illustration of frontogenetic effect of terms B and F of Miller's equation.

FIGURE 3. Illustration of terms of Miller's (1948) frontogenesis equation.
increase in the isobaric vorticity to satisfy the conservation of potential vorticity.  \(^{(1)}\)

Such developments are consistent with the concept of tropopause folding (see Danielsen, 1964 and Hoskins, 1971). This may be the explanation of why the tropopause on the cross-section of Palmén (1948) is indistinct near the polar front. Diagnostic studies carried out by Reed and Danielsen (1959), Danielsen (1964), Bosart (1970), and Shapiro (1976) show that quasi-conservative properties representative of stratospheric air are found within upper level frontal zones in support of the theory of the tropopause folding.

Newton (1954) investigated the time evolution of the vertical structure of the atmosphere associated with a developing surface cyclone below an upper-level trough in which the strong temperature gradient is wrapped around the west side. Miller's equation again formed the basis for the study. It is shown that in this case, in

\[ (\zeta \theta + f) \frac{\partial \theta}{\partial p} = p = (\zeta_p + f) \frac{\partial \theta}{\partial p} + \hat{\kappa} \cdot \frac{\partial h}{\partial p} \times \nabla_p \theta \]

Some investigators define potential vorticity by term two which is the component of potential vorticity on an isobaric surface. However, Reed shows that in the vicinity of fronts, term three should not be neglected since both the vertical wind shear and horizontal temperature gradient are large there.

\(^1\)Reed's definition for potential vorticity is
the beginning, the upper- and lower-level fronts differ in structure. It is only when the high-level front moves eastward over the lower-level front that the two combined into a single pronounced frontal layer extending from the surface to the tropopause level.

The lower-level front was formed by the horizontal advective processes described by Petterssen (1936) and Petterssen and Austin (1942). The middle and upper frontal layers were formed not only by the tilting due to the vertical component of the wind, but also by lateral confluence of the flow normal to the front and differential horizontal advections across the front; each term being of comparable magnitude.

As the complexity of frontal structures and processes was revealed by diagnostic studies, scientists began to use analytical and numerical solutions to simulate some of the major features observed in fronts. Two approaches are commonly followed. The first involves the simulation of frontogenesis as part of large scale cyclogenesis. A three-dimensional model is usually used in this approach and the velocity field contains stretching and shearing deformation as component parts. The second method requires only a two-dimensional model and usually concentrates on showing the frontogenetic effect of one of the above-mentioned components of the wind field. Both approaches start with initial fields which are believed to represent atmospheric conditions prior to frontogenesis. The model produces fields at subsequent time intervals showing the development of various aspects of fronts.
Stone (1966), using a two-dimensional model, and Williams and Plotkin (1968) using analytical techniques, both use the quasi-geostrophic system of equations to study the frontogenetic effects of the stretching deformation component of a wind field. They demonstrate that a discontinuity in potential temperature can be produced by the horizontal deformation field. However, their results were unrealistic in several respects. Their solutions produced a discontinuity only as time approached infinity, the slope of the front was vertical, the maximum tangential velocities occur in the center of the frontal zone and the maximum speed in this jet tends to infinity as time increases. Williams (1972), Gidel (1977), Hoskins (1971), and Hoskins and Bretherton (1972) point out that it is the neglect of the divergent part of the wind in the horizontal advection terms by the quasi-geostrophic approximation which account for the unrealistic solutions.

Hoskins (1971) obtains analytical solutions to frontogenesis using the full set of primitive equations for an inviscid, adiabatic compressible fluid over a plane surface. The structural solution of his models closely resemble cold front structures previously observed by the diagnostic studies.

Specifically, his two-fluid deformation model produces upper tropospheric fronts which contain many of the features observed by Reed and Sanders (1953) and Reed (1955). Most striking is that the tropopause is shown to fold with stratospheric air intruding into the troposphere not, however, down to 700 mb. Also, a realistic jet...
stream pattern developed near the tropopause though the wind speeds are too low.

Williams (1972) studied frontogenesis which is forced by a non-divergent horizontal wind field which contains stretching deformation. He numerically solved the linear and nonlinear hydrostatic primitive equations and compared the two solutions and the solutions of the quasi-geostrophic models of Stone (1966) and Williams and Plotkin (1968). His results show that the linear solution resembles that of the quasi-geostrophic solution in that there is no tilt to the frontal zone and it produces discontinuities only at the surface and only as time approaches infinity.

The non-linear solution gives much more realistic results. There is rapid development of a sloping frontal zone which gives rise to cyclonic vorticity in the zone. Also there is evidence that the discontinuity will form in a finite period of time.

Hoskins and Bretherton (1972) also discuss the differences between quasi-geostrophic and non-geostrophic frontogenesis. They conclude that large-scale geostrophic processes intensify the horizontal temperature gradients but that it is the smaller scale ageostrophic motions embedded in the baroclinic flow which lead to the rapid formation of a near discontinuity. It is when the relative vorticity predicted by the quasi-geostrophic theory approaches the value of the Coriolis parameter that small-scale ageostrophic circulations become dominant and a near discontinuity forms abruptly at a rigid boundary such as the earth's surface.
More recently, high resolution three-dimensional numerical studies by Mudrick (1974) and Price (1977) have extended the earlier two-dimensional results and demonstrated the complex physical nature of the vertical circulation patterns and frontogenesis mechanisms inherent in front-jet systems. As in the two-dimensional studies, the models are restricted to using dry, hydrostatic inviscid adiabatic motions. The solutions exhibit the development of frontal zones with features consistent with observations. The surface fronts are shown to arise from stretching and shearing deformation contributing to the increased horizontal temperature gradient. The vertical circulation pattern which accompanies this frontogenesis is thermally direct.

Mid-tropospheric frontal zones are found to be characterized by frontogenesis upstream of the baroclinity maximum due to horizontal deformation effects. The vertical circulation in this region is also thermally direct. Downstream, the frontogenesis is due to the tilting mechanism (see Figure 3) and the associated vertical circulation is found to be thermally indirect.

Bergeron (1959) points out that improved weather forecasts are made when meteorologists use more complete diagnostic models of atmospheric weather systems. Concurrently, more complete models have come about only with an increase of better observations of the four-dimensional atmospheric structure.

This is precisely what makes the work by the Norwegian School of Meteorology so important. Their model of the life cycle of cyclones
and the polar front theory combined the previously fragmented pieces of observed facts together with their newly observed ones to formulate the first relatively complete usable diagnostic model. By combining all the synoptic observations on a surface synoptic weather map, forecasters could more accurately pinpoint the future areas of weather events by referring to the Norwegian Cyclone Model which demonstrates the evolution of surface weather systems.

Similarly, World War II brought with it a need for even better forecasts, which concurrently saw an increase in the quality and number of atmospheric observations. The more complete structure of the atmosphere was revealed and the correlation of events occurring in the upper levels of the troposphere with those at the surface began to be documented. However, no existing model incorporated the features of the four-dimensional structure of the atmosphere with the inclusion of predictable weather events. The work of Palmén (1948), Palmén and Newton (1948) and others relating the jet stream to the polar front theory can be thought of as the first attempt at formulating a more complete diagnostic model.

The more recent diagnostic, analytical and numerical investigations have shown that atmospheric circulation systems, together with their component parts, are probably too diverse in nature to be explained by just one simple model for use by synoptic forecasters. Today, the diagnostic model used is fairly complex in which emphasis has shifted to understanding the dynamics of evolving atmospheric structures. The dynamics contain the mechanisms and processes necessary for atmospheric prediction.
These points are very important to this study for it has historically been the synoptic forecasters who day in and day out place the frontal positions on the synoptic weather maps. Though there exist criteria for the actual placement of the fronts on the maps, it is usually found that the positions are placed according to the forecasters understanding and knowledge of what constitutes a front and what processes act to create and destroy fronts; in other words, they are placed on the map according to the diagnostic model being used by the forecaster.

All previous climatological investigations on fronts and frontogenesis have been based on counting the number of fronts in a given area on the daily synoptic weather maps produced by the forecasters and collected and published by different national agencies.
III. PREVIOUS INVESTIGATIONS OF MEAN FRONTS AND FRONTOGENESIS

As can be seen in the previous section, fronts and frontogenesis have been the subjects of many investigators throughout the development of meteorology. While much work has been completed and can be found in the literature on the structure of fronts and the processes relating to frontogenesis, comparatively little can be found on the statistical distribution of fronts and frontogenesis (or the closely related problem of cyclogenesis) over the globe.

Petterssen (1941) was the first to put average frontal zones on a hemispheric map. He infers the mean positions of the principal frontal zones in the Northern Hemisphere from his mean surface wind charts. An important point which can be seen in his work is that fronts form in the regions of confluence marked by the air currents flowing in relation to the surface low- and high-pressure systems. He points out that the differential longitudinal heating due to the distribution of land mass in the Northern Hemisphere has a pronounced effect on the location of fronts. Of particular interest to this treatise is the position of the Atlantic Polar Front and the Pacific Polar Front in winter which lie to the east of the major continents.

Along these lines, Miller (1946) and Miller and Mantis (1946) studied average features of cyclones and cyclogenesis in the regions of the east coasts of North America and Asia during the winter months. Miller and Mantis in their study of the Far East point out
that there is a marked relationship between cyclone frequency and topography. Maximum frequencies occur over water areas that are partly surrounded by land and to a lesser extent over low lying land areas that are partly surrounded by higher terrain. Conversely, minima occur over island chains, the Korean Peninsula and mountainous areas. Another interesting point alluded to in this paper is that there is a tendency for the maximum frequency to follow the orientation of the sea-surface isotherms.

Schumann and van Rooy (1951) appear to be the first to apply a systematic statistical examination of the frequency of fronts in the Northern Hemisphere. (Unfortunately, due to lack of data corresponding to the deteriorated international relations, the sector which includes most of Asia was excluded from their study.) Fronts appearing on ten years of surface analyses were counted for each five-degree latitude-longitude section. By allowing for the fact that the area of the grids decrease with the cosine of the latitude, the investigators were able to depict percentage frequencies of fronts per unit area over most of the Northern Hemisphere. In winter, the line of maximum frequency corresponds closely to the polar frontal zones inferred by Petterssen (1941).

They also calculated average frequencies of fronts for each five-degree latitude belt and compared them to the mean temperature gradient within the layer between the 1000 and 500 mb levels. Upon studying the results, they document the observation suggested first by Bjerknes (1918) that "the greatest frequency of fronts is
undoubtedly to be sought in the region where the greatest tempera-
ture contrast prevails."

Klein (1956) studied the frequency of occurrence and formation
of migratory cyclones and anticyclones on daily weather maps at sea
level and related them to the 700 mb mean zonal wind speed. The
frequencies of the high- and low-pressure centers were compiled in
equal area boxes of length five degrees latitude, and average width
of five degrees longitude. No center was counted twice. The fre-
quencies were then averaged around 360 degrees of longitude.

He finds that the axis of high cyclone occurrence coincides
closely with the axis of maximum cyclonic geostrophic vorticity
computed from the 700 mb mean zonal wind speeds. By noting the first
day a cyclone appeared on the weather maps, he computed frequencies
of cyclogenesis in the same manner as cyclones. He confirms the
results of previous investigators, that new low-pressure centers
form in the vicinity of the jet stream where meridional temperature
gradients are large.

Reed (1960) is critical of inferring the locations of mean
fronts by distributions of other mean atmospheric variables. He
states,

in dealing with non-linear processes such as fronto-
genesis, it is generally not permissible to infer mean
behavior from component means . . . if the use of mean
charts is justifiable, it is pertinent to ask why the
average positions cannot be estimated more simply and
accurately by use of the wind and temperature criteria
ordinarily employed in locating fronts rather than by the
indirect method involving frontogenesis.
He uses a method similar to Schumann and van Rooy (1951) to obtain frontal frequencies in the Northern Hemisphere. Using the Daily Series Synoptic Weather Maps for the winter and summer seasons of 1952 through 1957 he calculated frequencies of occurrence of fronts in areas of 400,000 square kilometers. The axis of elongated regions of high frequencies refer to the principal frontal zones (see Figure 4).

In winter, he depicts three major frontal zones known as the Pacific, Atlantic and Eurasian Polar Fronts. The oceanic fronts are stronger and situated further to the south at their western extremities than at their eastern ones. The Eurasian Polar Front also weakens to the east but tends to remain at nearly the same latitude throughout its extent.

Several features appear to affect the distribution of frontal frequencies; most notably are mountain barriers, land-sea boundaries, and ocean currents. A very noticeable feature is the tendency for high frontal frequencies to be located in the western portions of the oceans. Reed points out the remarkable correspondence between regions of high frontal frequencies and zones of maximum sea-surface temperature concentration. He notes that "it is convenient to speak of ocean currents (the Gulf Stream-Labrador Current systems in the Atlantic and the Kuroshio-Oyashio systems in the Pacific) as controlling factors in the asymmetries of the frontal distributions."

Yoshino (1971) provides another study of the mean positions of fronts. He restricts his region of interest to include only East
FIGURE 4. From Reed (1960). Percentage frequency of fronts in squares of 400,000 km², December—February. Heavy lines denote axes of maximum frequency (principal frontal zones).
Asia and the Northwest Pacific. He obtained his data from the daily weather maps published by the Japan Meteorological Agency for the years 1959-1963. In this investigation, as in previous ones, fronts were counted, this time in areas of 2 degrees latitude by 2 degrees longitude. The frequency of occurrence of fronts within the areas were computed and analyzed for each month.

Yoshino finds the axis of the Pacific polar front to be about 10 degrees north of where it appears on the analysis of previous investigations. He attributes this difference to the different data used. (Earlier studies mainly used the Daily Series Synoptic Weather Maps published by the U.S. Weather Bureau.) By restricting the region of study and using a smaller grid size to compute frontal frequencies, he is able to obtain a finer resolution in defining the areas of maximum frontal activity than other studies (see Figure 5).

Thus he points out areas of "subfronts" as those areas with slightly less frontal activity than the principal Pacific polar front. A striking similarity exists between the location of these zones and the regions of high cyclone activity presented by Miller and Mantis (1946). For instance, minor frequency maxima can be seen in the Sea of Japan and low elevation areas of the Asian mainland.

Another interesting point of this study not present in others is that Yoshino appears to be the first to climatologically analyze fronts above the surface, and does so for the 850 mb level. In general, the maximum frequency of occurrence of fronts at this level are directly above those found at the surface.
FIGURE 5. From Yoshino (1971). Seasonal changes in the Pacific polar frontal zone.
Laboratory experiments of Fultz (1952), Faller (1955), and others have shown that fronts naturally form in rotating fluids which are differentially heated. On the earth, the distribution and contour of land masses especially in the Northern Hemisphere greatly affects the location of mean frontal zones in the atmosphere and oceans.

The evidence presented in this section indicates that the zones of atmospheric frontal activity near the surface are greatly affected by, among other things, mountain ranges, sea-surface isotherms, differential heating rates of land and water, the position of the jet stream and regions of large-scale confluence. Today, it is apparent that none of these factors is independent of the others.

On winter mean charts it is seen that the mean jet stream is situated in the mean upper-level trough, downstream of the major mountain ranges, at the boundaries of the continents which are bordered by warm ocean currents. The warm ocean currents are constrained to flow parallel to the continents. The different rates of heating of the land and water causes in winter a mean low-level high-pressure center to be located over the land masses. In addition, a mean low-level low-pressure center is found at high latitudes over the oceans between the continents. The flow about this pressure system greatly affects the synoptic situations off the east coasts.

While previous investigators have been instrumental in identifying the regions of greatest surface frontal activity, there appear to be some discrepancies among them. The differences appear to be
associated with the varying methods of surface frontal analysis by
different national agencies in preparing the surface maps used in
those studies.

Furthermore, I have found no previous attempt of statistically
identifying the average position of frontal zones in the upper
levels of the troposphere. The diagnostic and theoretical investi-
gations of frontal structure and evolution suggests that in many
cases, upper- and lower-level fronts differ in structure and that
different processes are involved in their formation and development.
It thus seems appropriate to investigate possible asymmetries in the
analyses of mean frontal positions at different levels.

It is quite apparent that this subject requires and deserves
much more attention. It is the intention of the author to illustrate
a method to be used in more thoroughly investigating these topics.
Throughout the history of meteorology, various methods have been used for defining and analyzing fronts. While there are many identifiable features associated with frontal phenomena, this thesis follows the reasoning of Petterssen (1936) in that the horizontal gradient of potential temperature is used to define frontal intensity. (At this point, it may be helpful to refer to Table 1 for a brief explanation of the symbols and operators used in the remainder of this thesis.) Specifically, Palmén and Newton (1948), Hoskins and Bretherton (1972) and others have shown that fronts (identified by their sharp temperature gradient) form in regions of large-scale baroclinity; Saucier (1955) demonstrates that the baroclinity (defined as $\nabla \alpha \times \nabla p$) can be redefined for a hydrostatic atmosphere and on an isobaric surface by:

$$\mathbf{B} = \frac{g}{\Theta} \nabla_p \Theta$$

(1)

Therefore, the author chooses to define a front as regions where the magnitude of the baroclinity is large. It may be noted that this would result in the placement of a front, on a weather map, in the interior of the zone of large temperature gradient rather than on the warm edge of the zone as is the common practice of weather analysts.

With a front defined by equation 1, the frontogenesis ($\mathbf{F}$) is defined to be the local time rate change of the baroclinity:

$$\mathbf{F} = \frac{\partial}{\partial t} \left[ \mathbf{B} \right] = g \nabla_p \Theta \frac{\partial}{\partial t} \left[ \frac{1}{\Theta} \right] + \frac{g}{\Theta} \frac{\partial}{\partial t} \left[ \nabla_p \Theta \right]$$
TABLE 1. List of symbols and operators.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>g</td>
<td>Acceleration due to gravity (assumed constant)</td>
</tr>
<tr>
<td>$R_d$</td>
<td>Gas constant for dry air</td>
</tr>
<tr>
<td>$C_p$</td>
<td>Specific heat capacity for dry air</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>The ratio $R_d/C_p = 0.286$</td>
</tr>
<tr>
<td>$\rho$</td>
<td>Air density</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Specific volume</td>
</tr>
<tr>
<td>$p$</td>
<td>Pressure</td>
</tr>
<tr>
<td>$T$</td>
<td>Temperature</td>
</tr>
<tr>
<td>$\Theta$</td>
<td>Potential temperature</td>
</tr>
<tr>
<td>$u$</td>
<td>East-west component of the wind velocity</td>
</tr>
<tr>
<td>$v$</td>
<td>North-south component of the wind velocity</td>
</tr>
<tr>
<td>$\omega$</td>
<td>$\frac{dp}{dt}$; the vertical velocity</td>
</tr>
<tr>
<td>$q$</td>
<td>Specific humidity</td>
</tr>
<tr>
<td>$\vec{B}$</td>
<td>Vector baroclinity on an isobaric surface</td>
</tr>
<tr>
<td>$\vec{F}$</td>
<td>Vector frontogenesis</td>
</tr>
<tr>
<td>$\nabla$</td>
<td>Three-dimensional gradient operator</td>
</tr>
<tr>
<td>$\nabla_p$</td>
<td>Horizontal (more specifically, isobaric) gradient operator</td>
</tr>
<tr>
<td>$\frac{d}{dt}$</td>
<td>$(\frac{3}{3t} + \nabla \cdot \nabla_p + \omega \frac{3}{3p})$ Total time derivative of a variable made up of a local change plus a change due to advection</td>
</tr>
<tr>
<td>$\left( \frac{t}{\Delta t} \right)$</td>
<td>$\frac{1}{\Delta t} \int_{t_1}^{t_2} \left( \right) dt$ Time mean of a variable</td>
</tr>
<tr>
<td>$\left( \right)'$</td>
<td>$(\ ) - \left( \right)$ Perturbation from the time mean</td>
</tr>
<tr>
<td>$\theta_s$</td>
<td>Sea-surface temperature</td>
</tr>
<tr>
<td>$C_D$</td>
<td>Drag coefficient for the bulk aerodynamic method</td>
</tr>
<tr>
<td>$H$</td>
<td>Heat addition</td>
</tr>
</tbody>
</table>
Since the first term on the right is negligible when compared to the second, we obtain after rearranging:

$$IF = \frac{\partial}{\partial t} \mathbf{\nabla} \cdot \mathbf{\nabla} \theta$$  \hspace{1cm} (2)

The local time rate of change of the potential temperature is given by:

$$\frac{\partial \theta}{\partial t} = \frac{d\theta}{dt} - \mathbf{\nabla} \cdot \mathbf{\nabla} \theta - \omega \frac{\partial \theta}{\partial p}$$

Therefore, the frontogenesis equation can be rewritten as

$$IF = \frac{\partial}{\partial t} \left[ |B| \right] = \frac{\mathbf{\nabla} \cdot \mathbf{\nabla} \theta}{\theta} \left[ \frac{d\theta}{dt} - \left( \mathbf{\nabla} \cdot \mathbf{\nabla} \theta + \omega \frac{\partial \theta}{\partial p} \right) \right]$$  \hspace{1cm} (3)

Now, in dealing with mean fronts, it seems intuitive and instructive to examine the time means of the equations describing the baroclinicity and the local time rate of change of the baroclinicity (Equations 1 and 3, respectively). By noting the mean operator of Table 1, we first have

$$\overline{B} = \frac{1}{\Delta t} \int_{t_1}^{t_2} \frac{\mathbf{\nabla} \cdot \mathbf{\nabla} \theta}{\theta} dt$$

By noting that $\frac{\mathbf{\nabla} \cdot \mathbf{\nabla} \theta}{\theta}$ varies little with time compared to $\mathbf{\nabla} \cdot \mathbf{\nabla} \theta$ and that the gradient operator is independent of time, we obtain an equation describing the mean baroclinicity as:

$$\overline{B} = \frac{\mathbf{\nabla} \cdot \mathbf{\nabla} \theta}{\theta}$$  \hspace{1cm} (4)

Next,

$$\overline{F} = \frac{1}{\Delta t} \int_{t_1}^{t_2} \frac{\partial \overline{\theta}}{\partial t} dt = \frac{1}{\Delta t} \int_{t_1}^{t_2} \frac{\mathbf{\nabla} \cdot \mathbf{\nabla} \theta}{\theta} \overline{\theta} dt$$
Term A is equal to the baroclinity at the end of the time period minus the baroclinity at the beginning divided by the time interval. It is quite apparent that, when the magnitude of the baroclinicity is bounded, the magnitude of term A decreases as the length of the time series increases. With a sufficiently long time series this term should converge to zero. Furthermore, if the mean operator is assumed to be a running mean, then the mean of the time derivative of baroclinicity is equal to the time derivative of mean baroclinity. Or,

\[
\overline{F} = \overline{\frac{\partial B}{\partial t}} = \frac{1}{\Delta t} \int_{t_1}^{t_2} \frac{g}{\theta} \nabla_p \frac{\partial \theta}{\partial t} \, dt
\]

The mean processes which act in changing the magnitude and direction of the mean baroclinicity are represented in term B. In order to evaluate the frontogenetic (or frontolytic) effects of both the mean and perturbation fields, we can separate the variables into a mean and a deviation from the mean as described in Table 1. Thus, term B becomes:

\[
\frac{1}{\Delta t} \int_{t_1}^{t_2} \left( \frac{g}{\theta} \nabla_p \frac{\partial \theta}{\partial t} \right) \, dt = \frac{1}{\Delta t} \int_{t_1}^{t_2} \frac{g}{\theta} \nabla_p \left\{ \frac{\partial \theta}{\partial t} - \left[ (\nabla + \nabla') \cdot \nabla_p (\bar{\theta} + \theta') + (\nabla + \omega') \frac{\partial}{\partial p} (\bar{\theta} + \theta') \right] \right\} \, dt
\]

As previously mentioned, we may move the gradient operator and \( \frac{g}{\theta} \) outside the integral operator. Upon expanding the terms on the right-hand side of the above equation we get
\[
\frac{1}{\Delta t} \int_{t_1}^{t_2} \left( \frac{\partial}{\partial \theta} \nabla_p \frac{\partial \theta}{\partial t} \right) dt = \frac{\partial}{\partial \theta} \nabla_p \left[ \frac{\partial \overline{\theta}}{\partial t} - \left( \nabla \cdot \nabla_p \overline{\theta} + \nabla' \cdot \nabla_p \theta' \right) + \omega \frac{\partial \theta}{\partial p} + \omega' \frac{\partial \theta'}{\partial p} \right]
\]

Again, for a sufficiently long time series, the left-hand side of the above equation should converge to zero. We therefore obtain the equation for the mean frontogenesis in the form:

\[
\overline{\nabla \Phi} = \frac{\partial \overline{\theta}}{\partial t} = \frac{\partial}{\partial \theta} \nabla_p \left[ \frac{\partial \overline{\theta}}{\partial t} - \left( \nabla \cdot \nabla_p \overline{\theta} + \nabla' \cdot \nabla_p \theta' \right) + \omega \frac{\partial \theta}{\partial p} + \omega' \frac{\partial \theta'}{\partial p} \right]
\]

(5)

The terms within the brackets of the right-hand side of Equation 5 can be described as:

C) The combined effects of all the mean diabatic heating and cooling processes and subgrid-scale heat transports.

D) Isobaric advection of the mean horizontal potential temperature field by the mean horizontal winds.

E) Averaged isobaric advections of the horizontal perturbation potential temperature field by the perturbation winds.

F) Vertical advection of the mean potential temperature by the mean vertical motion.

G) Averaged vertical advection of the perturbation potential temperature by the perturbation vertical motion.
The isobaric gradient of the sum of these terms thus represent mean frontogenesis. Moreover, because the mean frontogenesis is small (nearly steady state mean), the terms must sum to zero. It may thus be expected that some terms may be frontogenetic while others will be frontolytic.

Conceptually, the above methodology seems to provide an alternative and more illustrative manner of investigating mean frontal zones than previous studies which dealt with this topic by counting fronts. Not only are the mean fronts located, but the processes which act to maintain the magnitude and direction of the baroclinity can be quantitatively investigated.
V. PURPOSE OF THE STUDY

To locate the mean frontal positions of a region, several years of data are normally required. However, in a region covered mostly by water, real synoptic data are at best scarce. In this case, mean frontal positions either are inferred by using variables known to be associated with frontogenesis (such as large scale confluence) or are located by using a limited sample of data.

During the last two weeks of February 1975, AMTEX (Air Mass Transformation Experiment) executed its second field project. Lenschow and Agee (1976) discuss the collection of atmospheric and oceanographic variables and the synoptic events which occurred during the time period (see Table 2). A special array of stations was set up for the experiment to give sufficient data coverage over the East China Sea. The station spacing allowed for observations of synoptic and mesoscale features over the warm Kuroshio ocean current.

This treatise makes use of the wind and temperature data from the soundings of the stations shown in Figure 6 for the time period corresponding to AMTEX 1975. The soundings from stations on the Asian mainland were obtained from the National Climatic Center in Asheville, North Carolina. The soundings from the stations on the main Japanese Islands were obtained from Aerological Data of Japan. Finally, the soundings from the stations on the islands southwest of Japan as well as the surface data from the experimental
<table>
<thead>
<tr>
<th>Period</th>
<th>Feb. 1975</th>
<th>Major Synoptic Features</th>
</tr>
</thead>
<tbody>
<tr>
<td>Warm</td>
<td>13</td>
<td>Developing Taiwanese depression.</td>
</tr>
<tr>
<td></td>
<td>14</td>
<td>Taiwanese low NW Okinawa (1004 mb); cold front passage at Naha.</td>
</tr>
<tr>
<td>Cold</td>
<td>15</td>
<td>Taiwanese low east of Japan (982 mb); cold air outbreak (MCC).</td>
</tr>
<tr>
<td></td>
<td>16</td>
<td>Cold air outbreak continues (MCC).</td>
</tr>
<tr>
<td></td>
<td>17</td>
<td>Cold air outbreak weakens (MCC).</td>
</tr>
<tr>
<td>Cool</td>
<td>18</td>
<td>Trough low pressure over Yellow Sea.</td>
</tr>
<tr>
<td></td>
<td>19</td>
<td>Trough of low pressure east of Korea; cold air over Yellow Sea.</td>
</tr>
<tr>
<td>Cold</td>
<td>20</td>
<td>Cyclone east of Japan with PVA south of Korea; surge line.</td>
</tr>
<tr>
<td></td>
<td>21</td>
<td>Cold air outbreak (MCC).</td>
</tr>
<tr>
<td></td>
<td>22</td>
<td>Cold air outbreak continues (MCC).</td>
</tr>
<tr>
<td></td>
<td>23</td>
<td>Anticyclone over East China Sea; cold air outbreak weakens (MCC).</td>
</tr>
<tr>
<td>Warm</td>
<td>24</td>
<td>Anticyclone over Japan.</td>
</tr>
<tr>
<td></td>
<td>25</td>
<td>Cyclone over Sea of Japan NE of Korea.</td>
</tr>
<tr>
<td></td>
<td>26</td>
<td>East-west cold front over Korea.</td>
</tr>
<tr>
<td></td>
<td>27</td>
<td>East-west cold front over AMTEX region; frontal wave NW of Okinawa.</td>
</tr>
<tr>
<td></td>
<td>28</td>
<td>Weak cold air outbreak (MCC).</td>
</tr>
</tbody>
</table>

*From Lenschow and Agee (1976).*
FIGURE 6. Location of stations from which sounding data was obtained for this study. Also warm ocean currents (solid lines) and cold ocean currents (dashed lines) are approximately located. The boundary of the computational region is also identified.
ships used during AMTEX were obtained from the AMTEX '75 Data Report Volume 2.

It is understood that two weeks of data is hardly enough to constitute a study in synoptic climatology. However, several points can be raised to justify the short time period used for this study.

1) Since this thesis presents a new method for analyzing mean frontal positions, investigations by other researchers of AMTEX provide needed checks on the magnitude of terms in the frontogenesis equation.

2) Though the station spacing in some areas of the computational region leaves much to be desired, the supplemental data provided by AMTEX allows more accurate analyses to be made than otherwise possible, especially over the Kuroshio current where active cyclogenesis is common.

3) The two weeks during AMTEX 1975 appear to be "typical" in terms of weather sequence over the lands and seas of the Far East. This can be seen by relating the sequence of weather events of Table 2 to the following discussion of the synoptic climatology of the area.

The region of East Asia and the West Pacific is recognized as one of the most meteorologically active on the earth. Topography is known to be important in diversifying the synoptic situations which commonly occur over the region.
During the winter months, radiational cooling of the lower troposphere over Eurasia is a major factor for the development of a cold shallow air mass associated with the strong anticyclone situated over Siberia and Manchuria. This, coupled with the semi-permanent Aleutian low-pressure pattern, establishes the winter monsoon regime of the Far East. Surges of very cold, dry and stable air burst out of the Siberian anticyclone at intervals of four to seven days (Nitta et al., 1973 and Murahami, 1979). The surging cold air being only 1000-2000 meters in depth is greatly influenced by east-west hill ranges (such as the Chinling Shan and the Nanling Shan) and the warm ocean currents off the Asian East Coast.

Concurrent with the establishment of the winter monsoon regime in the lower levels, is the anchoring of the subtropical jet stream in the upper troposphere south of the Tibetan Plateau. Yeh (1950), Bolin (1950), Mohri (1953), and others, discuss the thermal and dynamical effects of the Tibetan Plateau on the small latitudinal variation of the subtropical jet. The polar front jet stream which in the mean extends across Siberia and North China and reaches West Japan fluctuates daily and monthly over a much wider latitudinal range. Frequently during the winter months the subtropical and polar front jet streams form a region of confluence downstream of the Tibetan Plateau. Namias and Clapp (1949) have shown that this situation leads to intensification of the jet downstream. It is no surprise then that the highest wind speeds recorded in the upper troposphere are found in the vicinity of Japan.
In connection with the jet streams of the region, wave depressions are very common, and tend to have preferred areas of formation and development. Nitta et al. (1973), Yoshizaki (1974), and Murakami (1979) have used spectral techniques to analyze the disturbances. Two types appear to predominate. The first is a long wave (from here to be termed synoptic scale wave) with a wavelength of approximately 4000-5000 km and a period of four to five days. The second, associated with medium-scale disturbances, has a wavelength of about 1000-2000 km and a period of 1.5 to 2 days. Both types appear to have the vertical structure of unstable baroclinic waves. Indeed, all the growing waves of the study by Nitta et al. (1973) were associated with a 500 mb trough, whereas the one damped case had no corresponding upper-level pattern. The medium-scale disturbance is primarily found below 600 mb and develop in the atmosphere of low Richardson number as suggested by theoretical studies (Gambo, 1970a, b and Tokioka, 1970).

The synoptic-scale waves are frequently associated with the surface wave depressions of the Yangtze Valley. The surface depressions form on the cold air surging southward out of the Siberian anticyclone. As the baroclinic cyclones develop, they propagate eastward and north-eastward. Upon crossing the warm water off the Asian East Coast they rapidly intensify while continuing to move northeast to become absorbed in the Aleutian semi-permanent low-pressure system.
With the passage of the cold front, the cold, dry and stable air flows out over the warm water. The sharp temperature contrast between the cold air mass and the warm water favors strong heat and moisture transfer at the air-sea interface. This air mass modification has pronounced effects on all scales of atmospheric motions.

The onset of medium-scale disturbances is less clear. They are commonly thought to generate over the East China Sea and propagate northward and north-eastward (for a detailed introductory review on the study of medium-scale waves, see Yoshizumi, 1976). More recently, Saito (1977) has found that the sea-level cyclogenesis commenced when a pre-existing cyclonic vortex at about 850 mb moved over the sea from the continent. While questions remain as to the origin of medium-scale disturbances, their structure is well documented. These systems are known to intensify in a large-scale baroclinic zone. They have the greatest intensification in the lowest layers and a rapid decrease of deepening with height.

There is overwhelming observational evidence of atmospheric circulation systems of all scales intensifying over the warm oceans to the east of the major continents. This suggests that the large energy input to the atmosphere from the warm oceans is responsible for the development of the disturbances.

Min and Horn (1974) have studied the climatology of the generation of available potential energy in the areas of the east coast of Asia and North America. They note that the generation term depends not only on the rate of diabatic heating, but also on an "efficiency
factor" which states that heat addition at relatively high pressure on an isentropic surface creates available potential energy. Since the isentropic surfaces slope upward from the warm oceans over the cold continents, the efficiency factors are greatest, as is the rate of diabatic heating, over the warm oceans. It is to be expected then that the development of cyclones would be frequent and intense in these areas of excess available potential energy.

AMTEX was undertaken in an effort to clarify the processes by which heat, moisture and momentum are transferred from the ocean surface through the boundary layer and into the free atmosphere overlying the East China Sea. To find answers to the problem of how the large energy transfer influences atmospheric motions on scales ranging from individual cumuli to synoptic-scale disturbances is a primary motive of the experiment.

This investigation, being concerned with mean frontal zones recognizes the opportunity presented by the expanded data coverage (over an area of high frontal frequency) provided by AMTEX. It is the purpose of this study to show that the previously developed methodology is applicable for investigating this topic. This will be accomplished by identifying the mean large-scale baroclinic zones at different levels over the Far East for the time period of AMTEX 1975. Furthermore, having identified the mean baroclinic zones, the terms in the mean frontogenesis equation will be evaluated in an effort to illustrate the dominant processes involved in maintaining the mean baroclinity field.
VI. DESCRIPTION OF THE DATA

The regular 00Z and 12Z soundings, as transmitted by the 68 stations shown on Figure 6, were used to obtain raw wind and temperature data. The data was extracted from the soundings for the period of 00Z 14 February 1975 through 12Z 28 February 1975 at the 1000, 850, 700 and 500 mb surfaces. Lack of higher level data from many stations on the Asian mainland restricted the analyses of data to those pressure surfaces. Also, the lack of data at each level by the mainland stations at 00Z 24 February resulted in that time period being eliminated from the study.

Considerable effort was undertaken in eliminating possible erroneous values from the data set. This was accomplished by a) checking for superadiabatic lapse rates in the potential temperature profile, and b) identifying extraordinarily large values of temperature gradient as calculated from the thermal wind relationship. Then, suspected bad data was compared with analyses presented in the AMTEX 1975 Data Report Volume 7. Data was only eliminated when found to deviate significantly from the analyses. Since some stations report infrequently, no attempt was made to fill in missing observations. The objective was to obtain as good a data coverage as possible for each day and height without producing possibly fictitious data. The number of stations reporting at each level and for each time period is given in Table 3.
TABLE 3.  Number of stations reporting observations at each time period.

<table>
<thead>
<tr>
<th>Feb. 1975</th>
<th>1000</th>
<th>850</th>
<th>700</th>
<th>500</th>
</tr>
</thead>
<tbody>
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<td>43</td>
<td>58</td>
<td>52</td>
</tr>
<tr>
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<tr>
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<tr>
<td>12Z 28</td>
<td>43</td>
<td>40</td>
<td>57</td>
<td>54</td>
</tr>
</tbody>
</table>
The u and v components of the wind, as well as the potential
temperature (\( \Theta \)), were interpolated to a latitude-longitude grid
with two-degree grid interval (see Figure 7) for each level and time
period by using a modified Barnes scheme. The interpolating and
filtering aspects of the conventional Barnes scheme are discussed at
length in Barnes (1964) and Barnes (1973).

Since the density of stations used in this study is far from
uniform, the conventional scheme was modified. The interpolating
scheme used for this study is

\[
\lambda_i = \frac{\sum_{j=1}^{N} \lambda_j c_j \exp \left( -\frac{D_{i,j}^2}{Q_k} \right)}{\sum_{j=1}^{N} c_j \exp \left( -\frac{D_{i,j}^2}{Q_k} \right)}
\]

where

- \( \lambda_i \) is the value of the variable at gridpoint i.
- \( \lambda_j \) is the value of the variable at station j.
- \( D_{i,j} \) is the distance between gridpoint i and station j.
- \( Q_k \) is a specified constant for the k\(^{th}\) pass through the
  scheme. (In this case, a two-pass scheme with \( Q_1 \) and
  \( Q_2 \) being 21 and 7 latitude degrees squared, respectively)
- \( N \) is the number of stations reporting at the time and
  for the height for which the interpolation was per-
  formed.
- \( c_j \) is a weight computed for each station j describing
  the local station density of its region. It is equal
FIGURE 7. Gridpoints at which the variables are interpolated to.
to the inverse of the number of stations which lie within a $7^\circ$ latitude by $7^\circ$ longitude box centered on station $j$. ($C_j$ changes as the number of reporting stations change with variable and time period.)

As with any interpolation scheme, errors are inherent. Since this treatise deals with locating mean large-scale baroclinic zones and identifying processes which maintain these mean zones, it is of interest to know the areas in which the interpolating scheme both concentrates and spreads the gradient of a variable.

To this end, a smooth bogus field was synthesized which resembles the mean temperature distribution of the region; low values are synthesized to the north over the continent, with higher values to the south and east over the warm waters. Care was taken to produce a strong constant gradient magnitude field. I approximated station values from this field and ran the interpolating scheme using these values.

Finite centered differencing is used to evaluate the gradients for this study. Analyzed on Figure 8 is the percentage difference of the magnitude of the bogus field from the value averaged over the computational region. As much as 30 percent deviation might be due to subjective errors due to synthesizing the bogus field and interpolating to the stations. But there are areas where we must attribute much of the observed deviation to the problems of the interpolating scheme. The computational problem areas are believed
FIGURE 8. Percentage difference of the magnitude of the gradient of the bogus field from the averaged value computed over the region. The sign of the difference is also indicated. Stippled areas enclose the areas where the percentage difference is more than 40 percent.
to be those regions which have values greater than or less than 40 percent of the average value (stippled areas).

Both time and funds restricted further investigations of ways to even further reduce the schemes negative aspects. Readers are referred to Doswell (1977) for a more theoretical discussion of the "Root Mean Squared" interpolation error associated with weighted averaged analysis schemes. In that paper, he notes that the RMS error will increase with the randomness of the station spacing and that boundary problems are associated with a phase shift in the analysis as the limits of the grid are approached.

It is important to note that errors are involved with any analysis scheme. Indeed, even manually-produced analyses are restricted by available data so that analysts are forced to use their knowledge of the physics of the phenomena to try to improve the output. As previously mentioned, Yoshino (1971) implies that this is the reason his principal zones of high frontal frequencies are north of those analyzed by Reed (1960).

Since the terms of the mean frontogenesis equation are highly non-linear, they are sensitive to noise in the analyses. Therefore, fairly large smoothing parameters (21 and 7 latitude degrees squared) were used in the interpolating scheme to produce fields which give meaningful results while retaining as much of the information in the original data as possible.

With gridpoint values of the wind components at the four levels, vertical velocities ($\omega = \frac{dp}{dt}$) were then calculated via
the kinematic method for each level and time period. As with any method of obtaining vertical velocities, errors are unavoidable and are a result of the underlying assumptions made and the computational scheme used. The following are the more notable sources of error for the $\omega$ fields of this study:

1) The usual lower boundary condition is assumed that $\omega$ is everywhere zero at 1000 mb.

2) Usually, $\omega$ is assumed to be zero also near the level of the tropopause. To satisfy this assumption while using the kinematic method, a correction is normally made to the $\omega$'s in the vertical to smoothly force the vertical motion to satisfy this condition (O'Brien, 1970). The argument correctly being that errors in the divergence fields in the lower levels propagate vertically to add to the errors of the divergence fields in the upper levels. The latter are assumed to be large due to the high wind speeds found there. However, due to a lack of data above 500 mb no upper boundary condition is assumed in this study. This may not be too drastic, however, since the corrections normally made are largest above this level. The consequence, though, is that the $\omega$ fields at 500 mb appear noisier than in the lower levels.

3) Since the kinematic method is based on derivatives of the wind field, the largest errors at each level are to be
expected in the regions where the interpolating scheme shows problems with reproducing the actual gradients (Figure 8).

4) Also, some stations, particularly near the boundary of the computational region, report wind observations infrequently. Figure 9 shows the mean divergence field at 1000 mb upon which subsequent mean \(\omega\)-values in the vertical are partly based. The dark solid line enclose the stations which report winds more than 66 percent of the time. It can easily be noted that not only would we expect large errors outside the enclosed region at particular time periods, but we also should be aware of averaging any term which is a function of the winds near the boundaries.

5) The \(\omega\)'s at each level are a result of averaging the divergence through deep layers. This assumes that the wind varies linearly throughout the layer and this assumption may be unrealistic at times.

In support of the vertical motion fields of this study, it is noted that there is some correlation between these fields and those computed by the six-level quasi-geostrophic model of the Japanese Meteorological Agency. Figure 10 is a scatter diagram of this study's omega values versus those appearing on the analyses in the AMTEX 1975 Data Report Volume 7 at 700 mb for 130 degrees E by 30 degrees N.
FIGURE 9. Mean divergence field at 1000 mb (thin line). Heavy solid line encloses stations reporting wind observations more than 66 per cent of the time.
FIGURE 10. Scatter diagram of the $\omega$-values of this study versus those of the Japanese six-level quasi-geostrophic model at 700 mb for 130 degrees E by 30 degrees N.
To briefly summarize, real synoptic data from the stations shown on Figure 6 was input into an interpolating scheme which produced gridpoint data (Figure 7) of potential temperature and wind components at 1000, 850, 700, and 500 mb for 29 of the synoptic time periods corresponding to AMTEX 1975. The interpolating scheme, as with all interpolating schemes, has definite problems but is believed to be as good as any in reproducing the large scale features which can be resolved by the available data. Gridpoint values of $\omega$ were calculated and are believed to be reasonable. However, due to the method of computation, the $\omega$ fields are generally noisier, especially in the upper-levels, than the other three variables.
VII. PRESENTATION OF THE MEAN FIELDS

Upon completing the interpolation of $u$, $v$ and $\theta$ and the calculations of $\omega$ the four variables were averaged at the gridpoints over the time series. In this section, the mean conditions are presented while discussing some of the resulting observations and consequences by comparing and contrasting them to previous investigations.

Appearing on the figures showing the mean thermal pattern are: a) potential temperature (K), and b) regions of large magnitude of baroclinity. The magnitude of the baroclinity within frontal zones is known to vary in the vertical. Table 4 gives the threshold value of mean baroclinity of which areas having larger values are defined to be the mean fronts of this thesis. On the figures illustrating the mean horizontal flow are: a) streamlines, and b) isotachs (meters per second). The mean vertical motion (millibar per hour) is shown in accompanying figures. Positive values of $\omega$ indicate large-scale subsidence while negative values correspond to rising motion.
TABLE 4. Threshold values of the magnitude of mean baroclinity at each level. Areas which have values larger than these identify the large-scale baroclinic zones of this study.

<table>
<thead>
<tr>
<th>Level</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>$5.0 \times 10^{-7}$ s$^{-2}$</td>
</tr>
<tr>
<td>850</td>
<td>$4.0 \times 10^{-7}$ s$^{-2}$</td>
</tr>
<tr>
<td>700</td>
<td>$5.0 \times 10^{-7}$ s$^{-2}$</td>
</tr>
<tr>
<td>500</td>
<td>$7.0 \times 10^{-7}$ s$^{-2}$</td>
</tr>
</tbody>
</table>
At 1000 mb (Figure 11) the coldest air is located over Manchuria with the mean thermal trough extending over northern Korea, the Yellow Sea, and eastern China. There is a tendency for the isotherms to "bulge" northward over the East China Sea, the Sea of Japan and the waters southeast of Japan. The higher mean temperatures are a reflection of both the synoptic situations typical to the area and the associated diabatic processes. First, cyclones are common and are characterized by the northward transport of warm moist subtropical air. Then after the passage of the disturbance and the occurrence of a cold air outbreak, intense diabatic heating of the air overlying these warm bodies of water ($\sim 500 \text{ Wm}^{-2}$) is prevalent (Ninomiya, 1972).

The large-scale baroclinic zones compare well with the mean frontal zones of Reed (1961) and Yoshino (1971) (see Figure 5). However, Yoshino notes that the baroclinic zone extending from Manchuria, over Korea and into the Sea of Japan, corresponding to the eastern portions of his Eurasian Polar front (from here to be termed the northern baroclinic zone), reaches its greatest intensity in the early spring and late autumn when the subtropical jet stream is situated at the northern edge of the Himalayas. The baroclinic zone extending along the Kuroshio current (from here to be termed the southern baroclinic zone) is associated with the Pacific polar frontal zone of previous investigations. For comparison with the
FIGURE 11. Mean thermal field at 1000 mb. Solid lines are potential temperature $(\kappa)_p$; dashed lines are the magnitude of mean baroclinity (times $10^8 s^{-1}$). Stippled areas indicate the large scale baroclinic zones being investigated.
large-scale baroclinic zones of this study, I counted the number of fronts in areas of 4 degrees latitude by 4 degrees longitude appearing on the 00Z and 12Z surface analyses of the AMTEX '75 Data Report Volume 7 (See Figure 12). It can be seen that local maxima of frontal frequency are associated with both the northern and southern baroclinic zones. Also, since closely related, I calculated relative frequencies of baroclinity at the gridpoints for the time period of this investigation. The percentage of times the baroclinity exceeded $4.5 \times 10^{-7} \text{ s}^{-2}$ at 1000 mb is shown on Figure 13. Though this field bears close resemblance to the magnitude of the mean baroclinity, the frequencies may be biased by the analysis difficulties. Therefore, I shall not pursue this manner of investigating mean fronts and continue to investigate the mean frontal zones by evaluating the terms in the frontogenesis equation (Equation 5).

The observations of Miller and Mantis (1946), Reed (1960), and others, relating low level frontal and cyclonic activity with surface topography are supported. It appears as though the processes involved with the different synoptic situations tend to align both the mean baroclinic zones and the mean temperature field with the warm Kuroshio and Tsushima ocean currents.

The mean wind field at 1000 mb (Figure 14) corresponds well in several respects with discussions of Petterssen (1941), Thompson (1951), Mohri (1953), Members of the Academia Sinica (1957), and others, who discuss the evolution and effects of the mean wind
FIGURE 12. Number of times fronts are located in 4-degree latitude by 4-degree longitude sections on the 00Z and 12Z surface analyses of the AMTEX '75 Data Report Vol. 7.
FIGURE 13. The percentage of time the magnitude of baroclinity exceeded $4.5 \times 10^{-8} \text{ S}^{-2}$ at 1000 mb.
FIGURE 14. Mean flow pattern at 1000 mb. Solid lines are streamlines and dashed lines are isotachs (ms$^{-1}$).
regimes of the area. Most notably, as can be seen by the curvature of the streamlines, is the dominance of the mean large-scale pressure distribution, in particular, the Siberian anticyclone and the semi-permanent Aleutian low-pressure system in the North Pacific. This mean flow typifies the winter monsoons of the Far East with the cold air in the north being transported southward and spreading out over China and the warm waters off the east coast of Asia.

The mean isotach pattern at this level indicates relatively low wind speeds except for a maximum south of Korea over the East China Sea. Since diffluence of the mean flow occurs downstream from this speed maximum, it is difficult to estimate the mean divergence by inspection.

At 850 mb the thermal pattern (Figure 15) closely resembles that of 1000 mb. The axis of the thermal trough is approximately at the same location as is the location of the coldest air. The isotherms flatten over the warm bodies of water, indicating that the horizontal variation (between air over the colder surfaces and air over the warm seas) is not as pronounced as at 1000 mb. Thus, the mean stability of the air between 1000 and 850 mb is seen to be the greatest over the Asian continent and the Yellow Sea. (In general, there is an upward increase of potential temperature of 6 to 10 K over the cold surfaces and only 2 to 7 K over the areas of the Kuroshio and Tsushima currents.)

This also is in good agreement with the observed synoptic situations. Ninomiya (1973) notes that when a disturbance is located
FIGURE 15. Mean thermal field at 850 mb. Solid lines are potential temperature (K); dashed lines are the magnitude of mean baroclinity (times 10 s\(^{-1}\)). Stippled areas indicate the large scale baroclinic zones being investigated.
over the East China Sea, the lapse rate is nearly moist adiabatic throughout a deep layer. Then when a cold air outbreak occurs, a mixed layer is formed with the mean inversion height found between 850 and 700 mb. and a dry adiabatic layer below (Murty, 1976; Nitta, 1976; Kondo, 1976; and others). Processes below the inversion layer act to redistribute the heat and moisture additions occurring at the air-sea interface throughout the boundary layer; thus reducing the stability.

The mean flow at 850 mb (Figure 16) still exhibits the influence of the Siberian anticyclone as is evidenced by the anticyclone curvature of the streamlines over East and Southeast China and the southwest portions of the East China Sea. However, the mean high-pressure cell is known to be a shallow feature and the upper-level trough which in the mean is observed to be located off the Asian East Coast (Petterssen, 1956) can be inferred to influence the flow at this level in all but the southwest corner of the computational region. Even the isotach pattern more closely resembles that of upper levels in that the zone of maximum wind speed reflects the position of the mean upper-level jet stream.

Since the wind and temperature fields were analyzed independently, it is encouraging to note that the two fields generally agree. The mean baroclinic zones at 850 mb (as with the mean frontal zones depicted by Yoshino (1971) at this level) are located directly above those at 1000 mb (see Figure 11). The mean wind field should reflect the deep baroclinic layers through the thermal wind rela-
FIGURE 16. Mean flow pattern at 850 mb. Solid lines are streamlines and dashed lines are isotachs (ms⁻¹).
tionship. Indeed, upon closer examination of the mean flow patterns at 1000 and 850 mb, one can note the largest wind shears are located in the baroclinic zones over the warm ocean currents. The thermal wind condition predicts values of mean baroclinity slightly less than observed in the thermal field, however the patterns show good correlation. The mean large-scale baroclinic zones of this study are those areas in which the magnitude of mean baroclinity (see Equation 4) exceed $4.0 \times 10^{-7} \, \text{s}^{-2}$. In general, values calculated from the vertical wind shear are greater than $3.5 \times 10^{-7} \, \text{s}^{-2}$ within those zones while lesser values are observed outside of them.

Furthermore, we may note that over the warm bodies of water, processes act to decrease the stability in the lower levels. These areas are also characterized by a mean deep baroclinic layer as can be seen by noting the large values of wind shear. This implies that over these areas the Richardson number ($R_i$)\(^2\) decreases due to the effects of the mean fields. As previously mentioned, medium-scale disturbances can theoretically develop in an atmosphere characterized by low values of Ri (Gambo 1970a, b, and Tokioka 1970). Mullen (1979), Nitta et al. (1973), Saito (1977), and others, have documented the fact that medium-scale disturbances frequently develop or intensify in these areas.

\(^2\)The Richardson number is defined by

$$R_i = \frac{g}{\theta} \frac{\partial \theta}{\partial p} / \left( \frac{\partial \nu}{\partial p} \right)^2$$
The mean vertical motion field at 850 mb (Figure 17), since calculated via the kinematic method, is also a reflection of the mean low-level divergence field. Rising motion is observed in the eastern sections of the computational region. The large values in the extreme northeast are most likely due to the boundary effects discussed earlier. Most of the region of the Far East is characterized by mean sinking motion with a maximum observed over the Yellow Sea and Southwest Korea.

This pattern is to be expected since, as discussed by Palmén and Newton (1969), as the cold air from the north is transported southward, it subsides and spreads out; though there is a theoretical limit, based on the conservation of absolute potential vorticity, as to its southward progression. The sinking of the cold air is thus associated with mean large-scale divergence in the lower levels as required by mass continuity. Conversely, the rising motion in the eastern sections is associated with convergence.
FIGURE 17. Mean vertical motion field (mb.hr$^{-1}$) for 850 mb.
Mid-Tropospheric Levels

The mid- and upper-troposphere over the Far East is characterized by the presence of two upper-level jet streams. As previously mentioned, during the winter months, the subtropical jet stream maintains a fairly constant position while the polar front jet stream is more migratory. The fact that these two jet streams are inherent to the region has pronounced effects on the mean wind and thermal fields of the area. Mohri (1953) notes that these two belts of strong winds seem to be closely related to two distinct surface frontal systems, the one on the southern sea, the other on the northern sea of the Japanese Islands. Sometimes, associated with the highly developed jet streams, deep nearly-isothermal layers are observed in the middle- and upper-troposphere over Japanese stations.

The mean thermal fields at 700 and 500 mb (see Figures 18 and 20) of this study exhibit the presence of two deep baroclinic layers. The mean baroclinic zones at each of these levels are directly above those found in the lower levels. Namias and Clapp (1949) related the regions of strongest winds on mean winter charts to the process of confluence. They point out that the mean subtropical and polar front jet streams come closest together over the Far East. The confluence acts to bring cold and warm air masses together which results in an intensified solenoid field (baroclinity).
FIGURE 18. Mean thermal field at 700 mb. Solid lines are potential temperature ($\theta$) $^2$, dashed lines are the magnitude of mean baroclinity (times $10^2$). Stippled areas indicate the large scale baroclinic zones being investigated.
The mean deep baroclinic layers are evident not only in the thermal field, but also in the mean wind field (see Figures 19 and 21). Associated with each of the mean baroclinic zones of this study is an increase (in the vertical) of the wind speed which, in part, satisfies the thermal wind relationship. While the mean subtropical and polar front jet streams are situated above 250 mb, mean wind speeds, already greater than 30 ms⁻¹, are observed in the areas of the mean baroclinic zones at 500 mb.

The mean streamlines at 700 and 500 mb indicate that the flow is northwesterly over the Asian continent and become westerly east of Japan. Since the winds above the surface layer are nearly geostrophic, one would conclude that the mean trough of this study is situated in approximately the same location as the mean trough depicted by Petterssen (1956).

The mean vertical motion fields of the mid-troposphere (see Figures 22 and 23) are characterized by strong sinking motion over the colder surface areas (i.e., Korea, Japan, and the Yellow Sea) and either weak sinking or rising motion over the ocean areas of the Kuroshio and Tsushima currents. This also is in good agreement with the fact that cold air is being transported southward and sinking over much of the area and that synoptic-scale rising motion, in general, is associated with disturbances which are most common over the warm bodies of water (Miller and Mantis, 1947). It should be pointed out again that the vertical motion fields of this study probably contain considerable error particularly in the upper
FIGURE 19. Mean flow pattern at 700 mb. Solid lines are streamlines and dashed lines are isotachs (ms⁻¹).
FIGURE 20. Mean thermal field at 500 mb. Solid lines are potential temperature ($\theta$); dashed lines are the magnitude of mean baroclinity (times 10° s$^{-2}$). Stippled areas indicate the large scale baroclinic zones being investigated.
FIGURE 21. Mean flow pattern at 500 mb. Solid lines are streamlines and dashed lines are isotachs (ms$^{-1}$).
FIGURE 22. Mean vertical motion field (mbh$^{-1}$) for 700 mb.
FIGURE 23. Mean vertical motion field (mb h$^{-1}$) at 500 mb.
levels. However, it is believed that much of the error may be random and filtered out of the mean field, although there appears to be a bias near the boundary of the computational region due to reasons discussed earlier.

Furthermore, at selected gridpoints at 700 mb. I averaged the \( \omega \) values produced by the Japanese Meteorological Agency's six-level quasi-geostrophic model which are reproduced in the AMTEX '75 Data Report Volume 7. Table 5 gives a comparison of those with the mean 700 mb values of this study. In support of the \( \omega \) values of this study, it can be seen that, in general, the signs of the \( \omega \) values are the same but that the magnitudes differ.

In conclusion, it is the belief of the author, that the mean fields presented in this section for the last two weeks of February 1975 are a fair representation of the actual wintertime mean situation. This is clearly evident by comparing the discussion in this section to those of the many authors referred to throughout this and previous sections. By fair representation I do not mean that the magnitudes of these fields nor the locations of the large-scale features match exactly the longer term mean. Indeed, the chances of any two-week average matching the long-term mean in even a few respects is very remote. Of interest to this study is that the mean large-scale processes, peculiar to the east side of the major continents along which warm ocean currents are observed and which have been investigated and discussed by the above mentioned authors, appear to be the same mean processes acting over the time period being investigated here.
TABLE 5. Mean 700 mb $\omega$-values (mb hr$^{-1}$) of this study ($\omega_s$) and those of the Japanese meteorological six-level quasi-geostrophic model ($\omega_j$) for various grid-points over the analysis domain.

<table>
<thead>
<tr>
<th>Long (E) by Lat (N)</th>
<th>$\omega_s$</th>
<th>$\omega_j$</th>
</tr>
</thead>
<tbody>
<tr>
<td>120 x 40</td>
<td>0.7</td>
<td>0.5</td>
</tr>
<tr>
<td>120 x 35</td>
<td>2.3</td>
<td>1.9</td>
</tr>
<tr>
<td>120 x 30</td>
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</tr>
<tr>
<td>120 x 25</td>
<td>0.3</td>
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<td>3.9</td>
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<tr>
<td>130 x 35</td>
<td>0.4</td>
<td>0.4</td>
</tr>
<tr>
<td>130 x 30</td>
<td>0.5</td>
<td>0.2</td>
</tr>
<tr>
<td>140 x 35</td>
<td>-0.5</td>
<td>0.2</td>
</tr>
<tr>
<td>140 x 30</td>
<td>-2.8</td>
<td>-0.7</td>
</tr>
</tbody>
</table>
It is encouraging to note, moreover, that the structure and locations of most of the large-scale features observed in the long-term mean also appear in the means of this study. Particularly to be noted, since this presentation deals with the mean large-scale frontal zones, is the appearance and location of the northern and southern baroclinic zones which are clearly evident in both the mean horizontal thermal field and the vertical structure of the mean wind field.
VIII. THE LARGE-SCALE THERMODYNAMIC BALANCES

The terms in the mean frontogenesis equation involve computing the gradients of non-linear fields. Before proceeding to the results of those computations it seems worthwhile to present and discuss the fields involved in the mean local time rate of change of potential temperature. The mean frontogenesis equation involves calculating the isobaric gradient of those fields (see Equation 5).

At each level and for each time period the perturbations of $u$, $v$, $\theta$ and $\omega$ were obtained by subtracting the mean fields from the original ones (see Table 1). Then the east-west and north-south components of the gradients of a) the mean potential temperature, and b) each perturbation potential temperature field were computed for each pressure level by using finite centered differencing and assuming a spherical earth.

The fields of mean stability, as well as the perturbation stabilities were evaluated using the available information in the vertical (i.e., at 1000, 850, 700, and 500 mb). Since data was obtained for only these levels, and since the pressure interval between them is not equal, the stability at those levels are assumed to be equal to the stability at some intermediate level as required by the centered differencing technique. This method assumes that the vertical profile of potential temperature varies linearly throughout the deep layers over which the differencing was performed. This may be unrepresentative in some cases, especially in
the lower levels over the warm waters, for as discussed by Murty (1976), Nitta (1976), and others, the mean mixed layer inversion is found between 850 and 700 mb, particularly after a cold air outbreak has occurred.

The mean and perturbation advective terms were then calculated and the perturbation advective terms were averaged over the time period. Since these terms involve the product of two fields, errors corresponding to wavelengths smaller than either of the two fields, undoubtedly are included in the resulting computation. A moving nine-point filter was used to remove these spurious fluctuations while retaining the information whose wavelength corresponds to those of the original fields.

From Equation 5, we see that the mean frontogenesis equation involves computing isobaric gradients of the terms in the following equation:

\[
\frac{1}{\Delta t} \int_{t_1}^{t_2} \frac{\partial \theta}{\partial t} \, dt = \frac{\partial \theta}{\partial t} - \left[ \nabla \cdot \nabla_p \bar{\theta} + \nabla' \cdot \nabla_p \bar{\theta}' \right] + \frac{F}{\omega} \frac{\partial \bar{\theta}}{\partial p} + \frac{G}{\omega} \frac{\partial \bar{\theta}'}{\partial p}
\]

For a sufficiently long period of time, the mean local time rate change of potential temperature is nearly zero (nearly steady state mean). Even for the relatively short time period of this study, the left hand side of the above equation is an order of magnitude less than the other terms at 1000 mb and 850 mb. At 700 mb and 500 mb the maximum value is less than \(1.4 \times 10^{-5} \text{ K s}^{-1}\). This value corresponds
to an 18 degree temperature change over the two week period. Thus we can evaluate the mean diabatic heating term (term C above) as a residual of the remaining terms in the above equation. This term thus includes both the direct diabatic effects (latent heat release and flux divergence of radiation) and the effects of subgrid-scale heat transfers.

The diagrams included in this section serve not only to illustrate the distribution of these terms, but also act as a reference when discussing the computations of mean frontogenesis. Therefore, particular attention should be made of where and how the gradients of these fields vary over the region.
Mean Diabatic Heating

The magnitude of the diabatic heating term over the East China Sea and the Sea of Japan fluctuates according to the synoptic situation. According to Ninomiya (1973), the synoptic situations over the area can be classified into two categories: (1) the "cold air outbreak" characterized by the southward transport of the cold, dry continental air mass, and (2) the "warm period" characterized by the northward invasions of the warm moist subtropical air mass, which occurs when a disturbance is located in the area.

The bulk aerodynamic method (Kraus, 1967) relates the magnitude of sensible heat transfer between the ocean and the atmosphere to the surface wind speed and air-sea temperature difference. Similarly the latent heat transfer is proportional to the surface wind speed and the air-sea vapor pressure difference. Thus, when the cold air mass spreads out over the ocean areas of the Far East, intense heating from the surface is observed over the East China Sea and the Sea of Japan.

Manabe (1957, 1958) investigated a heat and moisture budget analysis and estimated heat and moisture supply from the Sea of Japan in winter. He found that the mean amount of transfer is about 250 Wm$^{-2}$ in winter but that a much larger value of heating, as much as 500 Wm$^{-2}$, is supplied during the period of cold air outbreak. Recently, detailed heat and moisture budgets for the East China Sea have been performed using AMTEX data. The budgets were evaluated by
Nitta (1976) for AMTEX 1974 and by Murty (1976) for AMTEX 1975. Kondo (1976) used data of the experimental ships participating in AMTEX to evaluate the energy transfers for AMTEX 1974 and AMTEX 1975 via the bulk aerodynamic method. He compared his results with those of Nitta (1976) and Murty (1976); Table 6 is the summary of his results. It can be seen that the budget method and bulk aerodynamic method of evaluating the diabatic heating give similar results. The mean diabatic heating over the East China Sea is presumably between 270 and 430 Wm$^{-2}$. Also, there is a wide variation in the daily values, ranging from 35 Wm$^{-2}$ to 850 Wm$^{-2}$ depending on the synoptic situation.

The mean values for AMTEX compare reasonably with the climatological values obtained by Ninomiya (1972), Mitsuta and Kato (1972) and Kondo (1976). In each of these studies, the mean diabatic heating is evaluated via the bulk aerodynamic method by inserting climatological values of the surface wind speed, the air-sea temperature difference and the rate of evaporation into the appropriate equations. To demonstrate the distribution of sensible heating by this method, I inserted the mean 1000 mb wind speeds of this study along with gridpoint values of air-sea temperature differences found in the Marine Climatic Atlas of the World, Volume 2 into the equations described by the bulk aerodynamic method. Figure 24 illustrates the resulting distribution of sensible heating which closely resembles those of the aforementioned investigators.
TABLE 6. Comparisons between the averaged value of $H + IE$ over the AMTEX area and that over the square area, or that for three Ocean Ship Stations, or the deduced flux from the data at Okinoerabujima Station ($\text{Wm}^{-2}$).

<table>
<thead>
<tr>
<th>Date</th>
<th>Average over the square region</th>
<th>Average over the AMTEX region</th>
<th>Estimated from data at Okinoerabujima</th>
<th>Average for three Ocean Stations</th>
<th>Average for three Ocean Stations</th>
<th>Square region Budget by Murty</th>
</tr>
</thead>
<tbody>
<tr>
<td>Feb</td>
<td>Budget Method by Nitta</td>
<td>Present bulk method</td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>15</td>
<td>193</td>
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<td>285</td>
<td>274</td>
<td>272</td>
<td>288</td>
<td>418</td>
</tr>
</tbody>
</table>

*From Nitta (1976)
FIGURE 24. Mean sensible heating (Wm$^{-2}$) evaluated by the bulk aerodynamic method.
These latter studies indicate that the mean value of diabatic heating becomes less over the Pacific Ocean, the South China Sea and the Yellow Sea. This seems to be a result, primarily, of the lower values of the mean air-sea temperature difference over those areas. This is a consequence of the lower values of mean sea-surface temperatures observed in the Yellow Sea and the fact that the cold air mass frequently does not penetrate the regions of the Pacific Ocean and South China Sea.

The mean distribution of diabatic heating over the Asian continent is less clear although it must be small. Ninomiya (1972) notes that over Southeast China the value is less than 50 Wm\(^{-2}\). Budyko (1958) has estimated the climatological values of sensible heat supply and evaporation rate at Shanghai to be 17 Wm\(^{-2}\) and 11 mm per day, respectively.

The above discussion serves to document the horizontal distribution of mean diabatic heating in the lowest layers of the atmosphere. Estimating the horizontal distribution of mean diabatic heating at higher levels is more complicated especially in "active" regions where data is sparse. No attempt has been made to evaluate the distribution of radiational heating or cooling. Murty (1976), Nitta (1976), and Kondo (1976) note that it is an order of magnitude smaller than other terms in the budget equation.

As previously mentioned, a primary motive of AMTEX is to investigate how the energy input from the East China Sea is redistributed in the vertical. Prior to AMTEX, Ninomiya and Akiyama (1973) used...
satellite and radar observations to study the characteristic features of cloud and echo distribution and their temporal variation over the East Coast of the Asian continent in February 1968. They identify and describe situations of "echo cluster," "weak echo," and "no echo." The following summarizes, qualitatively, each situation in relation to the vertical redistribution of the sensible and latent heat.

The situation of the echo cluster is associated with a disturbance over the area. Southerly winds prevail and the northward transport of the subtropical air mass is associated with large values of moisture convergence in the lower layers (below 600 mb). The intense stable layer associated with the cold air outbreak is replaced with a layer characterized by the moist adiabatic lapse rate. One might conclude that the low-level convergence associated with the disturbance enhances the convective activity and results in the formation of the echo cluster. The release of latent heat due to condensation in the middle troposphere results in the calculations of positive and negative values of the individual change of temperature \( \frac{d\theta}{dt} \) and moisture \( \frac{dm}{dt} \), respectively. Mean precipitation rates of 1.1 mm per hour were observed. Thus, though diabatic heating from the warm oceans is small during the "warm period," the low-level moisture convergence and associated latent heat release complicates the distribution of diabatic heating in the upper levels. Ninomiya (1973) estimates that more than 30 percent of the heat energy, which is supplied from the sea surface, is transported
upward across the 700 mb surface through the work of cumulus convection developing in the thick layer of moist neutral stratification.

After the passage of the disturbance, a "cold air outbreak" takes place and is associated with the "weak echo" or "no echo" situation. A strong stable layer is formed and northerly winds prevail below the inversion and westerlies above it. The heat energy budget is characterized by large values of horizontal divergence of water vapor flux and very large positive values of both the individual change of temperature \(\frac{d\theta}{dt}\) and specific humidity \(\frac{dq}{dt}\) in the lower troposphere below the inversion. The value of both \(\frac{d\theta}{dt}\) and \(\frac{dq}{dt}\) in the middle troposphere (above 700 mb) are, on the contrary, quite small and the convective transport of the total heat energy almost vanishes at the 700 mb surface (Ninomiya, 1973). This would be due to the fact that the top of the cumulus convection does not penetrate the stable layer which develops in the subsiding polar air mass.

The increased data resolution, both spatial and temporal, of AMTEX allowed more detailed and quantitative investigations of the heat and moisture budgets throughout the troposphere over the East China Sea. The studies of Nitta (1976), Murty (1976), Kondo (1976), Ninomiya (1976), and Kung and Siegel (1977) confirm the main results of the older investigations reviewed here. Readers are referred to those investigations for a more detailed description of these subjects.
The fields of mean diabatic heating of this study agree with previous investigations, especially with respect to horizontal distribution. At 1000 mb (Figure 25) a maximum is situated over the East China Sea with a ridge of high values extending into the Sea of Japan and along the axis of the warm Kuroshio ocean current. Values decrease with distance from these ridges so that lower values are noted over Japan, the Asian continent and the colder parts of the Pacific Ocean. The distribution of diabatic heating of this study at 1000 mb is similar to that calculated by the bulk aerodynamic method (see Figure 24).

At 850 mb (Figure 26) the pattern is slightly different from 1000 mb. The ridge of high values is at the same location, but the maximum has shifted from the East China Sea north-eastward to be situated over the Kuroshio current south of the main island of Japan. Also a maximum can be seen over the Sea of Japan. This pattern is consistent with the fact that the many disturbances associated with this area are intensifying as they propagate northeast over the warm waters. The large amounts of released latent heat associated with the disturbances surely affect the distribution of mean diabatic heating above the lower levels. Again, lower values are found over the Asian continent, Japan and the Yellow Sea; which are the areas of both low values of frontal frequencies (Yoshino, 1971) and areas where diabatic heating is small owing to their relatively cold surface temperatures.
FIGURE 25. Mean diabatic heating field (times $10^5$ K s$^{-1}$) at 1000 mb.
FIGURE 26. Mean diabatic heating field (times $10^5$ Ks$^{-1}$) at 850 mb.
The mean diabatic heating at 700 mb (Figure 27) and 500 mb (Figure 28) appear to be more complicated. As in the lower levels, large positive values can be noted over the warm surface areas, but now a maximum can also be noted over the Asian continent. The very large positive values over the southeastern portion of the analysis domain may, in part, be associated with latent heat release, but are largely due to analysis problems in this data-sparse area. Also, mean cooling can be seen over Japan, the Yellow Sea and the southern parts of the East China Sea. At 500 mb mean cooling is occurring over much of the East China Sea where the warm Kuroshio current flows. This latter observation disagrees with the study of Murty (1976) who finds both a mean heat source and moisture sink over the East China Sea at 500 mb during AMTEX 1975.

The discrepancies between the mean diabatic heating fields of this study and those of previous investigations can be explained in part by the different methods of calculation. Since, in this treatise, the term is evaluated as a residual of other terms, the errors inherent to each of those terms are present in the mean diabatic heating fields of this study. In particular, it is believed that the vertical velocities become more noisy in the upper levels and that at these levels, small errors in the direction of the horizontal winds, greatly affect the computation of horizontal advections.
FIGURE 27. Mean diabatic heating field (times $10^5$ K s$^{-1}$) at 700 mb.
FIGURE 28. Mean diabatic heating field (times $10^5$ Ks$^{-1}$) at 500 mb.
In light of previous discussions, it is not surprising to find mean horizontal cold advection over the entire computational region below 500 mb (solid lines on Figures 29 through 31). Kung and Seigel (1977) note that cold dry advection over the Far East in the lower layers greatly increases when the cold air outbreak occurs in association with the northwesterly flow aloft. At 500 mb (Figure 32) mean cold air advection can be seen everywhere except for the southern part of the East China Sea. This disagrees with the results of Kung and Seigel (1977) who find that over this area during AMTEX 1975 very pronounced cold air advection dominated the mid- and upper-troposphere with an intensified jet.

Being the product of two fields, the largest values occur in the regions where the winds cross the isotherms at large angles and where the mean baroclinity and mean wind speeds have large magnitudes.

At 1000 mb (Figure 29) a maximum is situated over the East China Sea with ridges of high values extending northward into the Sea of Japan and eastward along the axis of the Kuroshio current. The maximum is a result of the strong northerly winds flowing almost directly across the mean isotherms in an area where the mean baroclinity is large (see Figures 11 and 14). Over the Pacific Ocean the winds become more westerly and therefore cross the isotherms at smaller angles. Since the winds are fairly uniform in this region, the ridge of high values of cold air advection result from the large
FIGURE 29. Advection (times $10^5$ Ks$^{-1}$) of the mean potential temperature by the mean horizontal wind at 1000 mb.
FIGURE 30. Advection (times $10^5$ K s$^{-1}$) of the mean potential temperature by the mean horizontal wind (solid) and mean vertical motion (dashed) at 850 mb.
FIGURE 31. Advection (times $10^5$ K s$^{-1}$) of the mean potential temperature by the mean horizontal wind (solid) and mean vertical motion (dashed) at 700 mb.
FIGURE 32. Advection (times $10^5$ K s$^{-1}$) of the mean potential temperature by the mean horizontal wind (solid) and mean vertical motion (dashed) at 500 mb.
magnitude of mean baroclinity there. Smaller values occur over Japan, the Asian continent and the colder parts of the Pacific Ocean. In these regions the mean baroclinicity is small and the winds are light.

At 850 mb (Figure 30) the pattern of cold air advection closely resembles that of 1000 mb. Maxima occur over the warm bodies of water where the baroclinity and mean wind speeds are large. Increasingly smaller values are observed as one moves over the colder surfaces. By studying the mean flow (Figure 16) and mean thermal field (Figure 15) of this level, one can note the cause of the cold air advection over the entire region being considered here.

At 700 mb (Figure 31) and 500 mb (Figure 32) the patterns are more complicated. As in the lower levels, high values of cold advection can be seen over the Sea of Japan and over the Kuroshio current south of the main island of Japan. However, at 700 mb, a trough of lower values is present over parts of the East China Sea with the axis crossing the southern baroclinic zone. At 500 mb, over this area, warm air advection is prevalent. Due to the complexities involved with the interactions of the two jet streams common to this area, it is difficult to explain this observation. It should be clear, though, that due to the strong wind speeds at these levels, a small error in either the wind direction or thermal field would result in large errors in the computation of this term. Small values of cold air advections are also noted, as expected, over the regions of relatively cold surface temperatures where the baroclinicity is small.
Mean Vertical Advections

Owing to the mean stability of the atmosphere, the sign of this term is entirely determined by the sign of the vertical motion. Thus, mean sinking is associated with vertical advection of warmer potential temperature, whereas rising motion with vertical advection of colder potential temperature. As noted earlier, the southward transport of cold air is generally accompanied by the air sinking and spreading out in the low levels. Also, the vertical motion about cyclones is such that the warm air in the low levels is lifted above the cold air which is replacing it. These observations are consistent with the maintenance of the mean stability of the atmosphere.

At 850 mb (dashed lines on Figure 30) a maximum positive value (warm advection) can be seen over the Yellow Sea with ridges extending over the cold surface areas. Cold advection is occurring in the eastern sections and troughs of low values of warm advection extend over the Sea of Japan and along the path of the Kuroshio current. These observations are consistent with the discussions of the processes involved with the two types of synoptic situations common to the area. Namely, that the mean stability and mean sinking motion are weakened in the lower levels over the warm waters where cyclones are frequent.

At 700 and 500 mb (dashed lines on Figures 31 and 32) the patterns are similar in many respects to the field at 850 mb. At these levels, though, two maxima are present, one over the Yellow Sea
and another over Japan. The larger magnitudes observed in the middle troposphere are a result of the larger mean vertical velocities there. Also at these levels smaller values of warm advections are observed over the warm seas and even small values of cold advection can be noted over parts of the Sea of Japan and East China Sea at 500 mb. Moreover, cold advection is observed in the eastern portions of the computational region.

As mentioned previously, the vertical velocities are noisy at these levels and the large values of this term observed around the edges reflect the difficulties with the data distribution and boundaries. While much of these fields seem reasonable, in light of the synoptic situations discussed earlier, the large errors (especially in the upper levels) have pronounced effects on the distribution of the gradient of this term.
Perturbation Advection Term

The magnitude and sign of the perturbation terms depend on the overall correlation between the perturbation wind field and the perturbation temperature gradient field throughout the time series. Although the perturbations of this study (being associated with synoptic scale features) are sometimes large in magnitude, the correlations of the above-mentioned components appear to be small. Although the magnitude of the perturbation terms are small, their frontogenetic contribution to maintaining the large-scale baroclinic zones should not be overlooked.

The distribution of the perturbation horizontal advection term may be more easily explained by considering the idealized model shown on Figure 33. On that figure, a perturbation temperature field (dashed lines) is superimposed on a mean temperature field (horizontal solid lines) with the plus and minus sign indicating the areas where the perturbation temperature is greater than and less than the mean temperature, respectively. A cyclone and anticyclone are also depicted (heavy solid lines) with warm air ahead of the cyclone and behind the anticyclone and cold air between. It can easily be seen that the flow about these perturbations results in cold advection to the north and warm advection to the south. Upon averaging over a time series we would thus expect to see local cooling to the north and local warming to the south of the mean path along which the perturbations tend to travel.
Figure 33. Idealized model of perturbations. Thin horizontal lines are mean temperature; dashed lines are temperature perturbations from the mean. The plus and minus signs indicate areas where the perturbation temperatures are greater than and less than the mean respectively. A cyclone and anticyclone are indicated by the heavy solid lines.
The perturbation horizontal advection terms of this study (solid lines on Figures 34 through 37) seem to confirm this observation. The mean path of the perturbations undoubtedly lies within the mean large scale baroclinic zones and in general, we can note the distribution discussed above, particularly at 850 and 700 mb.

The distribution of the perturbation vertical advection term appear as dashed lines on Figures 34 through 37. The sign of this term is either positive or weakly negative throughout the region at each level. This is as expected since, in general, sinking motion is associated with greater stability and rising motion with weaker stability. While the sign of this term is reasonable, the errors involved with calculating the vertical motions of this study surely effect the horizontal distribution of this term so that their frontogenetic effect may be somewhat misrepresented.
FIGURE 34. Averaged advection (times $10^5$ Ks$^{-1}$) of perturbation potential temperature by the horizontal perturbation wind at 1000 mb.
FIGURE 35. Averaged advection (times $10^5 \text{K s}^{-1}$) of perturbation potential temperature by the horizontal perturbation wind (solid) and vertical motion (dashed) at 850 mb.
FIGURE 36. Averaged advection (times $10^5$ K s$^{-1}$) of perturbation potential temperature by the horizontal perturbation wind (solid) and vertical motion (dashed) at 700 mb.
FIGURE 37. Averaged advection (times $10^5 \text{ K s}^{-1}$) of perturbation potential temperature by the horizontal perturbation wind (solid) and vertical motion (dashed) at 500 mb.
Concluding Remarks on the Thermodynamic Fields

A summary of the important physical processes affecting the thermodynamic fields is based on the two types of synoptic situations common to the area. The warm period is associated with a disturbance in the area. Warm moist air is advected from the South China Sea over the East China Sea. The diabatic heating at the surface is small but a large amount of latent heat, associated with the low-level convergence, is released throughout the middle troposphere. When the disturbance moves to the northeast out of the area, a cold air outbreak occurs.

A cold, dry and stable air mass moves southward surging out of the Siberian anticyclone. Cold horizontal advection is prevalent in the low levels. The cold air sinks and spreads out as it moves southward thus warm advection occurs due to the vertical motion. When the cold air moves out over the warm oceans, intense diabatic heating occurs at the sea-surface. However, due to the strong inversion layer connected with the sinking cold air, the vertical transfer of heat only reaches the top of the mixed layer. Above the inversion, the diabatic heating is small and advections are dominated by the dynamics of the upper-level westerlies.

Qualitatively, the fields of the large-scale thermodynamic balance are, for the most part, consistent with results of other investigators. However, quantitative agreement is more complicated. Since the original fields of this presentation are quite smooth, the
magnitude of the terms presented here underestimate their climatological values. An exception to this is in the areas where errors are prominent. While Doswell (1977) gives an explanation for the causes of large RMS errors due to the properties of the interpolating scheme, further insight to irregularities in these fields are obtained by noting the frequency of wind observations by the stations included in this study (refer to Figure 9). Reliable estimates of these terms are expected only in the area where the number of observations by each station is large enough to produce reasonable estimates of the mean situation; especially upon averaging non-linear terms. Thus, while the qualitative discussion of these terms is quite reasonable throughout the computational region, only the area shown on Figure 9 is believed to give reliable values for the computation of mean frontogenesis.
IX. MEAN FRONTOGENESIS

The fields of mean baroclinity, being vector fields, are described by both direction and magnitude. The mean frontal zones of this study have been identified by those regions where the magnitude of the mean baroclinity is large. While (when considering the steady state case) the mean frontogenesis equation describes the processes which maintain the magnitude and direction of the mean baroclinity, it is worth investigating how just the magnitude changes with time. It can easily be shown that this can be written as:

\[
\frac{\partial}{\partial t} |\mathbf{B}| = \frac{|\mathbf{B}|}{|\mathbf{B}|} \cdot \frac{\partial \mathbf{B}}{\partial t}
\]

which from Equation 5 is just

\[
\frac{\partial}{\partial t} |\mathbf{B}| = \frac{|\mathbf{B}|}{|\mathbf{B}|} \cdot \mathbf{F}
\]

By substituting from Equations 4 and 5 this takes the form

\[
\begin{align*}
\frac{\partial}{\partial t} |\mathbf{B}| &= \frac{\nabla \cdot \overline{\theta}}{|\nabla \overline{\theta}|} \cdot \frac{\partial \overline{\theta}}{\partial t} - \left( \nabla \cdot \nabla \overline{\theta} + \nabla' \cdot \nabla \overline{\theta}' \right) \\
&+ \left( \frac{\partial \overline{\theta}}{\partial \mathbf{p}} + \frac{\partial \overline{\theta}'}{\partial \mathbf{p}'} \right)
\end{align*}
\]

Due to their nature, Equation 7 will be referred to as the scalar form of the mean frontogenesis equation and Equation 5 as the vector form. Term A is the local time rate change of the mean baroclinity, term B is a unit vector pointing in the direction of the
gradient of the mean baroclinity, and the remaining terms are as
described above (see page 33).

Having completed the computations of terms C through G, the
east-west and north-south components of the gradient of those terms
were calculated and multiplied by $g/\delta$ to obtain the components of
the terms in the vector form of the mean frontogenesis equation. The
figures in this section illustrate the effect each term has on
maintaining the field of mean baroclinity. For brevity, the fol-
lowing notation will be used on those figures in discussing the
frontogenetic effects of the various terms:

1) MHA - frontogenetic effect of the mean horizontal advect-
   ion term

2) MVA - frontogenetic effect of the mean vertical advect-
   ion term

3) PHA - frontogenetic effect of the average perturbation
   horizontal advection term

4) PVA - frontogenetic effect of the average perturbation
   vertical advection term

5) DHT - frontogenetic effect of the mean diabatic heating
   term

First, the gradient of each term was scalar multiplied by the unit
baroclinity vector (see Equation 7) and these terms (in units of
$10^{-16} \text{ s}^{-3}$) are superimposed on the field of the magnitude of mean
baroclinity (stippling indicating regions of large magnitude). Then
at the gridpoints indicated on Figure 41, diagrams are later pre-
sented which demonstrate the effect each term has on maintaining the
direction of the field of mean baroclinity.
Since the mean baroclinity vector, in general, has a component
directed toward the south, the distribution of the gradients of
terms C through G determine the contribution of those terms upon
maintaining the magnitude of mean baroclinity. In particular, it is
worth mentioning that the gradients change direction when crossing
minima, maxima and the axes of ridges and troughs; this is an impor-
tant fact when discussing the maintenance of the southern baroclinic
zone which lies along the path of the Kuroshio current. As observed
in the previous section, especially in the lower levels, the dis-
tribution of many of the terms involved in the frontogenesis equa-
tion tend to parallel the Kuroshio current.

Thus at 1000 mb we can see that the mean differential hori-
zontal advection term (Figure 38) is frontolytic north of the
southern baroclinic zone, frontogenetic south of it, and has rela-
tively small values within. In effect, the mean flow is advecting
the baroclinic zone southward. However, as discussed by Palmén and
Newton (1969) the cold air mass is sinking and spreading out in the
low levels so that, upon traveling great distances, it is more easily
modified, especially when moving out over vast bodies of warm water.
It should be no surprise then to find that the frontogenetic effects
of the mean diabatic heating term (Figure 39) counteract those of the
mean horizontal advection. Frontogenesis is noted north of the
baroclinic zone, frontolysis south of it and in general, relatively
small values within. This contradicts the belief that diabatic
heating of colder air is always frontolytic; one is reminded, how-
FIGURE 38. Frontogenetic effect \( (\text{times } 10^{16} \text{s}^{-3}) \) of the mean horizontal advection term on changing the magnitude of the mean baroclinity at 1000 mb.
FIGURE 39. Frontogenetic effect (times $10^{16} \cdot 3$) of the mean diabatic heating term on changing the magnitude of the baroclinity at 1000 mb.
ever, of the results of Min and Horn (1974) who show that heat added to the air over the warm oceans east of the major continents is responsible for cyclogenesis there. The mean frontogenetic effects of the perturbations at this level (Figure 40) appear to be as one would expect. Outside the mean baroclinic zones (where cyclones are less frequent) their effects are almost an order of magnitude smaller than the other terms. However, within these zones their effects are more important and cannot be neglected. Frontolysis is occurring due to the perturbations within the southern baroclinic zone, which is consistent with the fact that cyclones transport heat northward, thereby trying to weaken the temperature contrast which prevails between high and low latitudes. The combined effects mentioned above result in the maintenance of the magnitude of the southern baroclinic zone.

Unlike the southern baroclinic zone, which is situated almost entirely above the warm oceans, the northern baroclinic zone is located mainly above the colder surfaces of the Asian mainland, Korea, and the Yellow Sea and is above the relatively warmer surface of the Sea of Japan only in the eastern sections. Therefore, we would expect the frontogenetic effects of the various terms to act differently than they did while maintaining the southern baroclinic zone. Unfortunately, the northern baroclinic zone is situated in an area where the reliability of the terms in the mean frontogenesis equation is highly suspect, owing to the lack of original data over that area (refer to Figure 9). This being the case, I confined the
FIGURE 40. Frontogenetic effect (times $10^{16}$ s$^{-3}$) of the perturbation horizontal advection term on changing the magnitude of the mean baroclinity at 1000 mb.
analyses and discussions of mean frontogenesis to the southern baroclinic zone where the overall effects appear to be better represented.

The distribution of the terms (of the scalar form of the mean frontogenesis equation) in the levels above 1000 mb appear on Figures 41 through 49. (Figures 41 through 43 for 850 mb; Figures 44 through 46 for 700 mb, and Figures 47 through 49 for 500 mb). By referring to those figures, the effects of each term at the different levels can be noted. However, in order to estimate their "overall effect" on maintaining the magnitude of the baroclinity, I spatially averaged the values of the gridpoints lying within the southern baroclinic zone. Table 7 shows the results of those calculations. Thus, it can be noted that at 1000 mb, the effect of the mean differential horizontal advection term is to move the zone southward, while enhancing the overall magnitude of the baroclinity. While the mean diabatic heating term counteracts by trying to move the zone northward, its effect within the zone is almost negligible. Moreover, it is the disturbances which negates the frontogenetic effects of the mean horizontal advection term within the zone.

The maximum magnitude of the mean diabatic heating term almost coincides with the axis of the southern baroclinic zone (see Figures 11 and 25). However, the frontogenetic effect of the mean diabatic heating term is observed to be small there. This suggests that the mean baroclinic zone may have adjusted itself to the mean diabatic heating. This may be more easily seen for the lower levels
FIGURE 41. Frontogenetic effect \( \times 10^{16} \text{s}^{-3} \) of the mean diabatic heating term on changing the magnitude of the mean baroclinity at 850 mb.
FIGURE 42. Frontogenetic effect (times $10^{16}$ s$^{-3}$) of the mean horizontal advection term (solid) and mean vertical advection term (dashed) on changing the magnitude of the mean baroclinity at 850 mb.
FIGURE 43. Frontogenetic effect (times $10^{16} \text{s}^{-3}$) of the perturbation horizontal advection term (solid) and perturbation vertical advection term (dashed) on changing the magnitude of the mean baroclinity at 850 mb.
FIGURE 44. Frontogenetic effect (times $10^{16} \text{s}^{-3}$) of the mean diabatic heating term on changing the magnitude of the mean baroclinity at 700 mb.
FIGURE 45. Frontogenetic effect (times $10^{16} \text{s}^{-3}$) of the mean horizontal advection term (solid) and mean vertical advection term (dashed) on changing the magnitude of the mean baroclinity at 700 mb.
FIGURE 46. Frontogenetic effect (times $10^{16}$ s$^{-3}$) of the perturbation horizontal advection term (solid) and perturbation vertical advection term (dashed) on changing the magnitude of the mean baroclinity at 700 mb.
FIGURE 47. Frontogenetic effect (times $10^{16} s^{-3}$) of the mean diabatic heating term on changing the magnitude of the mean baroclinicity at 500 mb.
FIGURE 48. Frontogenetic effect (times $10^{16}$ s$^{-3}$) of the mean horizontal advection term (solid) and mean vertical advection term (dashed) on changing the magnitude of the mean baroclinity at 500 mb.
FIGURE 49. Frontogenetic effect (times $10^{16} \text{s}^{-3}$) of the perturbation horizontal advection term (solid) and perturbation vertical advection term (dashed) on changing the magnitude of the mean baroclinity at 850 mb.
TABLE 7.  Average frontogenetic effect \( \times 10^{16} \text{ s}^{-3} \) of various terms on maintaining the magnitude of mean baroclinity.

<table>
<thead>
<tr>
<th>Term</th>
<th>1000</th>
<th>850</th>
<th>700</th>
<th>500</th>
</tr>
</thead>
<tbody>
<tr>
<td>MBA</td>
<td>0.286</td>
<td>1.445</td>
<td>0.905</td>
<td>3.840</td>
</tr>
<tr>
<td>MVA</td>
<td>-0.600</td>
<td>-0.435</td>
<td>0.660</td>
<td></td>
</tr>
<tr>
<td>PHA</td>
<td>-0.282</td>
<td>-0.532</td>
<td>-0.485</td>
<td>-0.373</td>
</tr>
<tr>
<td>PVA</td>
<td>0.073</td>
<td>0.110</td>
<td></td>
<td>-0.213</td>
</tr>
<tr>
<td>DHT</td>
<td>0.086</td>
<td>-0.195</td>
<td>-0.025</td>
<td>-3.890</td>
</tr>
</tbody>
</table>
by considering the following simple model of cold air advection in which the temperature varies only in the $x$-direction. The diabatic heating is then given by

$$\frac{d\theta}{dt} = \frac{\partial \theta}{\partial t} + u \frac{\partial \theta}{\partial x}$$

Assuming a steady state temperature field, we obtain

$$\frac{d\theta}{dt} = u \frac{\partial \theta}{\partial x}$$

From the first law of thermodynamics we have

$$\frac{d\theta}{dt} = \left( \frac{1000}{p} \right)^k \frac{1}{c_p} \frac{dH}{dt} = u \frac{\partial \theta}{\partial x}$$

Now, the heat addition can be estimated from the bulk aerodynamic method which states that the heat addition from the oceans is proportional to the wind speed and the air-sea temperature difference

$$\frac{dH}{dt} = \rho C_D u (\theta_s - \theta)$$

Substituting for $\frac{dH}{dt}$ in the previous equation and assuming the heat is distributed through a depth $\Delta p$, we have

$$\left( \frac{1000}{p} \right)^k \frac{g}{\Delta p} \rho C_D (\theta_s - \theta) = \frac{\partial \theta}{\partial x}$$

where $(\theta_s - \theta)$ is the air-sea temperature difference, $C_D$ is the drag coefficient, which has an estimated value of $1.7 \times 10^{-3}$, and the remaining variables have their customary symbols (see Table 1).

Upon substituting representative values for the variables in the above equation we have

$$\frac{\partial \theta}{\partial x} = 8.5 \times 10^{-7} (\theta_s - \theta)$$
Thus we can see that the maximum temperature gradient is located at the location of the maximum air-sea temperature difference. Assuming a constant wind speed, this is also the location of the maximum diabatic heating. Furthermore, the gradient of the diabatic heating is zero at the maximum and therefore, the frontogenetic effect of the diabatic heating is zero there also. The climatological value of the mean air-sea temperature difference has a maximum over the Kuroshio current of about $10^0$C. From the above equation, we would expect to find a maximum temperature gradient there of around $0.85 \times 10^{-5} \circ C m^{-1}$. This corresponds to values of around $1.5 \times 10^{-5} \circ C m^{-1}$ found within the southern baroclinic zone of this presentation.

At 850 mb, the mean horizontal advection is also seen to be important within the zone. One might get a better feel for this term by referring to the fields which depict the mean thermal and wind fields of this level (Figures 12 and 17). In particular, we can see on those figures that the flow shifts from having a predominantly northerly component in the west to having a more westerly component in the east; furthermore, the flow slows down south of the front. Thus, we would expect the confluence aspect of frontogenesis (see Figure 3) to be an important contribution to this mean advection term. The slightly lower value of this term at 700 mb may reflect the fact that the atmosphere becomes more nearly equivalent barotropic at higher levels. The larger-than-normal value of this term at 500 mb undoubtedly reflects the errors involved with computing the vertical motions and horizontal advections at higher levels.
Above 1000 mb, the terms involving the vertical motion must also be considered. At 850 mb and 700 mb the mean vertical advection term combine with the average horizontal perturbation term in trying to reduce the overall magnitude of baroclinity within the zone. Since, at these levels, the effects of the mean diabatic heating term and the average perturbation vertical advection term are almost negligible (refer to Table 7), the previous two terms appear to be responsible for counteracting the frontogenetic effects of the mean horizontal advection term. At 500 mb, the large value due to the mean diabatic heating is a consequence of the aforementioned problems.

While the scalar form of the mean frontogenesis equation illustrates the effects each term has on maintaining the magnitude of the mean baroclinity, the vector form demonstrates, also, the processes involved with maintaining the direction of the mean baroclinity vector. The following figures serve to illustrate this aspect of mean frontogenesis. At 850 mb the vectors of the gradients of terms C through G of Equation 7, along with the unit baroclinity vector, are presented for the gridpoints depicted on Figure 41. These gridpoints were chosen because of their location with respect to the southern baroclinic zone -- gridpoint A to the south, B within, and C to the north of that zone. The 850 mb level is believed to contain smaller errors in each term than higher levels and, unlike the 1000 mb level, explicitly include the influence of vertical motions. Each vector is graphed with north to the top and
east to the right of the page, and such that the length of the vector is proportional to its magnitude. The scalar projection of each term onto the unit baroclinity vector (the component responsible for changing the magnitude of the baroclinity) is also indicated. Pertinent information on the vectors are contained in Table 8; direction being measured in a cyclonic sense with the axis pointing east being zero degrees.

To the north and south of the zone (Figures 50 and 52) the dominance of the mean diabatic heating term and mean horizontal advection term on maintaining the magnitude of the baroclinity is clearly evident. However, the component of each term, perpendicular to the mean baroclinity vector, act to change the direction of the mean baroclinity with time. At each of these gridpoints, both these terms are, in a crude sense, trying to align the baroclinity vector along the direction of the Kuroshio current. While the importance of perturbation terms and the mean vertical advection term is almost negligible when discussing the maintenance of the magnitude of the baroclinity, their importance should not be underestimated when considering the maintenance of the direction. It can be noted that at these gridpoints these latter terms are entirely responsible for counteracting the twisting effects of the previous ones. As can be seen on Figure 51b the vector sum of all the terms is zero, as required by the steady-state assumption.

Also, within the mean baroclinic zone (Figure 51) the relative importance of each term is more complicated. At this gridpoint, as
TABLE 8. Information about the vectors which describe the frontogenetic effects of the various terms upon changing the direction of the mean baroclinity vector.

<table>
<thead>
<tr>
<th></th>
<th>A</th>
<th>B</th>
<th>C</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$28^\circ \times 130^\circ$</td>
<td>$30^\circ \times 130^\circ$</td>
<td>$32^\circ \times 130^\circ$</td>
</tr>
<tr>
<td>Direction (from east)</td>
<td>Magnitude (X $10^{16}$ s$^{-3}$)</td>
<td>Component along UBV</td>
<td>Direction (from east)</td>
</tr>
<tr>
<td>UBV*</td>
<td>268</td>
<td>1</td>
<td>263</td>
</tr>
<tr>
<td>MHA</td>
<td>288</td>
<td>2.84</td>
<td>2.67</td>
</tr>
<tr>
<td>MVA</td>
<td>227</td>
<td>1.05</td>
<td>0.79</td>
</tr>
<tr>
<td>PHA</td>
<td>125</td>
<td>0.62</td>
<td>-0.50</td>
</tr>
<tr>
<td>PVA</td>
<td>226</td>
<td>0.43</td>
<td>0.32</td>
</tr>
<tr>
<td>DHT</td>
<td>82</td>
<td>3.31</td>
<td>-3.29</td>
</tr>
</tbody>
</table>

*UBV = "Unit Baroclinity Vector"
FIGURE 50. Frontogenetic effect of each term upon changing the vector baroclinity at $130^\circ$ E x $28^\circ$ N at 850 mb. (Point A on Figure 41) The length of the "unit baroclinity vector" (UBV) defines the scale ($1.0 \times 10^{-16} \text{ s}^{-3}$) for the other vectors.
FIGURE 51.

a) Frontogenetic effect of each term upon changing the vector baroclinity at 130°E x 30°N at 850 mb. (Point B on Figure 41) The length of the "unit baroclinity vector" $\mathbf{UBV}$ defines the scale $(1.0 \times 10^{-16} \text{ s}^{-3})$ for the other vectors.

b) Illustration of the vector sum of the terms in a.
FIGURE 52. Frontogenetic effect of each term upon changing the vector baroclinity at $130^\circ$ E x $32^\circ$ N at 850 mb. (Point C on Figure 41) The length of the "unit baroclinity vector" (UBV) defines the scale ($1.0 \times 10^{-5}$ s$^{-1}$) for the other vectors.
with the average situation within the zone (see Table 7), the mean vertical advection term and the perturbation horizontal advection term act in the same sense, not only with respect to changing the magnitude of the baroclinity, but also in changing its direction. Unlike the average case, the perturbation vertical advection term is important, at this gridpoint, to maintaining the magnitude of the baroclinity, while the mean horizontal advection term is less important. However, this latter term is responsible in counteracting the twisting effects of the previously-mentioned terms.
X. SUMMARY AND CONCLUSIONS

Previous investigations have been instrumental not only in locating regions of frontal activity, but also in documenting some relationships that exist between the regions of low and high frontal frequency to both the earth's topography and the mean condition of the atmosphere. In this investigation the mean large scale baroclinic zones near the Asian East Coast were investigated for the last two weeks of February 1975. Two zones were presented and are located along the axes of high frontal frequencies depicted by earlier studies of the region's frontal climatology. This is not unexpected, though, since a) fronts commonly form in large-scale baroclinic zones, and b) once formed, fronts tend to be steered by the upper-level jet streams which, themselves, are associated with deep baroclinic zones through the thermal wind relationship.

While some early investigators located mean fronts by using mean atmospheric variables and their knowledge of frontogenesis mechanisms, I have found no attempt to quantitatively identify which mechanisms were actually dominant in maintaining the mean zones. A frontogenesis equation was developed which, when averaged over a time period, aid in clarifying some of these processes. The terms in this equation represent the frontogenetic effects of the mean fields and average effects of perturbations.

These effects were studied to estimate their roles in maintaining the magnitude of the southern baroclinic zone of this study.
It is found that, to the north and south of that zone, the frontogenetic effects of the mean horizontal advection term and the mean diabatic heating term are dominant and act opposite to each other. However, within the zone the mean diabatic heating term is negligible and the frontogenetic effect of the mean horizontal advection term is balanced mainly by the effects of the mean vertical advection term and averaged perturbation horizontal advection term. The effect of the perturbation vertical advection term is observed to be small but this may be due to problems inherent in calculating vertical velocities.

The frontogenetic effects of each term on maintaining the direction of the mean baroclinity field was briefly investigated. For this aspect of mean frontogenesis, it is found that the effects of the mean horizontal advection term and mean diabatic heating term act together to align the mean baroclinity with the warm Kuroshio ocean current. The effects of the mean vertical advection term and averaged perturbation advection terms counteract, thus maintaining the direction of the mean baroclinity. While these results are interesting, it should be noted that this aspect of mean frontogenesis was studied at only three gridpoints through the southern baroclinic zone. Vector averaging the terms within the zone might provide more representative results.
XI. SUGGESTIONS FOR FURTHER RESEARCH

The need to test the applicability of the frontogenesis equation required choosing a region where frontogenesis and cyclogenesis are frequent and also where the terms in the equation have been previously evaluated. However, virtually all areas of the globe experience frontal phenomena at some time and we are limited only by our ability to resolve them with our observations.

It would be interesting to study frontal climatology by this method in regions where mean processes, different from those found in this study may be dominant. For example, downstream of the mean jet stream we would expect to find a thermally indirect circulation pattern (Blackman, 1977). Thus we might find that the frontogenetic effect of the vertical advection terms are opposite their effects on maintaining the mean baroclinic zones of this study. However, we are still restricted by the fact that the mean jet streams are situated in data sparse areas, specifically, at the eastern edges of the major continents and western portions of the oceans. Remote sensing of atmospheric variables may someday be of high enough quality to provide much needed data over the vast ocean areas.

Furthermore, previous investigations have shown that local maxima of frontal frequencies exist on the lee side of the major mountain ranges upstream of the mean jet streams. Since these regions inherently are over land masses, data spacing may not be so troublesome (particularly to the east of the Rocky mountains where
data is available for many stations and many years). The increased data should not only identify the mean frontal zones accurately, but also might provide better fields of the winds at higher levels. Thus, the frontogenetic effect of the vertical velocities should be better represented.

More localized frontal phenomena could also be investigated by the method presented in this thesis. Ninomiya (1980) has documented some processes which might be responsible for maintaining the Baiu front. This front is a summertime phenomena which is situated over eastern Asia and is responsible for much of the annual precipitation there. Further investigations might help us understand these processes even better.


