Structure of a cold front over the ocean

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SUMMARY

The kinematic and thermodynamic structure of a strong cold front over the Gulf of Alaska has been determined using aircraft, dropsonde, and radiosonde data obtained during the Storm Transfer and Response Experiment. Synoptic-scale analysis using the Sawyer–Eliassen secondary circulation equation shows that friction and diabatic heating, as well as the geostrophic forcing, are important in accounting for the secondary circulation at the front. A fine-scale analysis, with horizontal resolution of 700 m and vertical resolution of 100 m, shows strong relative inflow of warm boundary layer air toward the front from the east and a weaker inflow of cold air from the west. Updraught velocities of greater than 6 m s⁻¹ over a 2 km width were measured at 875 mb near the leading edge of the front. The frontogenous effects of confluence, turbulent mixing, and tilting of isentropic surfaces are evaluated in the region of the front and compared with the results of previous studies. The high resolution of the thermodynamic fields provides new data for comparison with high resolution numerical models.

1. INTRODUCTION

The kinematic and thermodynamic structure of surface fronts has been studied for more than sixty years, since fronts were first discussed by J. Bjerknes (1919). This subject continues to be important because the dynamics of fronts are complex and are not yet entirely understood and, of course, because so much dramatic weather is associated with surface fronts. Especially notable from a historical perspective among many observational studies throughout this long period were the serial ascent investigations by Bjerknes and Palmén (1937), which established the vertical structure of surface fronts, and the study of frontogenous processes of a lower troposphere cold front carried out by Sanders (1955).

Browning and Harrold (1970) first used Doppler radar data to produce analyses of velocity fields in the vicinity of fronts with horizontal resolution of roughly 150 m. They resolved the abrupt turning of the wind in the lowest 1 km for a cold front over land, and were able to gain new insights into the distribution of precipitation at fronts. Hobbs and Persson (1982) have used Doppler radar to describe ‘precipitation cores’ along a cold front approaching the coast of Washington state. Carbone (1982) has used Doppler radar to study the flow associated with an intense cold front over the Central Valley of California. Temperature and wind data of about 150 m resolution have been presented by Shapiro (1984) for a strong dry cold front, based on tower measurements for the lowest 300 m of the atmosphere.

Results of observational studies have provided the background on which dynamical studies of fronts and frontogenesis have been carried out by Sawyer (1956), Eliassen (1959, 1962), Hoskins and Bretherton (1972), and others, and for development and verification of numerical models of fronts by Blumen (1980), Keyser and Anthes (1982), and others.

The Storm Transfer and Response Experiment (STREX) was carried out in the autumn of 1980 in the Gulf of Alaska to study the structure of the boundary layer and the distribution of surface fluxes in ocean storms. Aircraft were used to deploy dropsondes and to fly prescribed routes within the boundary layer as nine storms passed through the region of Ocean Station Papa (50°N 145°W). In the case presented here, two aircraft, the NCAR Electra and a NOAA P-3, flew through one of the strongest fronts encountered in the six-week observation period.
The purpose of this study is to determine the kinematic and thermodynamic structure of the front with the highest resolution permitted by the basic data, and to determine the magnitude and distribution of advective processes, condensational heating, and turbulent mixing in the vicinity of the front. This study adds to the results of earlier diagnostic studies in the following respects. First, it was carried out over the open ocean in a region minimally influenced by topography and ocean surface temperature gradients. Second, the data set includes not only good synoptic-scale coverage, but also aircraft data which permit analyses of higher resolution than was possible in previous studies.

2. DATA ANALYSIS

Separate analyses were carried out on the synoptic scale and on a sub-synoptic or mesoscale. On the synoptic scale, analysis was based on twenty-five soundings and eleven observations from surface ships of opportunity. On the mesoscale, a much finer resolution analysis was based on high resolution flight-level measurements by the research aircraft.

The soundings used in the synoptic analysis include ten radiosondes released from the Canadian weathership Vancouver and the NOAA Oceanographer at three-hour intervals from 1800 GMT 15 November to 0600 GMT 16 November, eleven dropsondes deployed from an Air Force C-130 between 1815 GMT 15 November and 0144 GMT 16 November, and four dropsondes deployed from the NOAA P-3 between 2044 GMT 15 November and 0244 GMT 16 November. These soundings were examined, and spurious data were subjectively deleted. Figure 1 shows the positions of these soundings relative to the front at 0000 GMT 16 November. The grid for the objective analysis is also shown. Grid points are oriented parallel and normal to the front, with a horizontal spacing of 115 km. Objective analyses of temperature, water vapour mixing ratio, cross-frontal wind, and along-frontal wind were executed at 50 mb intervals from the surface to 500 mb, following the scheme outlined by Cressman (1959). The final scan radius was approximately one grid interval. This method may result in underestimating the sharp gradients at a front. Therefore, separate analyses were performed in the post-frontal and pre-frontal regions, using only data from within the respective region. Soundings intersecting the front were used in both analyses. The combined output from the two analyses constitute the input fields for the calculations shown in the next section. Geopotential heights at each grid point were calculated by integrating the hypsometric equation upward from the surface pressure values shown in Fig. 2(c). Ageostrophic wind components were calculated by subtracting the geostrophic components from the objectively analysed measured winds.

On the average the spacing between observations is about double the grid spacing. However, observational grid spacing is not entirely uniform, and there are regions of poor resolution. To improve the effective resolution in the cross-frontal direction parameters of interest were calculated for each grid interval and averaged in the along-frontal direction. This results in collapsing the observations onto a single plane normal to the front and improves resolution in the cross-frontal direction. Variations along the front were considered less significant.

The mesoscale analysis of the front was based primarily on four crossings of the front by the aircraft along the 49°N latitude line. The NOAA P-3 crossed the front at 300 and 950 m at 2201 GMT 15 November and 0143 GMT 16 November, respectively. The NCAR Electra crossed the front at 4500 and 2100 m at 2210 and 0025 GMT, respectively. A dropsonde released by the Air Force C-130 at 2338 GMT reached the surface at about 2358 GMT at 49°5′N 137°8′W immediately to the west of the surface front. Locations of these data are shown in Fig. 9. These were the sources of the specific data used in the
mesoscale analysis, although other data provided necessary background information. The hourly time series of surface observations from the Oceanographer as the front passed around 1900 GMT was particularly valuable. The observations were transformed to a coordinate system relative to the front at 0000 GMT 16 November.

The most critical step in mesoscale analysis was the alignment of the three lower aircraft flights as they passed through the front. Alignment based on synoptic analysis of frontal positions at the times of frontal crossing was far too inaccurate. The procedure used was to identify the maximum vertical velocity measured by the P-3 in crossing the front at 300 m, and to align the maximum vertical velocities encountered on the 950 m and 2100 m crossings directly above the 300 m maximum. This procedure was suggested by the results of Browning and Harrold (1970) and Hobbs and Persson (1982) which showed vertical orientation of the region of maximum upward velocity. In the case discussed here the best agreement between measured vertical velocities and integrated one-dimensional divergence occurred when the measured maxima were aligned vertically. The less critical alignment of the 4500 m crossing was made to conform to the 2338 GMT C-130 sounding.

All observations were projected onto a vertical cross-section oriented normal to the surface front (the x direction) and horizontal velocities resolved into cross-front (u) and along-front (v) components. It was assumed that variations in fields along the front (y direction) were negligible compared to variations normal to the front. Some of the implications of this assumption are discussed later. Spectra derived from fast-response sensing of wind velocity, temperature and humidity show substantial energy near 0·4 Hz (period 2·5 s) and near 0·04 Hz (period 25 s). For aircraft flying at 100 m s$^{-1}$, this represents concentrations of energy at horizontal scales of about 250 m and 2–3 km. In order to average out the smaller-scale (turbulent) fluctuations while preserving the 2–3 km scale, the data were block-averaged in 5-second segments. Therefore the resolution is approximately 0·5 km, with data spacing east of the front somewhat greater at 300 m and somewhat less at 950 m. The 2338 GMT dropsonde data and other soundings near the front were block-averaged over vertical segments of 12 mb.

Fields of equivalent potential temperature ($\theta_e$) and velocity components $u$, $v$ and $w$ were produced by fitting all the aircraft and dropsonde data and completing the fields by subjective analysis. At points of intersection of the dropsonde and the aircraft flights the observed air temperatures differed by less than 0·5°C, and the velocity measurements differed by less than 1 m s$^{-1}$ in the cross-front direction and 4 m s$^{-1}$ in the along-front
Figure 2(a). Surface chart for 00 GMT 14 November 1980. Surface pressures and frontal analysis prepared by R. J. Reed and S. L. Mullen for the STREX Meteorological Atlas. Dashed lines indicate 1000–700 mb thickness in metres.

Figure 2(b). As Fig. 2(a), but for 00 GMT 15 November

Figure 2(c). As Fig. 2(a), but for 00 GMT 16 November
Figure 3(a). 850 mb chart for 00 GMT 16 November 1980. Solid lines indicate geopotential height (decametres), dashed lines indicate temperature (°C) and one full barb is 5 m s⁻¹. The position of the surface front is also shown.

Figure 3(b). As Fig. 3(a), but for 700 mb.

direction. Calculations using these analyses are based on grid points with spacing of 700 m in the cross-front direction and 12.5 mb (~100 m) in the vertical. The horizontal resolution of the kinematic analysis is somewhat less than that of the Doppler radar studies (roughly 150 to 500 m). On the other hand, this data set allows analysis of the thermodynamic fields with higher resolution than the radar studies.
3. Synoptic-scale analysis

Surface pressure and 1000–700 mb thickness charts for 00 GMT 14 November, 00 GMT 15 November and 00 GMT 16 November 1980 are shown in Figs. 2(a)–(c). The surface pressure fields are based on analyses by R. J. Reed and S. L. Mullen using standard data and additional observations not included in the conventional National Meteorological Center (NMC) analyses.* The 700 mb patterns used in the thickness calculations are based on NMC analyses for the first two maps. The thickness field at 00 GMT 16 November is based on a 700 mb analysis incorporating the additional soundings near the front at 50°N. Contour spacing of thirty metres is roughly equivalent to a layer-average temperature difference of 3°C.

Figure 2(a) shows the coldest air about 5° of latitude (500 km) south of the 986 mb low centre. Along the cold front south of 45°N weak geostrophic stretching deformation is tending to increase the strength of the front. At this early stage there was still relatively cool air to the east of the front. The 500 mb pattern (not shown) shows a N–S oriented trough along about 162°W and a strong ridge along the western coast of Canada. This ridge remained nearly stationary and intensified over the next 48 hours.

Comparison of Figs. 2(a) and (b) shows that substantial development of the frontal system occurred during the 24-hour interval. The 500 mb trough has intensified, resulting in strong warm advection from the SSW ahead of the front. The low pressure centre from the previous map has filled slightly and is located in the cold air to the west of the north–south frontal system. Figure 2 also shows a new rapidly deepening low at 51°N 148°W, and two open waves along the cold front.

Very rapid development took place in the 24 hours before 00 GMT 16 November (Fig. 2(c)), when the front reached its maximum strength. The northernmost open wave shown in Fig. 2(b) deepened by 40 mb in the following 24 hours to 965 mb and moved

* Meteorological Atlas, Storm Transfer and Response Experiment, STREX Data Center, University of Washington, June 30, 1981.
northward along the front. The old parent low and the low that was at 51°N 148°W in
Fig. 2(b) are shown in Fig. 2(c) as associated with tongues of low pressure to the south
and south-west of the new low centre. The southernmost open wave shown in the previous
figure had also moved along the front to 52°N 138°W and deepened 25 mb to 981 mb.
The strongest thermal gradients are located to the south of this feature in the region of
the intensive meteorological measurements. Little temperature gradient remains in the
along-front direction.

Constant pressure height analyses at 850, 700 and 500 mb at 00 GMT 16 November
are shown in Figs. 3(a), (b), (c). The charts are based on the archived NMC charts and
incorporate the additional soundings in the region of the front at 50°N. The position of
the surface front is indicated on each of these analyses.

The 850 mb map, Fig. 3(a), shows that the front is oriented nearly vertically from
the surface to this level. Strong flow from the south existed ahead of the front, with
velocities up to 45 m s\(^{-1}\) in a swatch a few hundred kilometres wide. The wind turned
sharply in a zone about 100 km in width behind the front. Almost all the temperature
drop occurred in this zone. Weak gradients are observed in the cold air west of the
100 km zone.

Figure 3(b) shows a 700 mb pattern similar to the 850 mb pattern. The frontal zone
at 700 mb is somewhat broader and is displaced about 1° of longitude to the west of the
850 mb front. The temperature change across the frontal zone is a maximum (~12 K) at
this level.

The 500 mb chart shown in Fig. 3(c) indicates wind speeds greater than 50 m s\(^{-1}\) just
ahead of the front and less abrupt turning of the wind than at 700 mb. The isotherms are
still close to parallel to the surface front with the maximum packing found 200 km behind
the front. The relatively short trough–ridge spacing implies large positive vorticity
advection above the surface front.

Calculations based on the synoptic-scale objective analysis can be used to compare
with other fronts and with simulations of fronts. The distributions of potential temperature
(\(\theta\)) and the vertical component of relative vorticity in the cross-frontal plane for this case
are shown in Fig. 4. A relatively narrow band (~150 km) of high cyclonic vorticity is
nearly vertical from the surface to 650 mb and above 650 mb slopes upward toward the
cold air. The slope of the axis of maximum vorticity is similar to that observed by Ogura
and Liou (1980) but steeper than that associated with the weaker front of Ogura and
Portis (1982). Below 600 mb, the largest horizontal temperature gradient is found just
behind or coincident with the vorticity maximum. The frontal zone is ahead of the
vorticity maximum above 600 mb. Figure 5 shows the horizontal velocity divergence and
kinematically calculated vertical velocity. Strong convergence was limited to the lowest
150 mb with a maximum at 950 mb where \(\text{div} V = -9\times10^{-5}\text{s}^{-1}\); divergence predominated
above 600 mb in the frontal zone. The low-level convergence is capped by a small region
of divergence at 750 mb. This is consistent with the results of Ogura and Portis and with
the simulations of Keyser and Anthes (1982) and Orlanski and Ross (1984).

4. SYNOPTIC-SCALE FRONTOGENESIS

The relation of kinematics to dynamics of the front can be analysed using the
theory developed by Sawyer (1956) and Eliassen (1959, 1962). Their theory relates the
ageostrophic (or secondary) circulation normal to the front to the frontogenesis due to
ageostrophic deformation. As frontogenetical processes increase the temperature gradient
across the front, the vertical shear in the along-front wind must increase to maintain
thermal wind balance. The accelerations in the along-front wind are accompanied by
ageostrophic motions in the cross-front plane. Two assumptions are included in this theory. First, the geostrophic momentum approximation is used, i.e. the acceleration of the ageostrophic wind is neglected but advection by the ageostrophic wind is included. Also, it is assumed that the wind component along a two-dimensional front is in geostrophic balance. Eliassen (1959) extended the theory by including the effects of turbulent stress and diabatic heating on the secondary circulation. Hoskins and Draghi (1977) generalized the theory of the secondary circulation to three dimensions. This theory, known as Q vector theory, has been applied to a variety of synoptic situations. Figures 2(c) and 3(a)–3(c) show that this front was virtually two-dimensional; therefore analysis of the ageostrophic flow is restricted to the cross-frontal plane. When terms representing effects of stress and diabatic heating are included, the Q vector theory reduces to the secondary circulation equation for a two-dimensional front in the form

$$\frac{\partial \theta}{\partial \omega} - \frac{2}{\gamma} \frac{\partial m}{\partial x} \frac{\partial u_x}{\partial x} + \frac{1}{\gamma} \frac{\partial m}{\partial x} \frac{\partial u_x}{\partial x} + \frac{1}{\gamma} \frac{\partial m}{\partial y} \frac{\partial u_x}{\partial y} + \frac{1}{\gamma} \frac{\partial m}{\partial z} \frac{\partial u_x}{\partial z} = -2 \frac{\partial U}{\partial x} \frac{\partial \theta}{\partial x} - 2 \frac{\partial V}{\partial y} \frac{\partial \theta}{\partial y} + \frac{1}{\gamma} \frac{\partial}{\partial x} \left( \frac{1}{\rho} \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial x} \left( \frac{d \theta}{dt} \right).$$

(1)

This equation is similar to Eq. (39) derived by Shapiro (1981). In Eq. (1) $\gamma = (RfP_0)/(P_0/P)^{\varepsilon/\varepsilon_f}$, $m = V + f x$, $u_x$ is the ageostrophic cross-frontal wind component, $\omega$ is the vertical velocity in pressure coordinates and $U, V$ are the geostrophic cross-frontal and along-frontal wind components, respectively. The left-hand side of Eq. (1) represents the sum of terms involving ageostrophic motions, and the right represents the sum of frontogenetical processes forcing the ageostrophic motions. The first two terms on the right represent frontogenesis due to the cross-front and along-front geostrophic wind, respectively. The third term on the right of Eq. (1) represents the effect of friction on the along-front vertical wind shear. The fourth term represents the frontogenesis due to the cross-front gradients in diabatic heating. By evaluating the terms in Eq. (1) the separate physical processes that may be important at a front can be assessed and compared.

The sum of the three terms which constitute the secondary circulation (left-hand side of Eq. (1)) is shown in Fig. 6. The general sense of the ageostrophic flow is counterclockwise around a positive centre and vice versa. In general, none of the three terms on the left of Eq. (1) can be neglected. However, the maximum of $1 \times 10^{-5}$ K m$^{-1}$ s$^{-1}$, located in the frontal zone at 700 mb, results largely from the second of the three terms,
\[-(2/\gamma)(\partial m/\partial p)(\partial u_a/\partial x)\]. The minimum of \(-1 \times 10^{-8} \text{K m}^{-1} \text{s}^{-1}\) ahead of the front at 900 mb results almost wholly from the third term, \((1/\gamma)(\partial m/\partial x)(\partial u_a/\partial p)\).

The geostrophic forcing, the sum of the first two terms on the right of Eq. (1), is shown in Fig. 7. The correspondence between Figs. 6 and 7 indicates the degree to which the secondary circulation can be attributed to geostrophic forcing. Figures 6 and 7 show positive centres located in the same region; the central magnitude in Fig. 6 is significantly larger than in Fig. 7. There is no negative centre in Fig. 7 corresponding to the negative centre in Fig. 6. The absence of strong geostrophic forcing near the surface is not surprising. As shown in Fig. 2(c), the isotherms became virtually parallel to the front at the mature stage of its development. Since both \(\partial U/\partial x\) and \(\partial \theta/\partial y\) are small, the total geostrophic forcing is weak.

The effect of turbulent stresses on the circulation has been estimated from the results found by Fleagle and Nuss (1985) for this case. They calculated surface stresses and sensible and latent heat fluxes during STREX using stability-corrected bulk aerodynamic formulae. Their values of surface stress along a line normal to the front at 49°N have been separated into contributions due to the along-front and cross-front wind. As shown in Eq. (1), the forcing is proportional to the vertical derivative of the along-front stress divergence. We assumed a linear stress profile from the surface value to zero at 900 mb and evaluated the stress divergence using a 50 mb vertical grid interval.

An estimate of the cross-front gradient in diabatic heating (last term in Eq. (1)) was made considering two processes: turbulent flux of sensible heat in the boundary layer and condensation or evaporation. Values of vertical heat flux at the ocean surface were taken from the results of Fleagle and Nuss, and it was assumed that these decreased linearly to zero at 900 mb. The heating due to the latent heat of condensation was determined from the change in mixing ratio following a parcel. For this computation we assumed that the mixing ratio was steady-state in a reference frame moving with the front. All of the condensate was assumed to be precipitated at the rate it was formed. Sensible heat flux at the surface of droplets (or ice crystals) was not included.

Figure 8 shows the total forcing due to the geostrophic, friction and diabatic terms of Eq. (1). The friction term is dominant at the low-level negative maximum at 900 mb in Fig. 8, accounting for about 80% of the total shown in Fig. 8. This implies that friction is principally responsible for the concentrated convergence and the vertical jet below 850 mb. Heating due to boundary layer turbulent heat flux divergence constituted a minor effect here. Condensational heating and cooling above the boundary layer accounts for

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**Figure 6.** Distribution of the sum of the cross-front ageostrophic circulation terms, \((\partial \theta/\partial p)(\partial u_a/\partial x) - (2/\gamma)(\partial m/\partial p)(\partial u_a/\partial x) + (1/\gamma)(\partial m/\partial x)(\partial u_a/\partial p)\) \((10^{-10}\text{K m}^{-1} \text{s}^{-1})\). The dash-dotted line represents the axis of maximum relative vorticity.

**Figure 7.** Distribution of the geostrophic forcing terms, \((-2\partial U/\partial x)(\partial \theta/\partial x) - 2\partial V/\partial x)(\partial \theta/\partial y)\) \((10^{-10}\text{K m}^{-1} \text{s}^{-1})\). The dash-dotted line represents the axis of maximum relative vorticity.
the greater value of the positive centre at 700 mb in Fig. 8 over the corresponding centre in Fig. 7. This is associated with a strong horizontal gradient of vertical motion.

Baldwin et al. (1984) found similar results in a diagnostic study of a two-dimensional model simulation of frontogenesis. They also used the Sawyer-Eliassen equation to examine the processes important to the forcing of the secondary circulation. Their study showed the importance of friction to the forcing of a vertical velocity maximum at low levels. Latent heating served to strengthen the thermally direct circulation cell in the frontal zone at mid levels. Turbulent heat exchange had a small effect in their model, which used a thermally insulated lower boundary. Our computations that included surface heat fluxes also show that boundary layer turbulent heat fluxes had a minimal effect on the synoptic-scale secondary circulation.

There remain substantial differences between the sum of the circulation terms (Fig. 6) and the total forcing (Fig. 8). Most prominent is the 40% underestimate of the intensity of the negative maximum in circulation at 900 mb. This may be due to a variety of factors including observation errors, inadequate data coverage, uncertainty in the frictional and diabatic heating parametrizations, and finite difference errors. The steady-state, two-dimensional, and geostrophic momentum assumptions are not perfectly valid, and the along-front wind may not be in complete geostrophic balance, especially in the boundary layer. We have not concluded which of these sources of error are likely to be most important. Above 600 mb the differences between Figs. 6 and 8 are probably associated with omitted terms expressing gradients parallel to the front. The most important point, however, is that Figs. 6 and 8 agree reasonably well, and this indicates that the dominant features of the synoptic-scale ageostrophic circulation at a strong front can be accounted for by the geostrophic forcing, plus the diabatic and frictional effects.

5. MESOSCALE STRUCTURE

Flight-level pressure measurements within a few kilometres of the front show significant variations from the smooth large-scale field of Fig. 2(c). These pressure perturbations appear to be non-hydrostatic since they are generally correlated with the large vertical accelerations shown in later figures. The pressure measurements show that the along-front wind was strongly ageostrophic near the front and that accelerations in the cross-frontal ageostrophic wind were large. We conclude that the approximations
required for Eq. (1) are not valid on the mesoscale in the immediate vicinity of the front. A more general equation containing inertia-gravity waves, convection cells, secondary flows in the boundary layer, and other mesoscale phenomena is too complicated to provide the useful insights of Eq. (1). It remains for future research to elucidate the dynamical effects of these phenomena on fronts. The following analysis, therefore, is addressed to a kinematical and thermodynamical description rather than to a dynamical equation.

Figure 9 shows the locations of the averaged flight-level data and dropsonde data, and the analysed field of $\theta_e$. The heavy solid lines in Fig. 9 and all subsequent figures represent the 307 K contour. The zero point on the horizontal scale was arbitrarily set at the intersection of the 307 K contour and the surface. Relative humidity below about 700 mb is greater than 95% except for one small region behind the front. Because the cross-section is so nearly saturated, $\theta_e$ is used to represent a quasi-conservative thermodynamic variable. The backward slope of the front below 950 mb is based on the sounding that reached the surface immediately to the west of the front. The strongest horizontal gradients of potential temperature ($\theta$) and $\theta_e$ were at 900 mb. Here, aircraft data recorded at one-second intervals show a maximum air temperature gradient of 1.6°C/(100 m)

At 955 mb the maximum horizontal air temperature gradient was 0.8°C/(100 m)

By comparison, Shapiro (1984) found a maximum potential temperature gradient of about 3°C/(100 m)

in the lowest 300 m of a dry cold front over land. The general pattern of $\theta_e$ suggests a gravity current (also called density current) as illustrated, for example, by Simpson (1972). The undulations behind the 'nose' at 890 mb and the low-level convective instability are features in common with a gravity current. Benjamin (1968) has shown that the speed of a gravity current can be expressed by

$$u' = \left( g \Delta z (T_{v1} - T_{v2}) / T_{v2} \right)^{1/2}$$

where $\kappa^2$ is the internal Froude number, $g$ is the acceleration of gravity, $\Delta z$ is the depth of the cold air, and $T_{v1}$ and $T_{v2}$ are the depth-averaged virtual temperatures in the warm and cold air, respectively. For the reported range of $\kappa$ (1.1 to 1.4), $\Delta z$ of 1.2 km, and

![Image of Figure 9](image_url)
$T_v$ and $T_e$ of 284 K and 280 K, respectively, Eq. (2) yields $u'$ ranging from 14 to 18 m s$^{-1}$, whereas the measured speed was 17 m s$^{-1}$.

Figure 10 shows the field of the wind component normal to the front ($u$) in a coordinate system moving with the front. As found by Carbone (1982) and Hobbs and Persson (1982), the greatest relative inflow toward the front was at low levels from the warm side. A vertically oriented axis of maximum two-dimensional convergence of magnitude $6 \times 10^{-3}$ s$^{-1}$ was present in the warm air below about 925 mb. Above this level to about 890 mb the convergence was of the same magnitude with an axis tilted back toward the cold air. The convergence dropped off rapidly with height above the nose of the front. Divergence of order $2 \times 10^{-5}$ s$^{-1}$ was found between 850 and 750 mb directly above the surface convergence. This distribution is qualitatively consistent with the synoptic-scale divergence pattern shown in Fig. 5 and with results of the other mesoscale studies referred to earlier.

Figure 11 shows the along-front wind component. The maximum velocities were analysed as greater than 40 m s$^{-1}$ in the warm air near 850 mb. The velocities were lower near the surface, possibly due to friction, and dropped off weakly with height above 850 mb. This front appears to have had a less prominent prefrontal low-level jet than those cases studied by Browning and Harrold (1970), Browning and Pardoe (1973) and Hobbs and Persson (1982). Relatively low velocities ($\sim 30$ m s$^{-1}$) are shown above and to the left of the nose of the front. The analysis presented below shows that air in this region had been advected from near the surface on the warm side of the front by the main frontal updraught. This implies that momentum in the along-front direction was quasi-conserved in the warm air. The one-dimensional relative vorticity ($\partial v/\partial x$) reaches a maximum of $1.2 \times 10^{-3}$ s$^{-1}$ at 950 mb in the same region as the maximum in convergence. Hobbs and Persson found a relative vorticity of $0.8 \times 10^{-3}$ s$^{-1}$ at a precipitation core, and in the two cases studied by Browning and Harrold maximum vorticities of $0.6 \times 10^{-3}$ s$^{-1}$ were found.

The vertical velocity ($w$) in the cross-front plane is shown in Fig. 12. This field is based on the direct measurements of $w$ made by the research aircraft. The maximum updraught velocity found here, 6 m s$^{-1}$ at 875 mb, can be compared to the vertical velocities of 8, 7 and 17 m s$^{-1}$ found in the studies of Browning and Harrold (1970), Hobbs and Persson (1982) and Carbone (1982), respectively. The updraught in this case was somewhat wider than in the earlier cases. A downdraught of 1 m s$^{-1}$ was found 2 km behind the main updraught. This feature is present in the other observational studies.
The most difficult problem in analysis arose from the fact that the integrated one-dimensional divergence near the front underestimates the directly measured maximum vertical velocity by about 35% and the vertical mass flux over a width of 5 km by 50%. It appears plausible that most of the discrepancy must reflect either: (1) change in the velocity field between successive crossings of the front by the P-3 at 300 m and 950 m; or (2) mesoscale variation in the parallel wind component in the along-front direction. Regarding the first of these possibilities, the synoptic analyses show that the pressure gradients on both sides of the front became more intense in the 12 hours preceding 00 GMT 16 November. Therefore the 300 m value of $\partial u / \partial x$ and the corresponding vertical velocities may have been greater at 0143 GMT than they were when measured at 2201 GMT. In examining the effect of variation along the front we considered the effects of precipitation cores which have been described by James and Browning (1979) and Hobbs and Persson (1982). These precipitation cores, which were associated with large low-level convergence at their leading edges, alternated with gap regions of smaller convergence. The two low-level crossings by the P-3 may have intersected the front at different phases with respect to precipitation core and gap regions. This could account for the discrepancy between the measured vertical velocity and vertically integrated divergence fields. Lacking radar data, it is not possible to verify the existence of precipitation cores or to determine mesoscale variations in the along-front direction. Therefore, our data set does not provide a basis for distinguishing between the effects of variation of divergence occurring in time or in the along-front direction.

The kinematic and thermodynamic structure of the front is shown in a single diagram in Fig. 13. Here the $\theta_e$ field shown in Fig. 9 is repeated, together with the computed streamlines relative to the front (equivalent to trajectories projected onto the vertical plane under the assumptions of steady state and independence of $y$). Calculations have been carried out as described earlier, and results have been faithfully represented in the figure. In this coordinate system which moves with the front the $\theta_e$ field is fixed while the ocean boundary moves to the left. Within the boundary layer air flows toward the front from both the cold and warm sides; the warm air approaches the front more rapidly than the cold air as shown by the 100-second intervals marked along the trajectories.
The surface winds observed at the *Oceanographer* following the passage of the front and the 300 m aircraft measurements show that the relative flow in the cold air was toward the front.

The converging cold and warm streams form an updraught within and east of the region of greatest horizontal gradient of $\theta_e$, with the strongest part of the updraught occurring in the warm air. The updraught is about 5 km in width and transports air from the boundary layer to the free atmosphere at a rate of $20 \times 10^3$ kg s$^{-1}$ per metre measured parallel to the front. The mass of water vapour carried upward by the updraught amounts to 160 kg s$^{-1}$ m$^{-1}$, equivalent to precipitation of 1 mm h$^{-1}$ extending over a west–east span of 8° longitude (580 km).

It is interesting to estimate the contribution of frictional convergence to the total mass flux of the updraught. Where curl of the surface stress is large and advection and local change can be neglected, the vorticity equation yields the following expression for the vertical velocity at the top of the boundary layer

$$\omega_i = (1/\rho f) \nabla \times \tau \cdot \mathbf{k}$$  \hspace{1cm} (3)

where $\mathbf{k}$ is the unit vertical vector, $\tau$ represents surface stress, $\rho$ is density, and $f$ is the Coriolis parameter. Surface stress for this case has been computed using the bulk aerodynamic formula,

$$\tau = C_d \rho |\mathbf{V}| \mathbf{V}$$  \hspace{1cm} (4)

where $C_d$ represents the drag coefficient. For neutral stability Large and Pond (1982) have expressed the drag coefficient by

$$C_d = (0.49 + 0.065|\mathbf{V}|) \times 10^{-3}$$

and this has been used for the following calculation, ignoring the relatively small effect.
of static stability of the surface layer. The surface stress in the along-front direction was computed at two points, 0 and +5 km along the x axis of Fig. 13. The main updraught and nearly all the along-front wind shear is contained within this interval. The difference between the two stress values yields a vertical velocity of 2.8 m s\(^{-1}\), which is 80% of the measured average vertical velocity over the 5 km width at 900 mb. This result indicates that frictional convergence accounted for nearly the entire updraught mass flux, the same conclusion reached by Browning and Harrold (1970) for two cold fronts over land.

Above 875 mb the updraught undergoes strong horizontal divergence with part of the warm air flowing eastward and continuing slowly upward, while another part flows westward along the sloping frontal surface. The vertical oscillations in and above the frontal zone resemble gravity waves generated by Kelvin–Helmholtz instability. The oscillations can be recognized as the visible part of the process of entrainment of relatively warm air from above the frontal zone. Much of the entrainment, of course, occurs at scales which are not resolvable in detail by this study.

Two vortices are shown in the cold air behind the front. The two-dimensional streamlines appear to terminate at the centre of these regions. This implies that there was some divergence in the along-front direction. The isolated small pocket of relatively cold air (\(\theta_c < 301\) K) between the two vortices cannot be accounted for by advection in the plane of the figure. It must be a result of diabatic effects and/or variations in \(\theta_c\) in the along-frontal direction.

The kinematic structure of this over-ocean front was similar in most respects to the fronts described earlier using Doppler radar. The more uniform lower boundary of this study appears not to have had unique or major influences on the kinematics of the front. The vertical oscillation above the leading edge of the front was found in each of the earlier studies. The double vortex in the cold air is also present in Carbone's (1982) results.

The thermodynamic structure of the lowest few hundred metres of this front is similar to the front analysed by Shapiro (1984). Figure 13 indicates that in the converging boundary layer flow convective instability was stronger on the cold side than on the warm side. The fields of streamlines relative to the front and of equivalent potential temperature are similar in their major features; this helps to confirm such features as the boundary layer flow, the strong updraught in the warm air ahead of the front, and horizontal divergence in the updraught above 875 mb with separation of the updraught into a current flowing eastward and a current flowing westward relative to the sloping front.

In the coordinate system relative to the front the rate of change of \(\theta_c\) experienced by individual air parcels can be expressed by

\[
\frac{d\theta_c}{dt} = \partial \theta_c/\partial t + u \partial \theta_c/\partial x + v \partial \theta_c/\partial y + w \partial \theta_c/\partial z. \tag{5}
\]

In order to compare the magnitudes of these terms we consider the idealized situation in which condensation and evaporation are the only diabatic processes (\(d\theta_c/dt = 0\)), steady-state exists with respect to the moving front (\(\partial \theta_c/\partial t = 0\)), and there is no variation along the front (\(\partial \theta_c/\partial y = 0\)). Under these conditions the two-dimensional velocity in Fig. 13 is exactly normal to the gradient of \(\theta_c\), that is the streamlines are everywhere parallel to \(\theta_c\) isotherms. It is clear from Fig. 13 that streamlines and isotherms tend to be parallel, especially in regions of high velocity or large gradient; in these regions \(u \partial \theta_c/\partial x + w \partial \theta_c/\partial z \approx 0\). The absolute magnitudes of each of these terms reaches about 6 K (100 s\(^{-1}\)) at the leading edge of the front at 890 mb. Figure 13 also indicates regions where streamlines and isotherms are not parallel. In these regions one or more of the omitted terms in Eq. (5) may be important. It is possible to identify the physical processes that may be important in these regions. In the region of the vertical oscillation above the
leading edge of the cold air the streamlines and $\theta_e$ isotherms are nearly normal to one another. As shown in Fig. 13, this region is just below the freezing level. A possible explanation proposed by Carbone (1982) is that the air in this region may be cooled by melting of ice crystals falling through it, so that $\theta_e$ is not conserved following the moving air. Then the magnitude of $d\theta_e/dt$ shown in Fig. 13 is as large as roughly $-1 \, \text{K} \, (100 \, \text{s})^{-1}$. The streamlines and isotherms also tend to cross in the region of the two vortices behind the front. Significant processes in this region may include turbulent mixing, changes with respect to time, or variations in the $y$ direction.

Turbulent fluctuations superimposed on the mean flow shown in Fig. 13 may bring air parcels of varying $\theta_e$ together and, by mixing, result in diabatic change in $\theta_e$ following the mean streamlines. The rate of diabatic change can be expressed by the divergence of the turbulent flux. Considering first the vertical component of the turbulent flux, the relation can be represented by

$$
(d\theta_e/dt)_{mz} = \dot{\partial} \frac{(\overline{w'}\theta_e')}{\partial z} = \frac{\partial}{\partial z} \left( K_z \frac{\overline{\partial \theta_e}}{\partial z} \right)
$$

(6)

where the subscript `m' designates the process of mixing, the overbar represents a mean value, the primes designate fluctuations from the mean, and $K_z$ represents the vertical turbulent transfer coefficient or eddy diffusivity. Similar equations may represent diabatic change associated with flux in the horizontal directions. Fluxes were calculated from the covariance of fluctuations measured by the aircraft. However, long averaging times and homogeneous conditions are necessary to calculate accurate values for the turbulent fluxes. Since these conditions were not fully met, we have represented flux using first-order closure or $K$ theory, and have assumed that the turbulent transfer coefficient or eddy diffusivity can be approximated as a constant. $K_z$ was estimated as the average of the ratio of $w'\theta_e'$ measured by the P-3 to the vertical gradient of $\theta_e$. This average was computed over approximately 15 minutes of flight time in the immediate vicinity of the front but not within the frontal zone itself. Fluxes were computed in 30-second blocks. Linear trends in the dependent variables were removed from each block before correlating fluctuations of temperature and humidity with those of vertical and cross-frontal velocities. Vertical gradients were calculated using increments of 25 mb.

Using an equation analogous to Eq. (6), similar calculations were made to determine values for the turbulent transfer coefficient in the $x$ direction using horizontal increments of 1-4 km.

The dependence of the turbulent transfer coefficients on the mean fields was investigated using Eq. (2.10) from Ross and Orlanski (1982), which relates the transfer coefficients to the bulk Richardson number ($Ri$). Calculations for this case assuming saturation showed that the vertical eddy diffusivity varied from 150 to 60 m$^2$s$^{-1}$ between regions of low and high $Ri$ ($<0$ to $>10$). The horizontal transfer coefficient was not dependent on $Ri$ but tended to be larger in regions of large horizontal wind shear. Using $K$s determined from the Ross and Orlanski equation, the diabatic heating rates were calculated. Results did not differ significantly from results using constant coefficients. Therefore, constant values of 100 m$^2$s$^{-1}$ and 600 m$^2$s$^{-1}$ for the vertical and horizontal transfer coefficients have been used for the calculations that follow.

The divergence of the vertical and horizontal fluxes yields the 'heating rate' due to turbulent mixing. At the resolution of this study, horizontal diffusion was less important than vertical diffusion by about a factor of four. Magnitudes of $d\theta_e/dt$ due to turbulent mixing ranged from cooling of $0.3 \, \text{K} \, (100 \, \text{s})^{-1}$ to warming of $0.6 \, \text{K} \, (100 \, \text{s})^{-1}$. In general, positive (negative) changes in $\theta_e$ following a parcel occurred in regions of increasing (decreasing) tendency in $\theta_e$ due to turbulent mixing. However, the turbulent mixing was
not strong enough to account for all of the Lagrangian change in $\theta_e$, especially in the regions of greatest gradients of $\theta_e$. Within the frontal zone itself, where aircraft turbulence data were not analysed, the eddy diffusivities may well have been much larger than those measured. A distribution of the residual, the difference between $d\theta_e/\!\! dt$ and $u \partial \theta_e/\!\! \partial x + w \partial \theta_e/\!\! \partial z$, reached magnitudes of \(1 \text{K}(100 \text{s})^{-1}\), approximately 1/6 of the maximum values of the dominant terms ($u \partial \theta_e/\!\! \partial x$ and $w \partial \theta_e/\!\! \partial z$). The residual includes the effects of radiation, cooling by melting ice, warming or cooling by falling rain, and errors of observation and analysis, as well as the explicit terms, $\partial \theta_e/\!\! t + v \partial \theta_e/\!\! y$.

The most important point is that with the exceptions noted the fields of relative velocity and equivalent potential temperature are similar in their major features: each of the fields provides confirmation of the other. It may be inferred that in the boundary layer turbulent mixing substantially affects the thermodynamic fields, but that above the boundary layer substantial diabatic effects other than evaporation and condensation occur only in limited regions. Although the effects of change with time and variation along the front have not been explicitly accounted for, they do not appear to have invalidated this analysis.

6. MESOSCALE FRONTOGENESIS

Sanders (1955), following Miller (1948) and Petterssen (1936), has expressed frontogenesis for the case of uniformity along the front by

$$\frac{d}{dt}(\partial \theta_e/\!\! \partial x) = \frac{\partial}{\partial x}(d\theta_e/\!\! dt) - (\partial u/\!\! \partial x)(\partial \theta_e/\!\! \partial x) - (\partial w/\!\! \partial x)(\partial \theta_e/\!\! \partial z)$$

(7)

where $d/\!\! dt$ signifies total differentiation in the Lagrangian sense, and the $x$ axis is directed perpendicular to the front toward the warm air. Variations in the along-front direction are neglected. This equation expresses frontogenesis as the sum of terms representing the effects of diabatic processes other than condensation and evaporation, confluence, and tilting of the $\theta_e$ surfaces. It describes the change experienced by a parcel and does not yield an estimate of the local change in the strength of the front.

The first term on the right of Eq. (7) represents all diabatic effects on $\theta_e$; in the following we have considered only the effects of turbulent mixing. The eddy flux divergences discussed in the previous section were differentiated in the cross-frontal direction, yielding a time rate of change in the horizontal gradient of $\theta_e$. In Fig. 14 the combined effects of vertical and horizontal diffusion are shown with positive values indicating frontogenesis. Following the notation of Eq. (6), the subscript in the caption designates the heating rate due to mixing. In general, the areas of frontogenesis and frontolysis alternate in the horizontal, and the pattern is 'noisy', reflecting the fact that it requires third-order derivatives. The largest positive and negative values were found near the 900 mb level in the region of largest gradients in the $\theta_e$ field. The largest absolute values were about $8 \times 10^{-6}$ K m$^{-1}$ s$^{-1}$, and typical values ranged from $1 \times 10^{-5}$ K m$^{-1}$ s$^{-1}$ at low levels to $0.3 \times 10^{-5}$ K m$^{-1}$ s$^{-1}$ above 850 mb.

The second term on the right of Eq. (7), the confluence term, is shown in Fig. 15. In the frontal zone at 900 mb positive values as large as $25 \times 10^{-6}$ K m$^{-1}$ s$^{-1}$ occurred. In the boundary layer below 900 mb maximum wind convergence was located on the warm side of the largest temperature gradient, thus tending to broaden the zone of significant confluence.

The third term on the right of Eq. (7), the tilting term, is shown in Fig. 16; it can be visualized as the change in horizontal gradient produced by differential vertical advection of the equivalent isentropic surfaces. Where the air is stable to moist processes,
this term is negative on the left or cold side of the updraught core and positive on the right. At lower levels, where the atmosphere was convectively unstable, the sign is reversed. As a result there was a region of frontogenesis due to tilting of magnitude greater than $10\times10^{-5}\text{K}\text{m}^{-1}\text{s}^{-1}$ near 950 mb in the frontal zone. Shapiro (1984) also found a frontogenetical tilting effect in the convectively unstable region of the frontal zone. An extreme value of frontolysis of approximately $90\times10^{-6}\text{K}\text{m}^{-1}\text{s}^{-1}$ occurs in the region of the very stably stratified frontal inversion at 875 mb. The tilting effect would have had a different distribution if potential temperature ($\theta$) rather than $\theta_e$ were plotted. As shown by Sanders (1955), a large frontogenetical effect would have been indicated in the warm air to the right of the main updraught. This effect is fictitious since the adiabatic cooling calculated in the updraught was actually compensated by warming due to latent heat release. When conditions are so nearly saturated, $\theta_e$ is a more appropriate thermodynamic variable.

The sum of the frontogenetical processes is shown in Fig. 17. The region of the strong updraught is of greatest interest. In the converging flow in the boundary layer the confluence and tilting terms contributed more or less equally to strong frontogenesis. Parcels experiencing the greatest frontogenesis achieve the strongest horizontal gradients in $\theta_e$ in the upper boundary layer near 900 mb. Above 900 mb the very large tilting term
is responsible for strong frontolysis. In visualizing the processes discussed here, it should be recognized that the numbers refer to rates of frontogenesis following the motion, and that these numbers can be multiplied by time in a region to yield change in horizontal gradient experienced by individual parcels within that region.

To summarize, the processes responsible for frontogenesis and frontolysis have been analysed based on nominal space scales of 700 m in the horizontal and 100 m in the vertical. These results provide another description of the coupled kinematics and thermodynamics of the front. Turbulent mixing is the only strongly frontolytic process in the boundary layer near the front, suggesting that boundary layer mixing may be the primary process which limits the sharpness of the front.

To compare this case to those analysed by Sanders (1955), Ogura and Portis (1982) and Shapiro (1984), and the model results of Keyser and Anthes (1982) and Baldwin et al. (1984), we have also calculated the confluence and tilting effects for potential temperature on a 50 km horizontal scale. The diabatic mixing term is negligible at this scale. The peak values are summarized in Table 1. The values for the study by Keyser and Anthes are derived from their high-resolution PBL model run, and crudely approximated from their Fig. 14.

<table>
<thead>
<tr>
<th>Study</th>
<th>$\frac{\partial u \partial \theta}{\partial x \partial x}$</th>
<th>$\frac{\partial w \partial \theta}{\partial x \partial z}$</th>
<th>Horizontal scale</th>
</tr>
</thead>
<tbody>
<tr>
<td>Baldwin et al. (1984)</td>
<td>$5 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>$-8 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>440 km</td>
</tr>
<tr>
<td>Ogura and Portis (1982)</td>
<td>$2 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>$-2 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>90 km</td>
</tr>
<tr>
<td>Bond and Fleagle</td>
<td>$5 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>$-4 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>50 km</td>
</tr>
<tr>
<td>Keyser and Anthes (1982)</td>
<td>$5 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>$-5 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>40 km</td>
</tr>
<tr>
<td>Sanders (1955)</td>
<td>$5 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>$-4 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>&lt;25 km</td>
</tr>
<tr>
<td>Bond and Fleagle</td>
<td>$2.5 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$ *</td>
<td>$-9 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$ *</td>
<td>700 m</td>
</tr>
<tr>
<td>Shapiro (1984)</td>
<td>$8 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>$-8 \times 10^{-6} \text{K m}^{-1} \text{s}^{-1}$</td>
<td>200 m</td>
</tr>
</tbody>
</table>

* These confluence and tilting terms are calculated for $\theta$.

On comparable scales the front analysed here was stronger than the fronts analysed by Ogura and Portis and those simulated by Keyser and Anthes. The front analysed by Sanders was characterized by stronger temperature gradients and weaker convergence than was this front when analysed on the 25 km scale. The front studied by Shapiro was somewhat stronger than this front; his values of convergence and temperature gradient on a 700 m horizontal scale are about 50% greater than those of this case.

7. CONCLUSIONS

Synoptic-scale analysis has shown that this mid-ocean front was similar to other observed fronts and those simulated by model calculations. Analysis based on the Sawyer–Eliassen secondary circulation equation reveals the importance of friction and condensational heating as well as geostrophic forcing in creating the ageostrophic circulation near the front. Friction was especially important in the boundary layer, where forcing by the geostrophic wind was weak.

Mesoscale analysis shows that the vertical transport of air and water vapour was highly concentrated within a zone about 2 km wide at the leading edge of the front.
Gradients of temperature and humidity and associated frontogenetical processes were also concentrated near the leading edge of the front. Frictional convergence in the boundary layer accounted for 80% of the observed updraught. The analysis disclosed frontal features similar to gravity currents in the laboratory, extension of the cold, dense air ahead of the cold air at the ocean surface, and vertical undulations of the frontal surface behind the surface front.

The low-level frontal zone, as defined by the region of wind shift, was about 2 km wide. This is close to the width observed by Browning and Harrold (1970), Carbone (1982) and Hobbs and Persson (1982) at other precipitating cold fronts. On the other hand, the dry cold fronts studied by Shapiro (1984) and Young and Johnson (1984) had frontal zone widths of about 200 and 500 m, respectively. Frontal widths near the surface are largely determined by the combined effect of confluent flow which tends to increase gradients, and turbulent mixing which tends to reduce gradients. Tilting can be a modifying effect on this relationship above the surface layer. Further study is necessary to examine the influences of stability, horizontal shear, and other aspects of the flow on turbulent mixing at a front.

Results of this study emphasize the importance of accurately representing boundary layer processes in studies of fronts using numerical models. The results presented here should contribute to the design and evaluation of such models.

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