Polar Lows over the Gulf of Alaska in Conditions of Reverse Shear

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ABSTRACT

The formation of two polar lows over the Gulf of Alaska are studied, using observations taken during the OCEAN STORMS field experiment. Synoptic-scale and mesoscale analyses were constructed using NOAA P-3 aircraft flight-level and radar data and dropwindsonde profiles, in addition to conventional data sources. The synoptic-scale analyses show that polar low development occurred in a low-level mesoscale baroclinic zone near the center of a mature, occluded synoptic-scale low pressure system. The disturbances propagated along the zone in the reverse shear sense, i.e., in the direction opposite to the thermal wind within the zone. The mesoscale analysis reveals a boundary layer jet of 38 m s⁻¹ on the cold side of the baroclinic zone accompanying the leading polar low. Indirect evidence suggests that a Sawyer–Eliassen secondary circulation was present; polar low development occurred in the region of a frontogenetical geostrophic deformation. Convective activity was not prominent during the growth phase of the polar lows, as determined from satellite imagery and radar reflectivity measurements. Extremely high ocean waves (~13 m) occurred in response to intense wind forcing on the mesoscale on a sea state preconditioned by moderate forcing on the synoptic scale. The observed synoptic-scale and mesoscale structures are compared with results from previously studied polar lows. This case appears to represent an example of polar low development due primarily to moist baroclinic processes.

1. Introduction

The physical processes responsible for the development of subsynoptic-scale (<500 km) vortices in polar airstreams, usually referred to as polar lows, have become a subject of increasing research interest. Polar lows are defined as subsynoptic-scale cyclones that develop poleward of the major polar front and jet stream. As reviewed by Businger and Reed (1989), interpretations differ regarding the role of baroclinic instability relative to mesoconvective processes in the lifecycles of polar lows. Because few polar lows have been observed to develop over land, it is generally agreed that air–sea interaction processes are important to their development. The debate focuses on whether latent heating is important by reducing the effective static stability and augmenting the baroclinicity of fronts (Eliassen 1959), thereby enhancing the baroclinic growth rate, or whether localized mesoconvective heating induces the vortex spin up, i.e., as in a hurricane.

Observational evidence for identifying a single key process is varied; e.g., Harrold and Browning (1969) suggested that polar lows were a manifestation of low-level baroclinity within the polar air, whereas Rasmussen (1979) emphasized the importance of mesoconvective heating through the conditional instability of the second kind (CISK) mechanism (Charney and Eliassen 1964). More recently, Forbes and Lottes (1985) composited the climatology of Atlantic polar low cases and suggested that both baroclinic and CISK mechanisms were important. In addition, they found relatively subtle differences in the synoptic environment separating developing and nondeveloping lows. Shapiro et al. (1987) were the first to use research aircraft measurements to study a polar low, and they attributed the initiation of the low to synoptic-scale baroclinity. Their detailed analysis featured observations of vigorous cumulus convection along a low-level baroclinic zone within the circulation of the low. These observations were collected during the mature phase of the storm, and it was not possible to determine to what extent the mesoscale baroclinity or convection were consequences of the development.

Theoretical and numerical studies have addressed the mechanisms responsible for polar low development. Mansfield (1974) and Duncan (1977) showed that a shallow baroclinic layer with low static stability could support development over realistically short time and length scales. Sardie and Warner (1985) used a prim-
itive equation numerical model to simulate an Atlantic and a Pacific polar low. They found that baroclinity and a convective precipitation parameterization were necessary for a proper simulation in both cases. Mudrick (1987) found that dry baroclinic instability could explain the initial stages of polar low development. Nordeng (1987) used an operational mesoscale numerical model initialized with conventional data in a study of two polar lows. He found that a parameterization of slantwise convection yielded stronger low-level winds for both simulated lows. The studies of Bratseth (1985) and Økland (1987) indicated that convective heating must be maximized at low levels (~2 km) for a CISK mechanism to account for a polar low development. Emanuel and Rotunno (1989) used a theoretical model for a hurricane to argue against CISK and in favor of an air–sea interaction instability mechanism as the primary energy source for polar low-like vortices.

These observational and modeling studies suggest a spectrum of polar lows, with differences in underlying physics as well as in structure. Businger and Reed (1989) found these differences significant enough to classify polar lows into three elementary types based on their baroclinivity, static stability, and surface fluxes of sensible and latent heat. These categories are the 1) shortwave–jet streak type (sometimes termed “comma clouds”); 2) Arctic front-type; and 3) cold-low type. They acknowledged the existence of hybrid-type polar lows.

The present study combines research aircraft measurements with conventional data to investigate the structure of a pair of polar lows that developed in the Gulf of Alaska on 1 December 1987. These aircraft measurements were collected by a National Oceanic and Atmospheric Administration (NOAA) P-3 during the field phase of OCEAN STORMS, an air–sea interaction experiment involving the Office of Naval Research (ONR), NOAA, the National Weather Service (NWS), the Atmospheric Environment Service (AES) and Institute of Ocean Sciences (IOS) from Canada, and United States and Canadian university scientists.

The vortices studied here are termed polar lows, in spite of the asymmetry of their low-level circulations. As is shown later, they did exhibit many of the characteristics of previously studied polar lows. Figures 1a, b show high-resolution infrared (IR) and visible satellite images of the two polar lows of this study. The polar low to the west had an almost symmetrical eye, similar to that of many previously studied polar lows (e.g., Rasmussen 1983). The polar low to the east was at a later stage in its lifecycle than the western polar low and was accompanied by more cumuliform clouds.

Progress in understanding polar lows has been slow due to their small size and their occurrence in data-sparse oceanic regions. Most previous studies (with the exceptions of Shapiro et al. (1987) and Douglas and Shapiro (1989) for arctic front-type polar lows, and Businger and Hobbs (1987) and Businger and Walter (1988) for comma cloud-type polar lows) have relied upon conventional data sources or unverified numerical model diagnostics. The OCEAN STORMS aircraft measurements presented here permitted specification of the synoptic-scale environment of polar low devel-

![Fig. 1. (a) NOAA polar orbiter IR satellite image for 2305 UTC 1 December. (b) NOAA Polar Orbiter visible satellite image for 2305 UTC 1 December.](image-url)
opment and analysis of some of the mesoscale aspects of their structure.

2. Data sources and analysis techniques

Much of this study is based on measurements collected on the flight mission of 1 December 1987; therefore the flight is first briefly reviewed and then the data sources and method of analysis are outlined.

The primary objectives for the flight mission of 1 December were to investigate the structure of a low-level westerly jet and to measure the oceanic mixed layer’s response to forcing by this jet. The first half of the flight consisted of an Omega dropwindsonde (ODW) survey of the wind, temperature, and humidity of the synoptic-scale environment. Low-level winds from a north–south-oriented line of ODWs revealed the presence of the westerly jet. The aircraft then descended to ~3000 m and later to ~1500 m to deploy air-expendable current profilers (AXCPs) along this same north–south track. Detailed low-level observations of a narrow (~20 km) horizontal shear zone in the vicinity of the center of the synoptic low were taken during the remainder of the flight. Additional ODWs were deployed on the return flight to Seattle, Washington. Satellite images reviewed after the flight revealed the polar low cloud features that motivated the analysis presented here. The synoptic-scale analysis describes the environment in which the polar lows developed. Surface analyses were based on observations from buoys, ships of opportunity, coastal stations, and surface reports from ODWs. Upper-air analyses were prepared from ODW and flight-level measurements from the P-3, from rawinsonde soundings from the Canadian weathership Parizeau, and from the standard network of upper-air stations bordering the Gulf of Alaska and eastern North Pacific. Asynoptic (off-time) observations were translated in space using a two-dimensional Taylor’s hypothesis.

Mesoscale analyses were carried out for the leading polar low from ODWs spaced ~100 km apart and flight-level observations. These observations were used to construct vertical cross sections across a low-level baroclinic zone accompanying the polar low. Flight-level observations collected in the vicinity of the trailing polar low are also shown. Surface weather observations associated with this disturbance as it approached the British Columbia coast are also presented. Analyses of ocean significant wave heights were constructed using reports from buoys and ships of opportunity.

3. Synoptic perspective

The synoptic-scale perspective was prepared from 1) surface maps analyzed at 6-h intervals for the period from 0000 UTC 1 December to 0000 UTC 2 December 1987, inclusive, 2) a constant-pressure analysis at 500 mb at 0000 UTC 1 December, 3) constant-pressure analyses at 850, 700, and 500 mb at 1200 UTC 1 December, and 4) satellite imagery. P-3 ODW data were used in analyses at 1200 and 1800 UTC 1 December.

The sea level pressure analysis, surface frontal analysis, and ship and buoy observations for 0000 UTC 1 December (Fig. 2a) show that the dominant synoptic feature was the mature low centered near 50°N, 145°W. A long fetch of strong westerlies occurred along a band between about 47° and 40°N, south of the center. Relatively warm surface air temperatures (~ 8°C) were observed near the low center, and colder (~1° to 5°C) temperatures were observed in the north-northwesterly flow farther north near 52°N, 152°W. The GOES-W IR satellite image at 0046 UTC 1 December (Fig. 2b) shows the cloud shield associated with the occluded front extended along a ~300-km wide band north of 50°N. This image suggests that a tongue of relatively warm air had encircled the north side of the low center in the form of a back-bent occlusion. Moderate cellular convection was prominent in the cooler and convectively unstable westerly flow to the south of the low center. No polar low cloud circulations were evident at this time, but the first polar low (PL1) developed within the stratiform patch of clouds near 49°N, 145°W shown in Fig. 2b. The stratuscumulus cloud streaks immediately upstream from this location and the surface analysis (Fig. 2a) suggest confluence of the low-level flow in this region, which enhanced the cross-stream thermal gradients and their attendant vertical circulations. A similar satellite image was shown by Reed and Albright (1986) for the mature stage of a rapidly developing cyclone; their analysis of buoy data also showed a tongue of warm air wrapping around the surface low center.

The National Meteorological Center (NMC) height and temperature analysis at 500 mb for 0000 UTC 1 December (Fig. 3) shows weak westerly geostrophic flow over the initiation region of PL1 near 50°N, 145°W. The composite of 21 Gulf of Alaska polar lows presented by Businger (1987) includes cold temperatures (< ~36°C) at the 500-mb low center and 1000–500 mb thicknesses on the order of 500 dam in the initiation region of polar lows. Businger’s composite sea level pressure and 1000–500 mb thickness patterns also show weak lower-tropospheric temperature advections on the synoptic scale. The present analysis shows 500-mb temperatures were ~ ~32°C, and 1000–500 mb thicknesses were about 540 dam, implying greater static stability than in the Businger (1987) composite. More important, the geostrophic vertical shear between the surface and 500 mb in the present case implies that significant cold air advection was occurring to the west and southwest of the surface low center, while weak temperature advection was occurring near the low center. This thermal structure was not resolved by the Businger (1987) analysis; his focus was on the larger-scale environment conducive to polar low development.
The sea level pressure and frontal analysis for 0600 UTC 1 December (Fig. 4a) shows that the synoptic low had deepened by 4 mb to 955 mb and moved to the east-northeast at about 12 m s⁻¹ during the previous 6 h. Ships in the northerly flow west of the synoptic low reported wind speeds up to 20 m s⁻¹. The corresponding IR satellite image at 0646 UTC (Fig. 4b) shows an expansion in the solid cloud deck near 50°N, 141°W, associated with PL1. At this time, PL1 was close to the synoptic low center.

The sea level pressure and frontal analysis for 1200 UTC 1 December (Fig. 5a) shows that the synoptic...
low propagated to the north-northeast at about 11 m s\(^{-1}\) and had little change in central pressure during the previous 6 h. The pressure gradient associated with PL1 intensified, as documented by the ODWs along 139°W. The report from the ODW at 48.5°N, 139.0°W included a wind speed of 36 m s\(^{-1}\) from 278° at 470
m above the surface. The IR satellite image for 1146 UTC 1 December (Fig. 5b) shows that the cloud mass associated with PL1 continued to expand, taking on the comma shape of a vortex circulation. Farther north, near 52°N, 138°W, the clouds associated with a second polar low (PL2) were beginning to organize. Cellular cumulus convection continued in the wedge of cold air between 50° and 42°N. This convection appeared to be suppressed south of PL1.

The 850-, 700-, and 500-mb height and temperature analyses at 1200 UTC 1 December (Figs. 6a–c) incorporated the ODW soundings taken between 1200 and 1500 UTC. All three analyses show the frontogenetical effects of differential cold air advection and confluence.
of warm and cold airstreams to the south of the synoptic low center. The 850-mb analysis (Fig. 6a) shows a \(\sim 2^\circ C\) temperature contrast and 20 m s\(^{-1}\) of cyclonic shear across the 100-km baroclinic zone associated with PL1. As shown in Figs. 6b–c, this baroclinic zone sloped south-southwestward with height toward colder air, and the cyclonic shear decreased with height. The low-level wind speed maximum associated with PL1 was situated on the cold side of this baroclinic zone. The thermal wind within the baroclinic zone was from the east-southeast, as shown by the temperature analyses and the vertical wind shear from the ODWs. Since the ther-

![850 mb](image_url)

**Fig. 6a.** Height (solid lines, 3-dam interval) and temperature (dashed contours, 2°C interval) at 850 mb for 1200 UTC 1 December. Line AA' is used in the cross-section analyses in section 4.
mal wind direction was opposite to the propagation direction of the polar lows, this case represents an example of reverse shear flow defined by Duncan (1978). The width of the baroclinic zone and associated reverse shear was <200 km, compared with a width of about 1000 km shown by Reed and Duncan (1987) in their
study of a family of polar lows. Reed and Duncan (1987) did not have the benefit of mesoscale observations; the baroclinity in their case may have been concentrated on a smaller and shallower scale. The synoptic-scale environment presented here resembles that of the majority of the cases in the composite of Forbes and Lotes (1985) and the polar low studied by Shapiro et al. (1987). In particular, Shapiro et al. (1987) showed that the polar low developed at the inside edge of a band of low-level cold advection from the west encircling a synoptic-scale surface low. The only upper-level observations upstream of the line of ODWs along 139°W in our study were the soundings from the Parizeau near 48.5°N, 142°W. The 700–500 mb temperatures from the 1200 UTC sounding were −1°–2°C colder than those from the ODWs about 200 km downstream, suggesting cold air advection at low levels along the cold side of the reverse shear baroclinic zone of PL1.

Radar reflectivity data from the P-3 provide another perspective on the structure of PL1. Figure 7 is a radar reflectivity composite for the period 1412–1425 UTC when the aircraft was over the frontal zone associated with PL1. Also plotted in Fig. 7 are the lowest winds measured by the ODWs during this leg. The distribution and magnitude of radar echoes was consistent with the satellite imagery at this time (e.g., Fig. 5b). Radar reflectivities were ∼25 dBZ in the center of the vortex and somewhat higher west of the center of the vortex. By comparison, Shapiro et al. (1987) found reflectivities exceeding 40 dBZ. Businger and Hobbs (1987) showed convective elements with reflectivities exceeding 40 dBZ embedded within rainbands associated with two comma-cloud-type polar lows.

The sea level pressure and frontal analysis for 1800 UTC 1 December (Fig. 8a) shows that PL1 propagated eastward at 20 m s⁻¹ during the last 6 h and intensified, as shown by the increased cyclonic curvature and pressure gradient. Surface wind speeds ranged from 20 to 25 m s⁻¹. Buoy 46036 at 48.2°N, 133.5°W reported its maximum wind speed at this time, 22 m s⁻¹ from 250° with gusts to 28 m s⁻¹. A sea level pressure trough was also analyzed for PL2 near 52°N, 137°W. The observations in this region clearly show the wind shift associated with this second vortex. In section 4 detailed low-level aircraft measurements of PL2 are presented. The IR satellite image for 1746 UTC 1 December (Fig. 8b) displays the tightly curved cloud circulation of PL1 and the blossoming of the cloud mass associated with

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1 This buoy has an anemometer height of 5 m; extrapolating this report to a typical ship anemometer height of 19.5 m using a Monin–Obukhov similarity relationship yields a sustained wind of 26 m s⁻¹.
PL2. There was suppression of the cellular convection immediately to the south of the two polar lows.

The sea level pressure and frontal analysis at 0000 UTC 2 December (Fig. 9a) shows that the synoptic low moved to 54°N, 137°W and filled to 961 mb. PL1 was situated near 52°N, 130°W and the British Columbia coast and was beginning to dissipate. Here PL1 and PL2 were separated by ~400 km. The sea level pressure gradients in the vicinity of PL2 increased during the last 6 h; surface wind speed reports in the westerly flow behind the trough associated with PL2 ranged between 16 and 25 m s⁻¹. The increase in the strength of the circulation around PL2 occurred while the sea level pressure in its vicinity was rising. The IR satellite image at 0046 UTC 2 December (Fig. 9b) shows that the clouds associated with PL1 merged to a certain
extent with other less organized convective clouds along the coast. The clouds of PL2 were now organized into a spiral pattern. Arctic air continued to flow from the west-northwest into the Gulf of Alaska, as indicated by the cloud streets in Fig. 9b and the surface reports of Fig. 9a. The cloud shield formerly associated with the occluded front was no longer prominent; it is speculated that relatively warm air was no longer available for additional secondary baroclinic developments in the vicinity of the synoptic low center.

The interpretation here of the satellite imagery (Figs. 1a–b, 2b, 4b, 5b, 8b, and 9b) is that the condensational heating accompanying PL1 and PL2 was mostly stratiform during their development. The high-resolution
IR image at 2305 UTC 1 December (Fig. 1a) reveals that cloud-top temperatures were \(-30^\circ\)C in the vicinity of PL2. More cumuliform clouds and cloud-top temperatures of \(-40^\circ\)C were associated with PL1, which was at a later stage in its lifecycle. Cumulus convection appears to have been much more prominent during the growth phases of the polar lows reviewed by Rasmussen (1983).

The upright convective clouds associated with PL1 during its mature phase appear to be a consequence rather than an important cause of its development. Surface heat and moisture fluxes were calculated from the ODWs released along 139\(^\circ\)W at about 1400 UTC using the bulk aerodynamic formulations of Large and Pond (1982) and a Monin–Obukhov similarity relationship. The maximum total surface heat fluxes (sensible and latent) were \(-600 \text{ W m}^{-2}\) at the ODW near 48.5\(^\circ\)N, 139.0\(^\circ\)W on the cold side of the baroclinic zone. Four hours later at buoy 46036 (48.2\(^\circ\)N, 133.5\(^\circ\)W), in a similar location relative to PL1, total heat surface fluxes were \(-350 \text{ W m}^{-2}\). These surface fluxes cause the equivalent potential temperature \((\theta_e)\) of the boundary layer to asymptotically approach a value equivalent to the sea surface temperature (SST) at saturation and act to reduce the low-level baroclinity associated with the disturbance. Evidently, this heating and moistening of the boundary layer was sufficient to destabilize a deep layer of PL1's circulation to upright convection by about 2100 UTC. Strong winds (and large surface heat fluxes) were associated with PL1 before upright cumulus convection became prominent, based on our interpretation of the satellite imagery. The air–sea interaction instability outlined by Emanuel and Rotunno (1989) may still pertain to these polar lows, since the surface heat and moisture fluxes were substantial.

An additional synoptic-scale perspective is provided by output from the NMC Nested Grid Model (NGM), which indicates the potential for secondary baroclinic development in the region of observed polar low development. Figures 10a–c show a selected region of the
12-h NGM forecast valid at 1200 UTC 1 December for sea level pressure and 1000–500 mb thickness, 700-mb vertical velocity, and 500-mb geopotential height and vertical vorticity, respectively. These forecasts capture much of the structure shown in the subjective analyses. The sea level pressure and 1000–500 mb thickness map (Fig. 10a) shows cold air advection in the northwesterly flow well behind the synoptic center and weaker warm air advection in the north-northeastly flow near 50°N, 142°W. This confluence and differential temperature advection is frontogenetical near 46°N, 142°W. The 700-mb vertical velocity (Fig. 10b) shows a band of upward motion near 48°N, 141°W and downward motion in a region of cold air advection to the southwest of this band. This vertical velocity pattern is in accord with a Sawyer–Eliassen secondary circulation at a front associated with confluence and cyclonic shear deformation of a positive alongfrontal thermal gradient. The 500-mb geopotential height and vorticity fields (Fig. 10c) show negative vorticity advection (NVA) by the 500-mb geostrophic wind in the region of upward motion. Since there was reverse shear (as indicated both in the constant pressure analyses of Figs. 6a–c and the 1000–500 mb thickness map of Fig. 10a), the NGM predicted upward motion in a region of midtropospheric positive vorticity advection (PVA) by the thermal wind, in agreement with the simple conceptual model of Sutcliffe (1947). Even though the NGM did not forecast the low-level pressure troughs associated with the polar lows, it did correctly forecast mesoscale conditions favorable for their development.

4. Mesoscale analysis
   a. Cross-section analyses

Analyses were prepared by projecting onto line AA' (shown in Fig. 6a) the data from 10 ODWs released along 139°W. The orientation of this cross section (023°–203°) is perpendicular to the 850-mb thermal wind vector at the ODW at 48.5°N, 139.0°W. Winds from the ODWs were projected into components along and normal to this section. The constant-pressure analyses (Figs. 6a–c) suggest that temperature and wind gradients along the cross section were greater than those normal to it. Separating the wind into components along and normal to the section was less justified for soundings at the northern and southern extremes of the section.

Figure 11 presents the cross-section analysis of potential temperature (θ), the component of the wind from 293° (U), and the selected wind barbs. The horizontal gradient in θ was concentrated within a ~100-km wide frontal zone extending from the boundary layer and toward the south with height. The slope of this frontal zone was ~1:80. The U analysis shows a boundary layer jet exceeding 38 m s⁻¹ at the cold side of the frontal zone. Velocities decreased sharply to the north, and there was a cyclonic shear vorticity of 2 × 10⁻⁶ s⁻¹. The anticyclonic horizontal shear to the

![Figure 11](https://example.com/figure11.png)

**FIG. 11.** Cross-section analyses of potential temperature (K, solid lines) and alongfrontal (from 293°) wind component (m s⁻¹, dashed lines) along the cross-section projection line AA' of Fig. 6a. Heavy solid lines signify the boundaries of the frontal zone. Selected wind reports from the ODWs are also plotted. The dots show the locations of the ODW reports; the release times of the ODWs are indicated at the bottom. The position of the low-level jet is indicated.
south of the jet was about $6 \times 10^{-5}$ s$^{-1}$. Vertical shear of $-6$ m s$^{-1}$ km$^{-1}$ in the reverse shear sense was found above the boundary layer jet; this magnitude was roughly the geostrophic shear associated with the observed horizontal temperature gradient of $\sim 2$ K (100 km)$^{-1}$. The high velocities shown at upper levels at the southern edge of Fig. 11 are associated with the baroclinic zone of the main polar front.

Analyses of $\theta_e$ and relative humidity are shown in Fig. 12. Convective instability ($\partial \theta_e / \partial p > 0$, shaded areas) was present in each sounding in the lower portion of the boundary layer; most of the other convectively unstable regions were collocated with relative humidities below 90% and do not imply instability to upright convection. An intrusion of low $\theta_e$ air extended northward and downward from $\sim 2.5$ km at the southern portion of this plot. The horizontal gradient in $\theta_e$ was concentrated in the frontal zone at the north edge of this tongue. High relative humidities were observed in the boundary layer (below about 1–1.5 km), over deep layers well north of the frontal zone, and in a relatively narrow column at and above the frontal zone. An intrusion of low relative humidity extended downward and northward from about 3 km at the southern portion of the cross section. Measurements from the airborne downward-looking radiometer (PRT-5) were used to determine cloud boundaries and assist the humidity analysis between sounding locations. The relative humidity in cloudy regions was assumed to be greater than 90%. The radiometer in particular defined the horizontal extent of the high relative humidity above the low-level frontal zone.

The analysis in Fig. 13 shows that $V$, the 203° cross-frontal component of the wind, was southwesterly on the south side of the low-level jet and frontal zone. A region of negative velocities (with northeasterly components) extended downward from 5 km into the boundary layer north of the frontal zone. One-dimensional horizontal convergence of about $-1 \times 10^{-4}$ s$^{-1}$ was found in the boundary layer within the frontal zone.

The frontal zone above the low-level jet exhibited substantial vertical wind shear and modest static stability with respect to moist processes, signifying an environment susceptible to slantwise convection (moist symmetric instability). Emanuel (1983) outlined a straightforward technique for evaluating the potential for slantwise convection in situations of straight, geostrophic flow. Sanders (1986) extended this technique for a circular vortex in gradient wind balance. The constant-pressure analyses of Figs. 5a–c show that the streamlines in the vicinity of PL1 were curved; at this stage of development PL1 was in a state somewhere between the extremes of straight and circular flow. For simplicity, the effects of curvature are neglected and the technique of Emanuel (1983) is applied.

Following Eliassen (1962), the absolute momentum $M$ is defined as

$$M = U - fy$$  \hspace{1cm} (1)

where $f$ is the Coriolis parameter and $y$ is the distance from the northernmost sounding of the cross section. A cross-sectional analysis of $M$ and $\theta_e$ is shown in Fig. 14. The stippled area indicates a region where the iso-

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**Fig. 12.** Cross-section analyses of equivalent potential temperature (K, solid lines) and relative humidity (percent, dashed lines) along the cross-section projection line AA' of Fig. 6a. The light shading indicates regions of convective instability.
pleths of $M$ were approximately parallel to the isopleths of $\theta_e$, implying moist symmetric neutrality (Emanuel 1983). Virtually all of this region had a relative humidity exceeding 90% and was within the solid cloud shown in the satellite image of Fig. 5b. Therefore, in this region the buoyancy force on parcels displaced along $M$ surfaces was near zero. Since the time scale for slantwise convection is usually less than that for baroclinic instability (Emanuel 1988), it is reasonable that the lower troposphere had undergone adjustment into a state neutral to slantwise convection.

The synoptic-scale constant-pressure analyses for

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**FIG. 13.** Cross-section analysis of the cross-frontal (from 203$^\circ$) wind component (m s$^{-1}$) along the cross-section projection line AA' of Fig. 6a. The arrows indicate wind direction.

**FIG. 14.** Cross-section analyses of absolute momentum (m s$^{-1}$, solid lines) and equivalent potential temperature (K, dashed lines). Selected values of absolute momentum are indicated with dotted lines. The stippled area indicates where isopleths of absolute momentum are approximately parallel to isopleths of equivalent potential temperature.
1200 UTC 1 December (Figs. 6a–c) show that differential cold air advection was occurring across the reverse shear front in the frontogenetical sense. Using the semigeostrophic framework, Eliassen (1962) showed that this type of thermal deformation \((\partial U/\partial y)(\partial \theta/\partial x) < 0\) forces a thermally direct secondary circulation at a front. The mesoscale analyses of relative humidity (Fig. 12) and cross-frontal component of wind (Fig. 13) suggest a circulation consisting of a relatively narrow upward plume above the low-level jet at the warm edge of the frontal zone with broader, weaker subsidence about 300 km to the south, consistent with a Sawyer–Eliassen ageostrophic circulation at a front with confluence and cyclonic shear accompanied by cold air advection.

The strong cyclonic vorticity \((\sim 2 \times 10^{-4} \text{ s}^{-1})\) implied by the \(U\) analysis of Fig. 11 suggests that Ekman pumping may have been an important forcing mechanism for frontal ascent in the boundary layer. Charney and Eliassen (1949) showed that the magnitude of Ekman pumping can be estimated from a simplified vorticity equation for the boundary layer. The result is that the vertical velocity at the top of the boundary layer \(w_H\) is proportional to the curl of the surface stress, i.e.,

\[
w_H = \nabla \times \tau / \rho f\tag{2}
\]

where \(\tau\) is the surface stress and \(\rho\) is density. Fleagle et al. (1988) showed that estimates of boundary layer vertical velocity from (2) compare well with kinematic calculations of vertical velocity at surface fronts. The surface stresses at two ODWs, released at 1411 and 1421 UTC, that bracketed the low-level frontal zone were estimated at 0.35 N m\(^{-2}\) and 1.60 N m\(^{-2}\), respectively. The values were found by using a bulk aerodynamic formulation based on Large and Pond (1982) and Monin–Obukhov similarity to reduce boundary layer winds to surface layer values. Considering only the alongfrontal components of these surface stresses, an estimate of the curl of the stress yields a vertical velocity of \(8 \times 10^{-2} \text{ m s}^{-1}\). Vertical integration of the convergence in \(V\), the cross-frontal wind component, yields a vertical velocity of about \(10 \times 10^{-2} \text{ m s}^{-1}\) at a height of 1 km. This simple calculation suggests that the frontal zone convergence below 1 km was largely induced by surface friction.

b. Fine-scale structure of PL2

High-resolution measurements of the low-level wind shift zone associated with PL2 were made during the last hour of on-station flight time. Three penetrations were made of the zone near 51\(^\circ\) N, 136\(^\circ\) W at about 1800 UTC, twice at 1500 m and once at 300 m, while the cloud shield accompanying PL2 (Fig. 8b) was still developing. The data show the magnitude and horizontal scale of the changes in wind speed and direction, and suggest that the circulation of PL2 strengthened over the period of these observations. Flight-level data from the two legs at 1500 m were averaged over 50 s \((\sim 5 \text{ km})\) and were plotted in a frame of reference moving with PL2, toward 120\(^\circ\) at 11.6 m s\(^{-1}\). Thermodynamic data and heights were adjusted to a common pressure of 800 mb using lapse rates from aircraft ascents. Both legs penetrated the shear zone along tracks toward 30\(^\circ\) in a similar location relative to PL2. The first leg was flown \(\sim 45\) min before the second leg.

Figure 15 presents 800-mb wind vectors, geopotential heights, air temperatures, and dewpoints. During the first leg, a wind shift zone was encountered near 51\(^\circ\) N; winds from \(\sim 305\(^\circ\)\) were found to the southwest of the zone and winds from \(\sim 220\(^\circ\)\) to the northeast of the zone. The relative vorticity in the vicinity of the wind shift zone was \(\sim 1 \times 10^{-3} \text{ s}^{-1}\). The wind speeds increased toward the southwest in the northwesterly flow, implying a cyclonic shear of \(\sim 10^{-4} \text{ s}^{-1}\). Temperatures were at a maximum in the wind shift zone and decreased toward the southwest. The temperature gradients were weak \([\sim 1^\circ(100 \text{ km})^{-1}]\). The dewpoint depression was near zero everywhere except to the northeast of the zone.

Figure 15 also shows that the south-southwest to north-northeast portion of the second leg was about 15 km to the northwest of the first leg (in a frame of reference moving with PL2). The wind shift encountered during the second leg was at the northernmost portion of the run, about 30 km due north of the wind shift zone penetrated earlier. These observations suggest that the wind shift zone and its associated trough were oriented approximately north–south. This is consistent with the orientation of the trough on the 1800 UTC sea level pressure analysis (Fig. 8a).

Some important changes occurred between the legs shown in Fig. 15. Southwest of the wind shift zone, the wind speeds in the northwesterly flow increased by \(\sim 5 \text{ m s}^{-1}\) and the air temperatures decreased by \(\sim 0.5 \text{ K}\). Geopotential heights along streamlines increased by \(\sim 7 – 9 \text{ m}\), and height rises increased toward the southwest. It appears that the circulation about PL2 was increasing, while pressures in its vicinity were rising. This interpretation is consistent with our sea level pressure analyses at 1800 UTC 1 December and 0000 UTC 2 December (Figs. 8a and 9a, respectively). Pressure rises at the southern periphery of PL2 can be attributed to increasing low-level cold advection during development.

At least qualitatively, the limited flight-level measurements in the vicinity of PL2 resemble the ODW measurements at low levels in the vicinity of PL1. In both cases, a frontal zone with large cyclonic shear was situated at the northeast edge of a region of cold northwesterly flow.

c. Surface station time series

Sequences of hourly surface observations collected from two island stations as PL1 approached the British Columbia coast are shown in Fig. 16. Satellite imagery
indicated that the center of PL1 passed between these two stations at about 2200 UTC 1 December. Here PL1 was propagating at about 14 m s\(^{-1}\); 1 h at a fixed point corresponds to 50 km in a frame of reference moving with the vortex. Both stations are situated on bluffs at altitudes near 100 m; their winds are more representative of the boundary layer than the surface layer. The time series from the northern station, Cape St. James (WFV), shows that an abrupt wind shift from south-southeast to southwest occurred within 1 h. The observer reported an apparent frontal passage at 2217 UTC. Winds gusted up to 25 m s\(^{-1}\) about 2 h before and 3 h after this wind shift. Convective precipitation activity was reported as early as 1600 UTC during the passage of the sea level pressure trough and, to a lesser extent, after the trough. Showers before 2000 UTC were associated with disorganized convective cloud clusters that preceded the clouds accompanying PL1 (see Fig. 1a).

The time series from the southern station, Sartine
WFV Cape St. James 51.9, 131.0 (91 m)

WFV Sartine Island 50.8, 128.9 (111 m)

Fig. 16. Surface observations of pressure (mb), air temperature (bold numerals, °C), and wind (one full barb = 5 m s⁻¹) from Cape St. James (WFV) and Sartine Island (WFG) during the passage of PL1. Observations are plotted using the standard meteorological convention; wind gust speeds (m s⁻¹) and observer comments are also shown. The locations of these stations are indicated by stars and bold pressure reports in Fig. 9a.

Island (WFG), shows a more gradual wind shift from south-southwest to west in ~4 h. Wind gusts of 30 m s⁻¹ were recorded before and during the wind shift and trough passage. This station is unstaffed; therefore, visual observations of cloud and precipitation type were not collected. Both stations experienced a slow drop in surface pressure ahead of the trough, but PL1 was followed by more rapid rises of about 2 mb h⁻¹.

5. Sea-state analysis

The chief hazard caused by these polar lows was not the surface wind, but rather the seas raised by the wind. Extremely high significant wave heights were caused by intense mesoscale forcing on seas preconditioned by moderate forcing on the synoptic scale. The significant wave height of 13.5 m at 2100 UTC at buoy 46005 (46.1°N, 131.0°W) was the greatest measured at any National Data Buoy Center (NDBC) buoy in the North Pacific during the fall and winter season of 1987–88. A detailed analysis of the momentum exchange at the ocean surface and the transformation of energy into long-period swell is beyond the scope of this study. The intent here is to show the rapid increase in significant wave height between about 47°N and 50°N, during the period between 1200 UTC 1 December and 0000 UTC 2 December as PL1 developed and propagated toward the coast. Analyses of significant wave height relied principally on buoy data and were therefore restricted to a region between about 51°N and 40°N and east of 140°W. Buoy reports 1 to 2 h before and after map times (not shown) were used to help determine wave-height gradients and locations of maxima.

Figure 17 shows the significant wave-height analysis for 1200 UTC 1 December. Also plotted are reports of wind, period of maximum wave energy, and swell propagation direction (when available). The highest seas were measured south of 45°N due to the long duration and fetch of westerlies to the south of the parent synoptic low. The significant wave height of 11.0 m reported by buoy 46006 (40.8°N, 137.6°W) is the height for a fully developed sea with winds of 23 m s⁻¹ (e.g., Apel 1987). Surface maps at 0000 and 0600 UTC 1 December (Figs. 2a and 4a, respectively) show winds of 20 to 22 m s⁻¹ upstream of the buoy. Significant wave heights declined rapidly north of 48°N. At 1200 UTC 1 December PL1 was developing rapidly near 50°N, 140°W; ship reports indicated winds of ~20 m s⁻¹ in this area.

Figure 18 shows the significant wave-height analysis for 1800 UTC 1 December. High seas continued to the south of 45°N, but the greatest heights are shown.

Fig. 17. Significant wave-height (Hₜ1/3, m) analysis and wind and swell propagation direction reports for 1200 UTC 1 December. Buoy and ship reports are indicated by open circles and solid dots, respectively. The station model shows the plotting convention.
near 47°N, 133°W. Advection of higher heights from the west appears minimal; most of the increase in this region is attributable to the winds of PL1. During PL1's development, surface air temperatures were about 4°C colder than the SST in the region of strong westerly winds. This static instability in the atmospheric surface layer enhanced the momentum transport to the ocean and contributed to the rapid buildup of the seas by 1800 UTC. The highest waves (~13 m) measured by a buoy were at 46.1°N, 131.0°W, south of the strongest forcing by PL1, but where the seas had been subject to greater synoptic-scale forcing.

Figure 19 shows the significant wave-height analysis for 0000 UTC 2 December. The patch of extremely high waves propagated eastward at ~14 m s⁻¹, which is the deep-water gravity wave speed for waves with a period of 18 s. These waves reached the buoys along the Pacific Northwest coast at about 0500 UTC 2 December. The reports of swell propagation direction available in the vicinity of the wave height maximum near 47°N, 130°W suggest divergence of the swell propagation vectors and therefore horizontal dispersion of the wave energy. The reports north of 50°N also imply significant surface wave generation by PL2, at this time near 52°N, 133°W.

6. Discussion

This study has documented the synoptic-scale and mesoscale environment in which a pair of polar lows developed and has described some of the mesoscale structure of these vortices. Here the findings are summarized, emphasizing comparisons with previous studies of polar lows.

The synoptic-scale analysis revealed that development occurred within a mature, occluded synopticscale cyclone in the Gulf of Alaska. Frontogenesis at low levels occurred as a result of confluence and differential cold air advection between the polar air stream to the west of the synoptic low and the relatively warm air sequestered near the core of the low. Indirect evidence suggests that a Sawyer–Eliassen secondary circulation occurred across this frontal zone: a pair of polar lows developed within the ascending branch of this circulation. The thermal wind in this region was directed in a reverse shear sense, i.e., opposite to the direction.
of propagation of the disturbances. These elements are summarized schematically in Fig. 20.

It is interesting to compare the synoptic-scale environment described here with that of previously studied polar low cases. Within the framework outlined by Businger and Reed (1989), the polar lows described here fit most closely into the arctic-front category. Businger and Reed (1989) found that the shallow baroclinity accompanying this type of polar low was usually due to differential boundary layer heating along an ice edge or continent. In the present case, the baroclinity was not caused by differential heating locally, but was more a result of frontogenesis between a modified arctic air mass and warm air near a mature, synoptic low. The most comprehensive statistical survey of polar lows was carried out by Forbes and Lottes (1985). Their schematic for developing polar lows shows a cold tongue (low 1000–500 mb thicknesses) in the northerly flow behind a nearly vertically stacked synoptic-scale low. Development is favored in this region of reverse vertical shear. These results are consistent with our synoptic-scale analysis. Shapiro et al. (1987) show a similar pattern in their large-scale European Center for Medium-Range Weather Forecasts (ECMWF) analyses, i.e., that a family of polar lows developed at the edge of a zone of lower tropospheric cold advection to the west of a synoptic-scale system. Both the Bering Sea and Gulf of Alaska polar lows described by Businger (1987) developed near the confluence of cold continental air and relatively warm air within a dissipating, occluded low. Rasmussen (1979, 1981, 1985) has carried out numerous case studies of polar lows occurring in situations of large air–sea temperature contrasts. On the basis of the satellite imagery, he suggests that vigorous cumulus convection was taking place during their growth. His analyses also show that development occurred at or near low-level baroclinic zones to the west and northwest of synoptic lows, apparently due primarily to differential surface heating. Grønås et al. (1987) simulated six polar lows that formed over the Norwegian Sea. All these polar lows developed in a baroclinic, reverse shear environment. This baroclinicity was sometimes attributed to differential heating from SST gradients and was sometimes a result of a tongue of warm air within an occluded, synoptic-scale cyclone.

Because of their small-scale and short lifetimes of polar lows, little is known about their mesoscale structure. The ODW, flight-level, and radar data collected by the P-3 permitted a rare opportunity to carry out a mesoscale analysis of a polar low. These results are summarized next and compared with the other analyses of polar lows using P-3 data by Shapiro et al. (1987) and Douglas and Shapiro (1989).

Our mesoscale analysis relied principally on ODWs transecting the leading polar low (PL1) during its growth phase. Vertical cross sections prepared from these ODWs show that a prominent low-level frontal zone accompanied the polar vortex. An alongfront wind maximum of 38 m s\(^{-1}\) was found at an altitude of 1 km; the decrease below this point was due to surface friction, and the decrease above this point was due to reverse geostrophic shear. Relative humidity and cross-frontal wind analyses (and NMC numerical model output) suggest that a thermally direct secondary circulation was occurring at the front. The vertical cross sections used data with a horizontal resolution of \(~\sim 100\) km; almost certainly gradients at low levels were concentrated on smaller scales. Fine-scale flight-level measurements collected in the vicinity of PL2 revealed that a large fraction of the wind shift below 2 km occurred over a scale of \(~\sim 10\) km. The relative vorticity on this scale was \(~\sim 1 \times 10^{-3}\) s\(^{-1}\).

This mesoscale structure exhibited both similarities and differences with the structure of the confluent frontal zone documented extensively by Shapiro et al. (1987). Both sets of analyses show alongfront jets exceeding 35 m s\(^{-1}\) at or below 1 km on the cold edge of the frontal zone, as well as convergence in the cross-frontal wind and ascent (implied in the present case) at low-levels in the frontal zone. The higher-resolution analysis of Shapiro et al. (1987) was able to diagnose subsidence above about 2.5 km just to the warm side of the frontal zone. The analysis of a zone of low relative humidity near 3 km on the warm side of the frontal zone is indirect evidence of a similar feature. Both studies found regions neutral to moist slantwise convection in the frontal zone above the boundary layer. The frontal study by Shapiro et al. (1987) exhibited more prominent maxima in \(\theta\) and \(\theta_v\) at its leading edge. A significant difference between the two cases was the degree of convective activity, as judged by radar reflectivities. Shapiro et al. (1987) found radar reflectivities exceeding 40 dBZ, compared with typical reflectivities of 25 dBZ for the present case. This may partly be because the polar low of Shapiro et al. (1987) was sampled later in its lifecycle.

The polar low analyzed by Douglas and Shapiro (1989) exhibited more features in common with the polar low analyzed by Shapiro et al. (1987) than with the case studied here. Douglas and Shapiro (1989) documented a low-level “warm frontal zone” extending toward the east from the polar low center. This frontal zone featured a \(~\sim 10\)-km wide band of enhanced confluence, shear, and thermal contrast. Shapiro et al. (1987) had less documentation but analyzed a similar structure extending northeastward from their polar low center. The aircraft observations and satellite imagery do not indicate a similar warm frontal-type feature in the present case. Douglas and Shapiro (1989) also show a confluence line extending toward the north-northwest from the surface center; this feature is analogous to but weaker than the southern confluent frontal zone analyzed by Shapiro et al. (1987) and the mesoscale frontal zone studied here. The satellite imagery presented by Douglas and Shapiro (1989) suggests that more cumulus convection was occurring during development than in the case studied here.
We have argued that frontogenesis by the synoptic-scale flow brought about conditions conducive for development. The satellite imagery, in particular the image at 1146 UTC (Fig. 5b), suggests that the polar lows were associated with mesoscale waves along the extension of the occluded front to the north and west of the synoptic low center. Theoretical studies of mesoscale baroclinic instability (e.g., Mansfield 1974; Duncan 1977; Reed and Duncan 1987; Moore and Peltier 1989) have shown that shallow baroclinic layers with low stability, hence, small Richardson numbers, can support rapid development over scales as short as 500 km on the scale of the polar lows observed here. Wiin-Nielsen (1989) and van Delden (1989) contend that baroclinic instability can account for the initial growth of polar lows and that CISK or air–sea interaction instability mechanism can contribute only to the latter stages of their growth.

Even though baroclinic instability, enhanced by latent heating, probably represented the primary source of eddy kinetic energy for these polar lows, other mechanisms or preexisting structures may have been important or even crucial for their initiation. Zick (1983) proposed that preexisting upper-level vorticity maxima migrating around a large-scale trough were associated with the development of subsynoptic-scale disturbances. The growth of these disturbances, observed in satellite imagery, occurs as vorticity maxima move through the base of the trough. Zick's mechanism may well have been operating in the present case. It remains an open question whether the initiation of these polar lows was due to preexisting structures of finite amplitude (e.g., Farrell 1982) or was more a consequence of normal-mode growth from small perturbations.

Barotropic instability represents another mechanism that may have been important. The necessary condition for barotropic instability was met (the low-level frontal zone and associated jet represented a local maximum in absolute vorticity), but it is not known if the barotropic growth rate was sufficiently large to have been important. The simple shear instability model of Haurwitz (1949) does imply a relatively short e-folding time scale of ~6 h. The 500-km spacing of the polar lows is not inconsistent with a barotropic wavelength of 6 to 8 times the width of the frontal zone. On the other hand, Sardie and Warner (1985) found a small barotropic instability growth rate (the time scale was ~7 days) for a jet with a broader scale than in the present case, but with a similar magnitude of horizontal shear.

Polar lows represent a challenge for the marine forecaster. Most operational weather prediction models are not likely to forecast their development accurately, although, as shown here, they are able to anticipate synoptic-scale environments favorable for development. The high-resolution operational model of the Norwegian Meteorological Institute has had success in forecasting polar lows, but underpredicts their intensity (Granás et al. 1987). The present case represented a particularly difficult forecast problem, since strong surface winds occurred before satellite imagery clearly revealed a tight spiral of clouds or the presence of organized convection.

This case would appear to be a good candidate for numerical model experiments. It would be interesting to see if current mesoscale models could simulate the observed structure. Because the sea state was so extreme, it would also be worthwhile to attempt a hindcast of the ocean wave field influenced by the polar lows.

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