A Comparison of Two Coastal Barrier Jet Events along the Southeast Alaskan Coast during the SARJET Field Experiment

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ABSTRACT
The Southeastern Alaskan Regional Jets experiment investigated the structures and physical processes of barrier jets along the coastal Fairweather Mountains near Juneau, Alaska, from 24 September to 21 October 2004. This paper compares in situ aircraft data and high-resolution simulations from the first intensive observation period (IOP1), which featured a nearly terrain-parallel barrier jet (classical jet) with another coastal jet event (IOP7) that was influenced by offshore-directed gap flows at the coast (hybrid jet). IOP1 featured southerly onshore flow preceding a landfalling trough, which was blocked by the coastal terrain and accelerated down the pressure gradient to produce a 5–10 m s$^{-1}$ wind enhancement in the alongshore direction in the lowest 1 km MSL. In contrast, IOP7 featured higher surface pressure and colder low-level temperatures to the east (inland) of the study area than did IOP1, which resulted in offshore-directed coastal gap flow exiting Cross Sound below 500 m that turned anticyclonically and merged with the ambient flow. Unlike the classical jet (IOP1), IOP7 had a surface warm anomaly adjacent to the steep coastal terrain, while a cold anomaly existed farther offshore within the gap outflow. Above the shallow gap flow (>500 m MSL), there were more classical barrier jet characteristics. High-resolution fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) simulations were performed to compare the structures and underlying dynamics between the two cases. Model trajectories show that coastal winds for IOP1 originated offshore, while much of the coastal flow in IOP7 had gap flow origins near the surface and offshore origins above the gap outflow. A model momentum budget suggests that the vertical mixing of southerly momentum from aloft forced the gap outflow in IOP7 to turn anticyclonically more rapidly than an inertial circle. A simulation of IOP7 with the Cross Sound gap removed (filled in) produced a coastal jet with similar maximum wind speeds to the control but resulted in a reduction in the width of the coastal jet by about 40%.

1. Introduction
Strong low-level terrain-parallel winds, known as barrier jets (Parish 1982), can reach high wind speeds (>30 m s$^{-1}$) along prominent two-dimensional mountain ranges. This phenomenon occurs frequently during the cool season along coastal southeastern Alaska (Loescher et al. 2006; Overland and Bond 1993, 1995; Macklin et al. 1990), and can result in hazardous conditions that affect the local fishing, shipping, and aviation industries (Macklin et al. 1990). Alaskan barrier jets typically occur when there is an upper-level trough over the Aleutian Islands and a ridge over western Canada, which favors low-level southerly flow impinging toward the Alaskan coastal terrain (Colle et al. 2006). The mesoscale structure of barrier jets in southeast Alaska was investigated using research aircraft measurements collected during the Southeastern Alas-
kan Regional Jets (SARJET) experiment between 24 September and 21 October 2004 (Fig. 1; Winstead et al. 2006). SARJET focused on a steep range of mountains known as the Fairweathers, which have peaks over 3000 m above mean sea level (MSL) within ~25 km of the coast (Fig. 1). The terrain in the SARJET region is highly complex with numerous coastal gaps, such as Cross Sound gap, which is a sea level gap ~50 km wide and located immediately to the southeast of the Fairweathers. Cross Sound is a common location for gap outflows as quantified using synthetic aperture radar observations (Loescher et al. 2006).

Barrier jet events have been studied over the Sierra Nevadas (Parish 1982; Marwitz 1983), along the coasts of California–Oregon (Doyle 1997; Yu and Smull 2000), Antarctica (Schwerdtfeger 1975), Norway (Barstad and Grønås 2005), and Taiwan (Li and Chen 1998, Yeh and Chen 2003), as well as in the Rockies (Colle and Mass 1995), Appalachians (Richwien 1980; Bell and Bosart 1988), and along the Alps (Chen and Smith 1987). These jets occur when a cold anomaly and associated high pressure perturbation develop against a steep mountain barrier. This cold anomaly can either result from a source region, such as in cold air damming (Bell and Bosart 1988; Colle and Mass 1995), adiabatic ascent over the barrier (Mass and Ferber 1990), or diabatic cooling from precipitation (Marwitz 1983). When the flow is blocked [Froude number, \( F_r = U/nh_m \) < 1, where \( U \) is the low-level flow speed perpendicular to the barrier, \( N \) is the Brunt–Väisälä frequency, and \( h_m \) is the effective mountain height], the wind can accelerate down the along-barrier pressure gradient to produce a barrier jet. For relatively long barriers, with an along-barrier Rossby number \( (V/fL) < 1 \), where \( V \) is the along-barrier flow, \( L \) is the length of the barrier, and \( f \) is the Coriolis parameter, an approximate geostrophic balance develops in the cross-barrier direction (Overland and Bond 1993), while an antitriptic balance typically occurs in the along-barrier direction (Bell and Bosart 1988; Colle and Mass 1995). The cross-barrier extent of this barrier jet extends about a Rossby radius of deformation from the terrain \( (L_R = Nh_m/f) \) (Pierrehumbert and Wyman 1985).

Enhanced terrain-parallel winds generated by the above mechanisms are regarded as “classical” barrier jets (Loescher et al. 2006), because they assume a quasi-two-dimensional terrain with flow impinging toward the barrier. Loesher et al. (2006) also discussed a “hybrid” barrier jet, which involved a nearby offshore-directed gap flow merging with a barrier jet near the coast. Colle et al. (2006) showed that these hybrid jets occur when there is a cold air source over interior Alaska, which favors an offshore-directed pressure gradient. Offshore-directed gap flows are quite common along the western coast of North America, such as through the Fraser River gap (Mass et al. 1995), the Shelikof Strait (Lackmann and Overland 1989), Cook Inlet (Macklin et al. 1990), the Strait of Juan de Fuca (Colle and Mass 2000; Doyle and Bond 2001), and California’s Petaluma Gap (Neiman et al. 2006).

Winstead et al. (2006) provides some SARJET observations of a classical barrier jet on 26 September 2004 [during intensive observing period 1 (IOP1)] and a hybrid barrier jet on 12 October 2004 (IOP7). They highlight some flow differences between the IOPs to motivate the SARJET analysis. The goal of this paper is to provide a more detailed three-dimensional analysis of both events in order to increase our understanding of the structural and dynamical differences between classical and hybrid barrier jets. This study addresses the following questions.

- How do mesoscale structures (e.g., jet morphologies and thermodynamic distributions) and ambient conditions differ between the classic and hybrid jet cases?
- What is the impact of gap outflow from Cross Sound and coastal downslope flows during hybrid barrier jet situations?
- How do the momentum balances and associated airflows differ between a classic and a hybrid barrier jet?
The following section describes the field observations and the setup of the mesoscale model simulations. Sections 3 and 4 present the observations and model simulations for IOP1 and IOP7, respectively. Section 5 compares a trajectory analysis for both IOPs, and section 6 presents a momentum budget analysis. Section 7 describes a sensitivity test to quantify the influence of the gap outflow on the hybrid jet. Our results are summarized in section 8.

2. Data and methods

Flight-level measurements for SARJET were obtained from the University of Wyoming’s King Air research aircraft. These measurements consisted of in situ observations from south of Cross Sound (labeled A in Fig. 1) to near Yakutat (labeled D), and from four southwest–northeast flight legs at various altitudes from the coast (labeled C) to 75 and 120 km offshore (labeled E and E’, respectively). This flight pattern was completed twice for each IOP, and will be referred to as flight 1 and flight 2, respectively. The aircraft data were augmented by some additional data, such as buoy 46083 and the sounding site at Yakutat (PAYA) (Fig. 1).

The fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5; Grell et al. 1994) was used to further elucidate the differences between the two SARJET cases. Three computational domains were used (Fig. 2b), with grid spacings of 36-, 12-, and 4-km. A 1.33-km nest was also utilized; however, the simulated structures were not significantly different from the 4-km nest, so the 1.33-km results are not highlighted in this paper. Each model domain was run simultaneously using one-way nesting, so that the impact of different model resolutions can be quantified. Thirty-two model sigma levels were applied, with 15 levels below 700 hPa in order to better resolve the boundary layer processes. A 10° and 1° land-use and topography dataset was utilized in the 36- and 12-km domains, respectively, while the 4-km domain used a 30° topography dataset.

Many different model configurations were tested to obtain the best simulation. A majority of the variance in the model solutions occurred when applying various planetary boundary layer (PBL) schemes. The control model configuration for IOP1 applied the Blackadar PBL (Zhang and Anthes 1982) scheme, while the 1.5-level closure [turbulent kinetic energy (TKE) based] Mellor–Yamada scheme (Mellor and Yamada 1974) was applied for IOP7. The TKE-based scheme was not utilized in IOP1, because the wind variations were not well simulated along the Fairweathers, while the Blackadar PBL scheme was more realistic. The reasons for these PBL differences will be explored in future work. For both IOPs, the Grell cumulus parameterization (Grell 1993) was used on the 36- and 12-km domains, while the precipitation was explicitly resolved in the 4-km domain using the simple ice microphysical scheme (Dudhia 1989). Klemp and Durran’s (1983) upper-radiative boundary condition was used in order to prevent gravity waves from being reflected off of the model top.

The initial and boundary conditions were provided by National Centers for Environmental Prediction Global Forecast System analyses at 1° resolution every 6 h. Four-dimensional data assimilation, as described in Stauffer and Seaman (1994), was applied to the 36-km domain, in which moisture, temperature, and winds fields were nudged during the first 12 h.

3. IOP1: 26 September 2004

a. Synoptic evolution

As highlighted in Winstead et al. (2006), the large-scale flow during 26 September 2004 was similar to the classic jet composite (Colle et al. 2006), with a high-amplitude 500-mb ridge over western North America and a broad trough over western Alaska (not shown). A surface cyclone was over the northern Gulf of Alaska at 1800 UTC 26 September (see Fig. 5 in Winstead et al. 2006), which was well forecast by the 36-km MM5 (Fig. 2a), with model sea level pressure errors typically <1 mb. Both the observations and model indicate a weak trough extending southward along 140°W, while the surface winds a few hundred kilometers east of this trough over the SARJET region were south-southeasterly at ∼20 m s⁻¹.

By 2100 UTC 26 September 2004 (Fig. 2b), the surface trough located 150 km southwest of the Fairweathers separated the enhanced (20–25 m s⁻¹) southeasterly winds near the coast and the 10–15 m s⁻¹ south-southwesterlies behind (west of) the trough. At 12-km grid spacing, there was weak flow blocking and deflection of the winds along the coast. The following mesoscale analysis further validates the model simulation at 4-km grid spacing.

b. Aircraft observations and model simulations

In situ observations taken as the aircraft ascended and descended to different altitudes between 150 and 1000 m MSL. ~90 km west of the coast (E in Fig. 1) were combined to illustrate the ambient low-level conditions during flight 1 (Fig. 3; black lines). Because the relative humidity in this layer was nearly saturated (~88%–98%), both a moist Brunt–Väisälä frequency
FIG. 2. (a) A 36-km MM5 simulation showing sea level pressure (black every 6 mb), surface temperature (thin dashed lines every 3°C), and surface winds barbs (full barb = 5 m s⁻¹) at 1800 UTC 26 Sep 2004. (b) A 12-km MM5 simulation showing sea level pressure (black every 4 mb), temperature (thin dashed lines every 2°C), wind speed (gray shaded every 5 m s⁻¹), and winds barbs (full barb = 5 m s⁻¹) at 2100 UTC 26 Sep 2004. Boldface dashed lines denote the positions of the surface troughs.
and dry $N$ were calculated. Figure 3b shows an average moist $N$ of $\sim 0.001$ s$^{-1}$, while the dry $N$ was 0.009 s$^{-1}$ (not shown). At point E there was a slight decrease in equivalent potential temperature with height (Fig. 3a), with the model having a slight cool bias (1–2 K). The mean wind speed in this layer was $\sim 27$ m s$^{-1}$, with the model having a 2–3 m s$^{-1}$ wind speed overprediction at 700–850 m MSL (Fig. 3c).

The flow and stability were also calculated for an upstream volume in the model (boxed U region in Fig. 1) between 0 and 2500 m in order to better represent the average ambient flow conditions approaching the Fairweathers. The average terrain-normal component was $\sim 12$ m s$^{-1}$, with a dry and moist $N$ values of 0.01 and 0.003 s$^{-1}$, respectively. This yields dry and moist Fr values of 0.48 and 1.6, respectively. Early in the IOP there were only scattered areas of precipitation over the windward slope (not shown), so the actual Fr was likely between the dry and moist values (Fr $\sim 1$). The nondimensional mountain slopes, given by the square root of the Burger number ($N_{hm}/fL_m$) (Cushman-Roisin 1994), are 8.0 and 2.4 for a dry and moist $N$ (where $h_m \sim 2500$ m, $f \sim 1.25 \times 10^{-4}$ s$^{-1}$, and $L_m \sim 25$ km), respectively, with both numbers suggesting the po-

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*1 Both a moist and a dry $N$ were calculated at this time since the relative humidity was 85%–95% upstream of the barrier (not shown).
tential for flow blocking (Smith 1979; Pierrehumbert and Wyman 1985).

During flight 1 (1445–1815 UTC), the observed near-coast wind speed enhancement at 150 m MSL was ~5 m s\(^{-1}\) relative to the winds ~60 km southwest of the Fairweathers (Fig. 4a). The 4-km MM5 wind enhancement at 1700 UTC 26 September 2004 was somewhat weaker than observed, with ~25 m s\(^{-1}\) southeasterlies adjacent to the Fairweathers decreasing to ~22 m s\(^{-1}\) about 80 km offshore (Fig. 4b). Meanwhile, at 1000 m (Figs. 4c and 4d), the observed and simulated winds were ~30 m s\(^{-1}\) and oriented nearly parallel to the coastal terrain. The observations at both levels show a slight cool anomaly (~1°C) near the coast.

During flight 2 (2100–2350 UTC) at point E, the ambient flow was saturated between 150 and 1000 m (not shown). Both the model and observations (Fig. 3b; thin gray lines) had an \(N_m\) of ~0.006 s\(^{-1}\), which is more stable than flight 1. Meanwhile, there was a slight increase in the onshore wind component to ~10 m s\(^{-1}\) between flights 1 and 2 (Figs. 3c and 4d). For the region ~200 km upstream of the Fairweathers (around point U in Fig. 1), the average flow component toward the barrier varied from 12 m s\(^{-1}\) below 1000 m to 18 m s\(^{-1}\) above 1800 m, while \(N_m\) increased from 0.006 s\(^{-1}\) below 1000 m to 0.01 s\(^{-1}\) above 1800 m. As a result, the moist Fr varied from the partially blocked regime of ~0.8 below 1000 m MSL to the unblocked regime (Fr ~ 2.5) above 1800 m MSL.

The observed and simulated winds at 150 m had a barrier jet maximum (to 25 m s\(^{-1}\)) adjacent to and slightly downstream (northwest) of the highest peak of the Fairweathers (Figs. 5a and 5b). The 150-m winds veered to southerly 150 km upstream (south) of the coast; however, some of this wind shift may have been associated with the approaching surface trough at 2100 UTC 26 September 2004 (Fig. 2b). At 1000 m MSL during flight 2 (Fig. 5c), the winds veered 40° from southeasterly to southerly with the offshore trough, with the model indicating a weak thermal ridge at this
location. There was a weak wind speed enhancement from $\sim 27 \text{ m s}^{-1}$ at point E to $\sim 30 \text{ m s}^{-1}$ about 20 km upstream (west) of the coast, which was well simulated by the model. During the next 3 h, the barrier jet width at 1000 m narrowed as the surface trough approached the coast (not shown).

The sounding at point C during flight 2 (Fig. 3; thick gray lines) represents the vertical profile through the barrier jet near the coast. Although the model had a 1–3-K cool bias at all levels (Fig. 3a), the moist static stability was similar to that observed below 1500 m MSL, with the exception of a spike observed in the 800–900-m layer (Fig. 3b). The peak winds of 34 m s$^{-1}$ occurred at 800–1000 m MSL (Fig. 3c). The 5–10 m s$^{-1}$ wind speed enhancement of the alongshore flow is consistent with the scale analysis of Overland and Bond (1995). They found for Fr < 1 (neglecting friction) that the alongshore wind enhancement was comparable to the onshore component of the incident flow. Above 1000 m MSL, the Fr was >1; therefore, the alongshore wind enhancement scales as $N_{m}h_{m}$, which is $\sim 7.5 \text{ m s}^{-1}$ ($N_{m} \sim 0.005 \text{ s}^{-1}$ and $h_{m} \sim 1500 \text{ m}$), and this enhancement decreases linearly to zero at the ridgetop (2500 m MSL).

A cross section of terrain-parallel winds and potential temperatures was constructed using the west to east (E to C) flight legs at 150, 300, 500, and 1000 m (Fig. 6). For flight 1 (Figs. 6a and 6b), the terrain-parallel winds were $\sim 30 \text{ m s}^{-1}$ near the coast between 500 and 1000 m MSL in both the aircraft and 4-km MM5 results. The terrain-parallel wind speed in the model decreased to $\sim 25 \text{ m s}^{-1}$ at 1500 m MSL about 20 km upstream of the barrier, with relatively few terrain flow enhancements above this level. This reduced blocking at $\sim 1500 \text{ m}$ likely results from an increase in the effective moist Fr above the midmountain level (Fr $\sim 3$, where $h_{m} \sim 1000 \text{ m}$, $U \sim 15 \text{ m s}^{-1}$, and $N_{m} \sim 0.005 \text{ s}^{-1}$), because

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*The profile at point C was not available during flight 1.*
this flow only has to surmount the remaining 1000 m of the Fairweathers. Both the model and the observations have a slight tilt of the isentropes upward toward the coast above 500 m, while there was a weak cold dome at the surface early in the event (Fig. 6a), which was somewhat weaker than observed in the model.

During flight 2 (2100–2350 UTC), the observed and simulated winds (valid for 2100 UTC) were directed more onshore as the weak surface trough approached (Figs. 6c and 6d). The terrain-parallel wind speed component in the 500–1000-m layer increased from ~20 m s⁻¹ offshore (point E) to over 30 m s⁻¹ at the coast. Meanwhile, the flow above 2000 m MSL was considerably less blocked, with terrain-parallel winds speeds <20 m s⁻¹ at the coast in the model.

To illustrate the small-scale variability within the jet during flight 2, Fig. 7 shows the observed time series for the C to E flight legs at 1000, 300, and 500 m MSL as well as the 15-min model data linearly interpolated in time and space along these flight legs. Southwestward along leg C to E at 1000 m for 2150–2207 UTC (Fig. 7a), the model accurately simulated the gradual increase in temperature of ~1°C (Fig. 7a), the small decrease in wind speed of about 2–3 m s⁻¹ (Fig. 7b), and the ~20° wind shift to more southerly (Fig. 7c); however, the model was slightly more southerly than observed toward point E. At 300 m (2208–2219 UTC), there was a reduction in wind speed of 2–3 m s⁻¹, and the observed horizontal wind speeds were more variable than at 1000 m, suggesting significant vertical transport of momentum associated with the +2 to −1 m s⁻¹ vertical motions (Fig. 7d). Although the model wind speeds were less variable, the model did capture the area of enhanced winds to 32 m s⁻¹ at the eastern end of the second leg at 300 m and the third leg at 500 m (near point C at 2220 UTC). The rainwater content (Fig. 7e), as measured by the aircraft’s 0.2-mm-resolution precipitation probe, had a maximum of ~1 g m⁻³ near the shore (C at 2220 UTC). The simulated rain maximum occurred over twice the horizontal distance as observed, but it remained near this location for much of

FIG. 6. Cross section between C and E showing winds (full barb = 5 m s⁻¹), terrain-parallel wind component (solid every 2 m s⁻¹), and potential temperatures (dashed every 1 K) for the (a) observations during flight 1 of IOP1 and (b) 4-km MM5 simulation at 1700 UTC 26 Sep 2004. (c) Same as in (a) but for flight 2. (d) Same as in (b) but at 2100 UTC 26 Sep 2004. The model terrain-parallel winds are also shaded (every 4 m s⁻¹).
the second flight (not shown). The model indicated that there was enhanced convergence at 500–1000 m MSL at point C (Fig. 6), which enhanced the upward motion between 500 and 1500 m and the precipitation rates upstream of the barrier. This precipitation enhancement upstream of the barrier associated with flow blocking is similar the “blocking front” events documented by Colle et al. (2005) along the Wasatch Mountains and Neiman et al. (2006) along the California coast. At 500 m (2220–2239 UTC), both the model and observations had a slight increase in temperature offshore, a decrease in wind speed, and a transition to a more southerly flow toward point E.

4. IOP 7: 12 October 2004

a. Synoptic setting

At the start of the aircraft mission at 1800 UTC 12 October 2004 (Fig. 8a), there was a surface cyclone (982 mb) near the southern tip of the Alaskan Peninsula, while a secondary low (~998 mb) over the central Gulf of Alaska was moving eastward toward the SARJET
study area. As compared with Fig. 10 of Winstead et al. (2006), the MM5 accurately simulated the position and depth of these features to within 1 mb. IOP7 had 4°–8°C colder inland surface temperatures than did IOP1 (cf. Fig. 2a). A 1028-mb surface high over western Canada during IOP7 resulted in an offshore pressure gradient, which forced cold continental air through the coastal mountain gaps.

By 0000 UTC 13 October 2004 (Fig. 8b), the secondary low pressure was within 500 km of the coast and had intensified to 995 mb. The circulation associated with this feature advected ~10°C warmer surface air into the
SARJET study area during the last 6 h. The surface trough at the leading edge of the warm air was about 2 h slow in the model, as revealed by the aircraft time series and buoy 46083 (not shown). Therefore, in order to directly compare the model with the flight-level data, the model analysis was shifted back 2 h for flight 2, such that 0100–0200 UTC was used in the model rather than 2300–0000 UTC.

**b. Aircraft observations and model simulations**

As compared with the King Air profile from 300 to 1000 m MSL for a location ~130 km offshore at 1848 UTC (point E in Fig. 1; see also Fig. 9, thin black lines), the model was 1°–2°C too cool (Fig. 9a). The average relative humidity in this layer was ~90% (not shown), with the model producing a similar moist static stability as observed (0.008–0.015 s⁻¹). The average observed wind speed in this layer was east-southeasterly at ~20 m s⁻¹ (Figs. 9c and 9d). The model winds were 1–2 m s⁻¹ stronger than observed below 500 m MSL and had a small (~10°) easterly wind direction error. Flow-blocking parameters, such as Fr, were not calculated because the low-level flow had an offshore component during flight 1.

During flight 1 (1700–1930 UTC) east-southeasterly flow existed throughout the study region at 150 m MSL (Fig. 10a). The winds were more easterly to the north.

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and south of the Fairweathers, where continental flow accelerated offshore through the coastal gaps. In particular, gap outflow of ~20 m s$^{-1}$ at the exit of the Cross Sound turned anticyclonically and merged with the southeasterly flow 40–80 km offshore. The temperatures were 2°–3°C cooler within this gap outflow than 200 km offshore. In contrast, there was a 3°C warm anomaly adjacent to the Fairweathers at 150 m, which resulted in a cross-shore temperature gradient near the coast that was opposite of IOP1 (cf. Fig. 5). The 4-km MM5 at 1900 UTC 12 October 2004 (Fig. 10b) illustrates more completely the anticyclonic turn-
ing of the cold gap outflow, which surrounds the area of warm air centered near the Fairweathers. A region of enhanced winds (≥20 m s⁻¹) was located between the cold gap outflow and the coastal warm anomaly.

At 500 m MSL (Figs. 10c and 10d), the observed and simulated flow was more coast parallel than at 150 m MSL, with 15–20 m s⁻¹ winds along flight leg C–E'. There was a slight wind enhancement and deflection of the ambient flow as it encountered the region of the gap outflow. The observed temperatures were >11°C at the coast, and decreased to <9°C about 20 km offshore. The model reproduced this temperature gradient, but it was ~1°C too cool.

At 1000 m MSL (Fig. 10e), the observed southeasterly winds of 15–17 m s⁻¹ were 2–3 m s⁻¹ stronger 80 km offshore at point E' than near the coast. The model wind speeds also increased slightly offshore in the region downstream of the tallest peaks, but were fairly uniform upstream (Fig. 10f). The temperatures were also more uniform at this level than below, thus indicating that the gap outflow and warm anomaly near the coast were fairly shallow.

During flight 2 (2300 UTC 12 October–0230 UTC 13 October), the temperature profile for 300–1000 m MSL at point E' was 10–12°C warmer than it was for flight 1 as a result of warm advection from the offshore cyclone (Fig. 9a), while the winds had weakened to 18 m s⁻¹ (Fig. 9e) and veered ~60° to more southerly (Fig. 9d). A similar low-level temperature increase was obtained upstream (south) of the Fairweathers for the region around point U (cf. Fig. 1), where the average moist Fr below 2000 m was ~0.75 (U ~ 15 m s⁻¹, Nₘ ~ 0.01 s⁻¹, and hₘ ~ 2500 m), so flow blocking is expected.

A vertical profile at point C (Fig. 9, heavy gray lines) through the barrier jet core at 0140 UTC 13 October 2004 indicates that the model equivalent potential temperature profile was within 1 K of the observed below 1500 m MSL (Fig. 9a), but was slightly too cool at ~500 m MSL. The temperatures at point E' between 500 and 1000 m MSL were ~4–5 K warmer than those at point C. The stabilities below 500 m MSL were weaker in the warm anomaly region near the coast than offshore at E. The model accurately simulated the jet profile above 400 m MSL, with wind reaching ~27 m s⁻¹ and veering from east-southeasterly near the surface to southerly above 1000 m MSL.

The weak surface trough noted earlier at 1800 UTC (cf. Fig. 8) entered the region and began interacting with the gap outflow at 0100 UTC 13 October 2004. The observed and simulated winds at 150 m MSL became southeasterly and exceeded 25 m s⁻¹ 20–60 km offshore of the coast (Figs. 11a and 11b). The low-level temperatures near the coast were >11°C, coincident with a 5–10 m s⁻¹ decrease in wind speed. Meanwhile, the flow was more south-southeasterly at 15 m s⁻¹ offshore at point E'. The surface temperature gradient was largest in the outer region of the offshore flight leg given the flow confluence between the advancing warm air offshore with the surface trough and the offshore-directed gap outflow. The model suggests that the temperature gradient of ~2°C (100 km)⁻¹ in the offshore portion of the 4-km MM5 simulation (Fig. 10b) increased to ~10°C (100 km)⁻¹ (Fig. 11b) as the trough interacted with the gap outflow. A similar pattern of low-level frontogenesis has been documented for cold gap outflow from the Strait of Juan de Fuca interacting with a landfalling warm front (Doyle and Bond 2001).

At 500 m MSL (Figs. 11c and 11d), the flow exceeded 25 m s⁻¹ 20–80 km offshore, while the winds were ~13 m s⁻¹ offshore near point E'. The coldest temperatures associated with the gap outflow were ~20 km closer to the Fairweathers than at 150 m MSL, while a narrow (~20 km) warm anomaly (+2°C) persisted against the southwest portion of the barrier at 500 m MSL. The model had a similar jet structure between C–E', and it also indicates that the jet became wider and slightly stronger ~50 km farther downstream.

At 1000 m MSL (Figs. 11e and 11f), south-southwesterly flow at 17 m s⁻¹ accelerated to 20 m s⁻¹ and became southerly at the coast, with the simulated winds (Fig. 11f) 2–5 m s⁻¹ too strong. In contrast to the warm anomaly near the coast at 500 m, there was weak cooling toward the coast at 1000 m in the observations, while the model temperatures were more uniform. The deflection of the onshore flow to more terrain parallel and the cold anomaly near the coast are reminiscent of the more classical barrier jet observed in IOP1.

Figure 12 shows cross section C–E' of the observed and simulated winds and potential temperatures for both flights. The flight 1 coastal winds (Figs. 12a and 12b) were 15–20 m s⁻¹ and directed mainly offshore between 500 and 2000 m. The center of the gap outflow is identified by the shallow cold dome below 500 m centered ~80–90 km offshore. There was a strong vertical potential temperature gradient capping the shallow gap outflow at 500 m; thus, the stratification (Nₘ ~ 0.012 s⁻¹) was at least twice as large as for IOP1 (cf. Fig. 6). Meanwhile, below 1 km MSL and at the coast the isentropes tilted downward toward the coast associated with the low-level warm anomaly.

During flight 2 (Figs. 12c and 12d), there was a well-defined barrier-parallel jet, with speeds to 30 m s⁻¹ within 50 km of the coast at 300–500 m MSL. The barrier jet was much weaker above 1 km MSL and the winds were more south-southwesterly. Between 1000 and 2000 m the ambient flow became more southerly
(terrain parallel) only within 10–20 km of the coast. The flow blocking above 1000 m was weaker given the Fr for the box at \( U \sim 1.3 \) (\( U \sim 20 \text{ m s}^{-1} \), \( N_m \sim 0.01 \text{ s}^{-1} \), and terrain height above 1000 m of \( \sim 1500 \text{ m} \)).

Time series for the C–E' flight legs show that the warmest temperatures (\( \sim 12^\circ \text{C} \)) were located at the coast at 150 m MSL (Fig. 13a). The model and observed winds decreased rapidly to 15 m s\(^{-1}\) while becoming southerly approaching point E' (Figs. 13b and 13c). Over the eastern portion of the 500-m flight leg (near 0030 UTC) there were large wind speed variations (21–28 m s\(^{-1}\)) and vertical velocities exceeding \( \pm 1 \text{ m s}^{-1} \). The peaks of these oscillations were separated every \( \sim 5 \text{ km} \), which is suggestive of wave like perturbations at the top of the stable layer capping the gap outflow. The upward (downward) motions at 500 m were asso-
associated with a 2–3 m s\(^{-1}\) wind speed increase (decrease), as higher (lower) momentum was transported from below (above) this level. In contrast to IOP1, there were higher concentrations of rainwater observed for the offshore portion of the flight leg (Fig. 13e), while lower concentrations were observed near the coast at point C. At the 500-m level (Fig. 13), there was an increase in wind speed and temperature, and a decrease in relative humidity toward the coast, but the simulated wind shift was slightly less pronounced.

5. Trajectory analysis

In order to illustrate the origins of the coastal winds and temperature structures for both IOPs, backward trajectories were calculated along cross section C–E. A trajectory time step of 5 min was used with the spatial and temporal interpolation of 15-min model data. The times of release were 2100 UTC 26 September and 0100 UTC 13 October for IOP1 and IOP7, respectively.

Figure 14 shows the trajectories released during the second flight of each IOP at 150 and 500 m MSL. The relatively uniform south-southeasterly winds during IOP1 are evident in all the near-coast trajectories (3–6 in Fig. 14a and 9–12 in Fig. 14b), with all trajectories originating offshore at 150 and 500 m. The parcels released at the coast (Nos. 6 and 12) show a slight deflection to a more east-southeasterly component near the Fairweathers. The greater onshore component for trajectories 1–2 and 7–8 is a result of the weak trough offshore immediately west of the strongest winds (cf. Fig. 5).

The IOP7 trajectories (Nos. 3 and 4) released in or near the jet maximum within 100 km of the coast originate inland (Fig. 14c). Trajectories 5 and 6 follow the coastline, ascend the gap outflow to 1900 m, then descend adjacent to the Fairweathers to create the warm anomaly noted in Figs. 10, 11, and 12. Trajectory 6 experienced a potential temperature increase of \(\sim 6\) K as it ascended over the gap outflow and the southern portion of the Fairweathers from 600 to 2000 m (not shown). There was light to moderate precipitation in this region, so latent heating was likely occurring. Sub-
sequently, trajectory 6 descended from 2000 to 150 m; however, the potential temperature decreased $\sim 3$ K due to evaporative cooling and mixing. Therefore, the net diabatic heating of trajectory 6 was $\sim 3$ K, which was likely important in enhancing the warm anomaly at the coast. The trajectories released at the 500-m level (Fig. 14d) resemble the IOP1 trajectories at this level (cf. Fig. 14b), with offshore origins and confluent flow 100–150 km from the coast. Trajectories 11 and 12 ascend the gap outflow to 700–1300 m and then descend to 500 m at the base of the Fairweathers.

6. Momentum budget

A momentum budget was calculated for both IOPs in order to diagnose the mechanisms for the wind variations near the coast. The following momentum equation was separated into its components:

$$\frac{dV}{dt} = -\frac{1}{\rho} \nabla p + f \times V - F, \quad (1)$$

where $d/dt$ is the total derivative with respect to time and $V$ is the horizontal velocity vector. The first two terms on the right-hand side represent the pressure gradient and Coriolis acceleration, respectively. The last term, $F$, shows the tendencies output from the PBL schemes, which include friction and turbulent mixing. Using this equation, the zonal and meridional momentum balances were output at all grid points within the 4-km MM5 nest every 5 min and averaged for a specified period.
Figure 15 shows the momentum terms plotted at 150, 500, and 1000 m MSL for a representative 30-min period of flight 2 for both IOPs. For IOP1 (2045–2115 UTC), the ambient flow (south of point E in Fig. 15a) is nearly geostrophic at 150 and 500 m MSL (Figs. 15c and 15e). As the parcels approach to within ~50 km of the Fairweathers, they are accelerated to the left by an offshore-directed pressure gradient. Subsequently, they accelerate down the pressure gradient parallel to the Fairweather Mountains, where at 150 m MSL an approximate antitropical balance exists in the along-barrier direction near point C between the pressure gradient force and friction exists (Fig. 15a). After the flow passes the Fairweathers (just offshore of point D in Fig. 15a), an approximate geostrophic balance is obtained in the cross-barrier direction at 500 and 1000 m MSL (Figs. 15c and 15e). The total accelerations are largest near the Fairweathers at 500 m MSL (Fig. 15c). The friction/mixing term is small everywhere except at 150 m, while both the pressure gradient force and the net acceleration terms do not vary much with height within the coastal region.

In contrast, for IOP7 the flow accelerates out of the Cross Sound gap down the pressure gradient and turns to the right in part by the Coriolis force (Fig. 15b). If the Coriolis was the only force, then the flow along the axis of the pressure ridge (dashed line) of the gap outflow would follow an inertial circle, with inertial radius $R_i = U/f$ (Holton 2004). With a gap outflow wind speed of 20 m s$^{-1}$ and $f = 1.25 \times 10^{-1}$ s$^{-1}$, $R_i \sim 160$ km; however, the radius for this trajectory is only ~100 km. This suggests that another force is accelerating the flow to the north. The friction/mixing term has a relatively large component in the gap outflow region to the right of the flow, due to the mixing of southerly momentum into the gap outflow from above. The moist Richardson number $[R_t = N^2(du/dz)^{-2}]$ was ~0.25, where $\Delta U = 20$ m s$^{-1}$ between 150 and 750 m MSL and $N_m \sim 0.005$ s$^{-1}$, thus suggesting the potential for mixing. Unfortunately, there were no observations collected around this loca-
tion, but the aircraft time series between CE suggested areas of significant vertical momentum transport within the gap outflow (cf. Fig. 13). The ambient flow at all levels entering the domain from the south is decelerated ~80 km upstream of the coast as it impinges on the gap outflow (Figs. 15b, 15d, and 15f). The deceleration is caused by the positive pressure perturbation associated with the gap outflow, thus acting to oppose the
impinging flow, which causes the impact of terrain to extend farther upstream than for IOP1. Above the gap outflow (Figs. 15d and 15f), the momentum balances adjacent to the Fairweathers resembles more of IOP1, with the impinging parcels being slightly deflected northward by the terrain-induced pressure gradient. Downstream of the Fairweathers (near point D), the flow at 500 and 1000 m MSL obtains an approximate geostrophic balance in the cross-barrier direction, similar to that in IOP1.

7. Impact of gap outflow

An objective of this study was to determine how the gap outflow influenced the structure and intensity of the barrier jet during IOP7. We examined this by filling Cross Sound gap (hereafter referred to as NOGAP) in the 12- and 4-km domains. The added NOGAP terrain connects the south side of the Fairweathers to Chichagof Island (Fig. 1) and slopes linearly from 2000 to 500 m over a 100-km distance (Fig. 16b). The effectiveness of the artificial ridge in blocking the flow across Cross Sound is shown in a cross section (Y–Y' in Fig. 16a) through Cross Sound. The gap outflow was reduced from \( \sim 15 \text{ m s}^{-1} \) in the control run (CTL) (Fig. 16c) to \(< 5 \text{ m s}^{-1} \) in the NOGAP run (Fig. 16d), while the flow became more southerly (upslope) over the windward slope of the added barrier. However, the isentropes still have an upward tilting toward the coast associated with a \( \sim 2^\circ \text{C} \) cold anomaly, which suggests that at least a portion of this cooling was from adiabatic ascent of the southerly flow over the barrier.

At 0000 UTC 13 October 2004 (Fig. 17), the wind speeds exceed 25 m s\(^{-1}\) for both the NOGAP and CTL simulations; however, the region of enhanced winds extends about 40 km wider (west–east) and about 30 km longer (north–south) for the CTL than NOGAP. The cross sections taken through points X–X' (Figs. 17a and
17b) illustrates that this change in width of the coastal jet is reduced from $\sim 100$ km in CTL (Fig. 17a) to $\sim 60$ km in NOGAP (Fig. 17b). Removing the Cross Sound gap also changes the position of the jet maximum, from slightly offshore to one centered closer to the coast than the CTL, which is similar to IOP1.

A trajectory analysis was completed for the NOGAP experiment (Fig. 18) to diagnose how the NOGAP flow changed the airflow of the hybrid jet. Trajectories 5 and 6 at 150 m MSL (Fig. 18a) still originate along the coast, but trajectories 3 and 4 no longer exit from Cross Sound as in the control run (cf. Fig. 14c). Trajectory 4 originates inland for the NOGAP run as it is deflected around the filled gap, which suggests that a portion of the cold anomaly found in section Y–Y’ in Fig. 16d is a result of upstream leakage of cold continental air. Trajectory 3 originates offshore in NOGAP and experiences a slight deflection as it approached the coastal region. At 500 m MSL (Fig. 18b), the trajectories have a more classical barrier jet signature, similar to IOP1. Overall, these results suggest that the gap outflow through Cross Sound had a large impact on the barrier jet structures during IOP7.

8. Summary and conclusions

Aircraft in situ measurements were collected during the SARJET experiment to investigate the structure

![Fig. 17. Cross section X–X’ showing the (a) CTL and (a) NOGAP simulations of the terrain-parallel wind speeds (gray shaded every 5 m s$^{-1}$) and potential temperature (dashed every 1 K) at 0000 UTC 13 Oct 2004. The location of X–X’ is shown in Fig. 16a.](image1)

![Fig. 18. Backward trajectories for the NOGAP experiment released at 0100 UTC Oct 2004 at (a) 150 and (b) 500 m MSL. The height of the trajectory is proportional to the width of the trajectories (spaced at hourly intervals). Wind speed is gray shaded (every 5 m s$^{-1}$).](image2)
and physical processes of coastal barrier jets along the Fairweather Mountains near Juneau, Alaska. Much of the success of SARJET can be attributed to the detailed observations collected in a variety of conditions favoring strong low-level winds in the coastal zone. This paper presented in situ aircraft data and high-resolution simulations to compare a classical barrier jet (IOP1) with a hybrid jet (IOP7) that had gap flow influences at low levels.

During IOP1 there was south-southeasterly flow preceding a landfalling trough, which became blocked by the coastal terrain and accelerated down the pressure gradient to produce a $5-10$ m s$^{-1}$ wind enhancement (maximum wind speeds $\sim 30$ m s$^{-1}$) in the alongshore direction near the Fairweathers. These features were similar to other classical barrier jet structures studied by Parish (1982), Doyle (1997), and Yeh and Chen (2003), in which the windward pressure ridging and associated cold anomalies were produced from low-level upslope flow. In contrast, IOP7 featured greater surface pressure and colder low-level temperatures to the east (inland) of the study area than did IOP1, which resulted in offshore-directed coastal gap flows below $\sim 500$ m. This event was similar to a barrier jet event near California’s Petaluma Gap that was investigated by Neiman et al. (2006). In their conceptual model, the gap outflows turn to the north, forming a hybrid jet similar to the one found in IOP7.

MM5 simulations were performed to further investigate the airflow through the jet structures and the dynamics governing their morphology. The simulations presented for both events were shown to adequately reproduce the low-level pressure perturbations and orographic flow response. The IOP7 simulation had a small timing bias associated with the approach of a pressure trough and warmer air; however, the model was still able to accurately simulate the sampled hybrid jet structure.

Momentum budget analysis revealed the dynamical differences between the two IOPs. The air parcels in the hybrid jet turned anticyclonically along a smaller radius than an inertial circle as they exited Cross Sound. Farther downstream (to the north), the momentum balance along the coast was more characteristic of a classical jet, with approximate geostrophy in the cross-barrier direction. The flow $\sim 150$ km upstream of the SARJET region in IOP1 was nearly geostrophic, while the upstream flow in IOP7 was undergoing a deceleration as it interacted with the gap outflow.

Model trajectories illustrated that IOP1 only had onshore flow origins, while the coastal winds in IOP7 had both gap and offshore origins. Low-level trajectories in IOP7 originated offshore and were deflected westward by the gap outflow rather than the coastal terrain. To test the impact of the gap flow on the hybrid jet, a simulation was performed with the Cross Sound gap filled (NOGAP). This produced a coastal jet with a similar maximum wind speed to the control run, but the filled gap reduced the offshore width of the coastal jet by about 40% and a shifted the maximum winds more toward the coast, which is similar to the classical jet in IOP1.

Figure 19 presents a conceptual model summarizing the three-dimensional structures of the southeast Alaskan hybrid jet using IOP7 results and those presented in Winstead et al. (2006). For these events the gap outflow rotates anticyclonically out of the coastal gap and merges with the ambient coastal jet adjacent to the steep coastal terrain. Unlike the classical jets (IOP1), there is a warm anomaly near the coast resulting from the downslope flow off the southern end of the Fairweathers (Fig. 19). A cold anomaly exists farther offshore associated with the gap outflow. Above the shallow gap flow at midmountain level, the flow is more representative of a classical barrier jet, with southerly flow deflecting and accelerating more parallel to the Fairweathers, and there is a weak cold anomaly against the barrier. Farther downstream, the hybrid barrier jet structure becomes more similar to a terrain-parallel classical jet.

The coastal jets of the southeastern Alaskan Coast are often influenced by the gap outflows through the numerous sea level mountain gaps in this region. This study will help forecasters recognize the different mechanisms between hybrid and classic barrier jets. A subsequent paper will further quantify the changes of the gap outflow and hybrid jet structures using a series
of idealized simulations with varying ambient flow and stability conditions, as well as the depth of the inland cold anomaly.

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