Numerical Simulation of a Low-Level Jet over Complex Terrain in Southern Iran

MING LIU,* DOUGLAS L. WESTPHAL, TEDDY R. HOLT, AND QIN XU
Naval Research Laboratory, Marine Meteorology Division, Monterey, California

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ABSTRACT

The Lut Desert of Iran is an elongated valley oriented north-northwest to south-southeast. The valley descends southward to the Jaz Murian dry lake through a pass. The Navy’s Coupled Ocean–Atmosphere Mesoscale Prediction System is used to study a northerly low-level jet in the valley and across the dry lake. The dynamics of the jet are investigated with force balance and Froude numbers to determine the contribution of various mechanisms to the jet formation and maintenance. The jet is initiated as a channeled gap flow in the convergent topography of the Lut valley by the valley-parallel pressure gradients generated by the large-scale processes and by the presence of cold air over the valley’s sloping terrain. The pressure gradient is mainly counteracted by the frictional force. The imbalance between them controls the intensity and persistence of the jet in the valley. Farther south, the jet evolves into a downslope flow resembling a hydraulic jump on the steep slope of the dry lake. A transition of subcritical situation to supercritical faster flow is found at the mountain crest between the Lut valley and dry lake. The depth of stably stratified cold layer, the static stability of upstream inversion, and magnitude of upstream winds all determine the jet configuration over the dry lake. The lee troughing over the Gulf of Oman and the Persian Gulf, as the large-scale inland flow crosses the coastal mountains, supports this low-level jet through the increased along-jet pressure gradient. The jet is also influenced by diurnal forcing, being strong at night and weak during daytime.

1. Introduction

Low-level jets (LLJs) occur in many regions of the world and have attracted a great deal of interest in the past decades because of their importance to air-pollutant transport, deep convection activities, cyclogenesis, wind energy production, and aviation safety (Doyle and Warner 1991; Stensrud 1996). In this paper, we discuss a previously undocumented northerly LLJ event occurring in central and southern Iran on 12, 13, and 14 February 1995. The Lut Desert is an elongated confluent valley oriented north-northwest to south-southeast along the foothills of the Zagros Mountain (Fig. 1). It connects with the Kavir Desert basin at a higher altitude in northern Iran and extends southward through a pass to the Jaz Murian dry lake. To the south of the dry lake, there exists another pass to the Gulf of Oman between the Zagros Mountain and the central Makran Range (not apparent in Fig. 1 due to coarse resolution), which otherwise act as barriers to the inland airflow reaching the coast. The topography surrounding the Lut Desert valley and the dry lake likely plays an important role in the development of the LLJ through the interaction of complex topographic forcing and meteorological conditions. The northern portion of the jet appears to be a gap wind in the Lut valley, while the southern portion of the jet appears as a strong downslope wind over the dry lake. The jet probably experiences a dynamic transition as it traverses from the Lut valley southward to the dry lake in responding to different features of surrounding topography.

A number of forcing mechanisms have been proposed or identified to explain the various aspects of LLJs in a wide variety of environments. Inertial oscillation (Blackadar 1957; Bonner 1968; Uccellini 1980; Douglas 1995), shallow baroclinicity (Djuric and Damiani 1980; Forbes et al. 1987; Hsu and Sun 1994; Stensrud 1996), and terrain effects (Stull 1988; Xu 1990; Holt 1996; Doyle 1997) are often seen in case studies to be the major causes of LLJs. Complex terrain impacts airflow in various ways with actions such as downsloping, blocking, and convergent channeling. Under the influence of dynamically or thermally created baroclinicity, topographically induced winds take different forms of LLJs.

The well-known examples include the gap wind LLJs observed in Cook Inlet and Shelikof Strait in Alaska (Lackmann and Overland 1989; Macklin et al. 1990),
in Fraser Gap and the Strait of Juan de Fuca of Washington State and British Columbia (Overland and Walter 1981; Mass et al. 1995), and in Howe Sound of British Columbia (Jackson and Steyn 1994), as well as in the Isthmus of Tehuantepec in Mexico (Steenburgh et al. 1998). These LLJs are highly terrain dependent as the jets are accelerated within the channels (e.g., gaps or straits) by the channel-parallel pressure gradients. Another type of terrain-dependent LLJs are downslope winds dominated by mountain-wave amplification, wave breaking, or hydraulic-jump mechanisms (Klemp and Lilly 1975, 1978; Clark and Peltier 1984; Smith 1985; Durran 1986, 1991; Durran and Klemp 1987). In such downslope events, strong winds are generated by mountain gravity waves amplified by vertical reflection due to the existence of a critical or near-neutral layer, or generated in an analog to hydraulic jumps for stratified fluids. As a location of flow reversal, a critical layer can also result from wave breaking through locally self-induced processes. The Front Range of Colorado on the western slopes of the Rocky Mountains is famous for its frequent damaging downslope windstorms (Brinkmann 1974; Klemp and Lilly 1975; Scheetz et al. 1976; Oard 1993; Kapela et al. 1995) studied with both mountain wave and hydraulic theories.

These previous case investigations and theoretical or modeling studies were all conducted for either gap wind events or for downslope winds. The LLJ in the Lut valley and the dry lake carries the features of both gap winds and downslope winds in three-dimensional complex terrain. It bears some similarity to the wind system that occurred in the Stampede Gap and on the lee side of the Cascade Mountains, Washington State, on 12 February 1995 (Colle and Mass 1998a,b). The topography in both the Lut valley along with the dry lake and the Stampede Gap exerts combined channeling and sloping effects on the airflows. In both cases, the low-level gap and downslope winds are within cold air masses. How-

**Fig. 1.** Terrain height (m) field from the 81-km grid mesh showing geographical features of the Lut Desert valley, the Jaz Marian dry lake, and the surrounding areas.
ever, the temporal and spatial scales are different. The LLJ in this study occurred in an area of approximately 1000 km by 400 km and lasted for a period of three days; while the wind event investigated by Colle and Mass was in a much smaller region and lasted for only one day. Therefore, we expect different magnitudes of topographic and dynamic forcing. In addition, the three-day LLJ in our case also bears some diurnal variation in its vertical structure.

The LLJ in the Lut Desert valley and the Jaz Murian dry lake is worthy of study because it may cause hazardous operating conditions in the region in terms of turbulence and dust mobilization. The strong winds can mobilize dust in the inland deserts and transport it to the Gulf of Oman, severely restricting the visibility in the area (Walters and Sjoberg 1988). Some evidence of dust was found during the Ship Antisubmarine Warfare Readiness Effectiveness Measurement (SHAREM) experiment, conducted in February 1995 in the Persian Gulf and the Gulf of Oman. The radiosondes released from the ships in the Gulf of Oman from 13 to 15 February detected layers of unusually dry air between 800 and 2500 m (Byers 1995). In coincidence with the dry layers, dust streaks were seen over the area in the satellite imagery retrieved from National Oceanic and Atmospheric Administration Advanced Very High Resolution Radiometer satellite data on 14 February (Teadt 1996). On the same day, the British C-130 research aircraft used in the SHAREM experiment encountered haze layers below 1300 m in the Gulf of Oman. Similar dust streaks have been found and reported by Perrone (1979) during a postfront period in January 1973 in the same area. The dust streaks appear to emanate from the deserts of Iran, Afghanistan, and Pakistan. Electromagnetic and optical propagation can be dramatically degraded in the presence of dust and haze (Condray and Edson 1988). A dust storm, strong winds, and severe turbulence in the same region of Iran under similar synoptic situations were significant causes for the failure of the Iranian hostage rescue mission in 1980 (Thomas 1987).

In this paper, the dynamics of the LLJ and its interaction with the surrounding topography are studied. Observations in the region are insufficient to detect the mesoscale nature of the jet, so we use the U.S. Navy’s Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS) numerical model to investigate the event. We first give a brief overview of the model. We then discuss the synoptic and mesoscale conditions of the LLJ and verify the numerical results. We use the modeled fields to analyze the force balances for the gap winds in the Lut valley and apply hydraulic-jump theories to the downslope winds in the dry lake in order to identify various forcing mechanisms responsible for the initiation and maintenance of the jet.

2. Model description

The model used for this LLJ case study is the navy’s Coupled Ocean–Atmosphere Mesoscale Prediction System. It is a nonhydrostatic, compressible dynamics model applied in a terrain-following sigma vertical coordinate. It predicts turbulent kinetic energy (TKE) for subgrid-scale turbulent diffusion, uses a force–restore method in the surface energy budget, and contains explicit cloud microphysics. The complete details of the model structure, dynamics, and physics are described by Hodur (1997). The data assimilation is performed at 12-h incremental update cycles (Baker 1992; Barker 1992). The analysis and forecast fields of the navy’s operational global circulation model (Hogan and Rosmond 1991) are used as the regional model’s initial state and to update the lateral boundary conditions every 6 h.

The model domain extends vertically to 35 km with 30 expanding grid layers ranging from 20 m thick at the surface to 7.5 km thick at the top. High resolution near the surface achieves better simulation of the planetary boundary layer structure. In the horizontal, the model is single nested by using a 63 × 61 gridpoint coarse mesh with 81-km grid spacing and an 82 × 67 relatively fine mesh with 27-km resolution. These meshes capture both the synoptic and the mesoscale systems related to the LLJ over the Lut Desert valley and the dry lake.

The model was run from 6 to 16 February 1995, generating overlapping 24-h forecasts every 12 h starting at the 0000 and 1200 UTC analysis times. The first two days of simulation enable the model to reach dynamic equilibrium and develop mesoscale features necessary for realistic predictions. Consecutive 12-h through 24-h forecast fields, rather than the analyses or short-term forecasts, are used in this paper to study the jet event. The purpose of using longer-term forecasts is to evaluate the potential for the COAMPS to predict the synoptic forcing that leads to the LLJs in the area. In section 4 on model validation, COAMPS 12-h forecasts compare well with observational data and capture the synoptic weather features for the study period. Observations in the Lut valley and the dry lake are insufficient to verify the capability of COAMPS to predict the mesoscale features of the jet. Instead, it must be inferred from the favorable comparison of the simulation with the theoretical analyses (sections 6 and 7) and the presence of dust over the Gulf of Oman on 14 February.

3. Synoptic-scale conditions

Figure 2 shows the COAMPS 12-h forecasts (81-km grid mesh) of geopotential height on the 500-mb surface from 10 to 15 February with observed winds. The 850-mb surface and sea level pressure for the same times are shown in Figs. 3 and 4. A deep 500-mb trough east of the Caspian Sea stretches southwestward to the Red Sea on 10 February (Fig. 2a). Behind the trough there is a cold center showing cold advection by northwesterly winds. The corresponding surface high is over Iraq and the Arabian Peninsula (Fig. 4a). This upper-air disturbance and the low-level northwesterlies combine to
bring cold air down from the north. The southward cold air intrusion can be seen in the 850-mb temperature field (Fig. 3a). The high northwesterly wind event over the Persian Gulf at this time is called Shamal, an Arabic word meaning north. Shipboard-measured 10-m winds in the Persian Gulf during SHAREM exceeded 10 m s$^{-1}$ continuously for 3 days from 9 to 11 February (Byers 1995). As the upper-level trough progresses eastward (Fig. 2b), a new surface high moves to the Caspian Sea and the Iran Plateau on 12 February north of the Lut Desert valley (Figs. 3b and 4b), and the surface low becomes stationary near the border of Pakistan and India. From the low center, a deep surface trough extends westward to the Gulf of Oman and the Persian Gulf.

The trough results from the effect of mountain lee troughing as the large-scale northeasterly inland flow crosses the Zagros Mountains and the central Makran Range. Both Brody (1977) and Perrone (1979), from their climatological and case studies, found that the mountain ranges parallel to the gulf coastline, combined with the warm seawater as a surface heat source, make the gulf area favorable for surface low formation or cyclogenesis. It is clear that in the lower troposphere on 12 and 13 February (Figs. 3c and 4c) the Lut Desert valley and the dry lake are under the influence of a synoptic southward pressure gradient force. Such a circulation pattern creates a prerequisite meteorological condition for the formation of a northerly LLJ. The Lut

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Fig. 2. The 12-h forecasts of geopotential height (m) (solid lines) and temperature (K) (dashed) on the 500-mb level using the 81-km grid and observed wind barbs (a full tick = 10 m s$^{-1}$) at times indicated.
valley and the dry lake lie along the eastern foothills of the Zagros Mountains. The mountains can block the low-level easterly airflow that is driven by the synoptic north–south pressure gradient and force the airflow to turn to the south along the mountain foothills. During the jet occurrence on 12, 13, and 14 February, the 500-mb winds are weak and westerly over the area (Figs. 2b and 2c). At 850 mb and at the surface (Figs. 3c and 4c), a strong lee trough remains in the south and a high pressure area remains in the north thus maintaining the circulation. Figure 3c also shows a pool of cold air mass occupying central Iran. By 15 February, a warm upper-air ridge has replaced the previous trough over the Iran Plateau (Fig. 2d). The surface highs in the Iran Plateau and Saudi Arabia both weaken, and the lee trough in the gulf area fills (Figs. 3d and 4d). As a result, the surface pressure gradient over Iran relaxes. This coincides with the decay of the northerly LLJ.

4. Modeling validation
In the above section, the 12-h forecasts of the 81-km grid mesh have been used to describe the synoptic situations of the LLJ since there are too few upper-air and surface observations in the model domain and none in the study area for analysis. Before using the modeled fields to study the dynamics of the LLJ, it is necessary to validate the COAMPS synoptic forecasts. Close inspection of Figs. 2 and 3 reveals good agreement between the observed winds and the 12-h forecast synoptic
height fields. A quantitative evaluation also verifies COAMPS’s skill in predicting the synoptic circulation related to the LLJ. The evaluation makes use of 81-km 12-h forecast fields of geopotential height and wind speed on 850-, 700-, and 500-mb levels, as well as sea level pressure. It consists of two statistical quantities: root-mean-square (rms) and bias score errors against the available observational data. The rms is calculated as

$$\sqrt{\frac{1}{N} \sum_{N} (F - O)^2}$$

and the bias score defined as

$$\frac{1}{N} \sum_{N} (F - O),$$

where $O$ is the observation, $F$ the forecast, and $N$ the total number of points involved. The bias represents the average modeling deviation from the observation at a valid time. The average number of observations at each analysis time covered in this evaluation is about 60 radiosondes and 440 surface stations. The approximate distribution of 45–50 radiosondes, which had both observed winds and geopotential heights, can be found in Figs. 2 and 3, while the surface observations are too dense to be shown in Fig. 4. The forecast fields are bilinearly interpolated to these observation sites, and then the differences with the observational data are calculated for the rms and the bias score. Tables 1 and 2 list the statistics averaged over the time period covering the LLJ event, that is, from 0000 UTC 12 February to
May 2000 1315Liu et al.

Table 1. Root-mean-square and bias errors for 12-h forecasts of sea level pressure (mb) and geopotential height (m) on 850-, 700-, and 500-mb surfaces, averaged from 0000 UTC 12 Feb to 0000 UTC 15 Feb.

<table>
<thead>
<tr>
<th>Pressure (mb) on sea level</th>
<th>Height (m) on 850 mb</th>
<th>Height (m) on 700 mb</th>
<th>Height (m) on 500 mb</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rms</td>
<td>2.595</td>
<td>12.628</td>
<td>15.059</td>
</tr>
<tr>
<td>Bias</td>
<td>0.966</td>
<td>3.847</td>
<td>0.940</td>
</tr>
</tbody>
</table>

Table 2. Root-mean-square and bias errors for 12-h forecasts of wind speeds on the sea level, 850-, 700-, and 500-mb surfaces, averaged from 0000 UTC 12 Feb to 0000 UTC 15 Feb.

<table>
<thead>
<tr>
<th>Winds (m s^-1)</th>
<th>Winds (m s^-1)</th>
<th>Winds (m s^-1)</th>
<th>Winds (m s^-1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>at surface</td>
<td>on 850 mb</td>
<td>on 700 mb</td>
<td>on 500 mb</td>
</tr>
<tr>
<td>Rms</td>
<td>2.364</td>
<td>2.193</td>
<td>2.464</td>
</tr>
<tr>
<td>Bias</td>
<td>0.952</td>
<td>0.044</td>
<td>-0.029</td>
</tr>
</tbody>
</table>

0000 UTC 15 February. Compared with similar evaluations conducted for the National Meteorological Center’s (now known as the National Centers for Environmental Prediction) regional forecast models (Junker et al. 1989), these errors are smaller than those of typical operational weather prediction. As for the wind speeds, both the rms and bias score are small considering typical observed wind speeds on each level.

Two radiosondes are picked to directly compare with the model output from the 81-km grid mesh. One station is Mashhad, Iran, at (36.30°N, 59.60°E) and north of the LLJ and the other Muscat, Oman, at (23.60°N, 58.30°E) and south of the LLJ (see points G and H in Fig. 5 for the locations). These two are the closest available radiosonde stations to the LLJ area, and their spatial distance from the north to the south represents the synoptic conditions of the LLJ. The 12-h forecasts at 0000 UTC 15 February. Compared with similar evaluations conducted for the National Meteorological Center’s (now known as the National Centers for Environmental Prediction) regional forecast models (Junker et al. 1989), these errors are smaller than those of typical operational weather prediction. As for the wind speeds, both the rms and bias score are small considering typical observed wind speeds on each level.

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UTC 13 February are bilinearly interpolated to these two points and tested against the observations (Figs. 6 and 7). Both modeled temperature and dewpoint depression profiles agree very well with the soundings. On the average, the wind vectors match each other satisfactorily, particularly the weak midtropospheric winds at Mashhad. We conclude that COAMPS forecasts capture the synoptic features for the study period. This gives us confidence in the simulation of the LLJ.

Note that the 12-h forecasts have been used for the validation in this section, since the forecasts will be used in the following LLJ analyses. The validation using the COAMPS analyses (i.e., after the data assimilation cycle) would have been more favorable, yet less relevant.

5. LLJ characteristics

In examining the numerical results from the 27-km grid mesh for the two-week period, it is found that the jet occurs on 12, 13, and 14 February. Prior to and after this period, the local winds are primarily southerly or weak northerly. The structure and persistence of the LLJ for 12–14 February are shown in Fig. 8 in a sequence of vertical, cross-valley sections (along the line A to B in Fig. 5) of the along-valley wind component. The jet is contained in the lowest 800 m and the jet core is 300 m above the ground with a speed varying from 10 to 20 m s$^{-1}$. The intensity of the jet core bears some diurnal variation, being strong and shallow at night and weak but deep during daytime due to turbulent mixing. Local time is 3 h ahead of UTC. Figure 9 shows similar vertical cross sections across the northern slope of the dry lake (along the line from C to D in Fig. 5) at 1800 UTC 12, 13, and 14 February. The jet at this location is stronger because of the steeper terrain slope, but is shallower and wider than that in the valley (Fig. 8) due to the fanning effect by the absence of constricting mountains. The west–east mountain range separating the Lut valley and the dry lake (Figs. 1 and 5) acts as the boundary where the LLJ changes from a gap flow to a downslope flow. Figure 10 is a vertical cross section of wind speed along the centerline of the valley and the dry lake (north–south line in Fig. 5) at 0000 UTC 14 February. The spatial scale of the jet stretches nearly 1000 km from the upper valley to the dry lake. The winds exceed 10 m s$^{-1}$ in the lowest kilometer from the Kavir Desert to the Gulf of Oman with the maximum winds 20 m s$^{-1}$ in the valley and 26 m s$^{-1}$ on the steep northern slope of the dry lake basin. Wind directions are mainly northwesterly to northerly following the topography of the valley and the dry lake. During the whole event, the LLJ reaches a maximum gap wind of 20.4 m s$^{-1}$ at 0000 UTC 14 February in the middle of the valley and a maximum downslope wind of 28.8 m s$^{-1}$ at 0300 UTC 14 February on the northern slope of the dry lake basin.

Figure 11 shows the 12-h forecast wind vectors 10 m above the ground along with the available observations at 1200 UTC 13 February. The modeled winds follow the topography of the Kavir Desert, the Lut Desert valley, and the dry lake. The wind directions and speeds generally match the few nearby observations. Note that in southern Iran, the onshore flow from the Gulf of Oman in the afternoon reaches only a short distance inland because it is met and weakened by the jet’s southward flow. A similar pattern is seen in southern Pakistan. In contrast, the onshore flows (a combination of sea and mountain breezes) reach far up the Zagros Mountains (along the east shore of the Persian
Gulf) and the coastal mountains of Oman (along the south shore of the Gulf of Oman). At night, the jet reinforces the downslope land breeze at the coast (not shown).

It will be shown in the next section that the cold air supply is a key factor in the formation of the LLJ in terms of pressure gradient and negative buoyancy over sloping terrain. Examining the numerical simulations for 12 and 13 February, strong low-level cold advection is evident at the surface and 850-mb level resulting from the cold pool of air filling the Kavir Desert basin (an upwind elevated plateau) on 12 and 13 February. The cold air is channeled into the Lut valley and accelerates down the valley, through the pass and across the dry
Fig. 9. Same as Fig. 8 but crossing the dry lake along the line from C to D in Fig. 5.

lake. The cold advection can be seen in Fig. 12 as winds in the valley blow from the north to south, up the temperature gradient. Similar advection also occurs in the desert valleys of Afghanistan and Pakistan. The cold air structure is further illustrated in Figs. 13a and 13b with vertical cross sections of potential temperature at 0000 UTC 14 February along A±B and C±D in Fig. 5 (across the Lut valley, Fig. 13a, and across the northern slope of the dry lake, Fig. 13b, respectively). The potential temperature contours in the valley tilt upward on the west side due to the Coriolis force associated with the northerly jet. This Coriolis force causes cold air to pile up in the west and, therefore, the temperature contours tilt. The tilting isentropes generate a local cross-valley pressure gradient to balance the Coriolis force as the result of geostrophic adjustment (Overland and Walter 1981; Lackmann and Overland 1989). The topography of the dry lake is different from that in the valley, as is the potential temperature structure. Without the constricting mountains, the cold pool is shallower and wider and there is little tilting. Comparing the temperature fields prior to and after the LLJ period (not shown), the strong low-level inversion 100–300 m above the ground in Figs. 13a and 13b is caused by the cold air advection, which is maximized at the jet core. This vertically differential cold advection leads to destabilization of the atmosphere at low levels in the upstream Lut valley (Fig. 13a). There is enhanced TKE in that unstable region due to buoyancy and strong wind shear. The turbulent mixing works to remove the instability altogether or minimize it. Near the surface, the effect of nocturnal radiational cooling is apparent.

In the following two sections we will study the dynamics of the LLJ in the valley and the dry lake separately, since their topographic features interact with meteorological conditions differently and result in different forms of the jet, that is, the gap jet and the down-slope jet. The forcing mechanisms behind these two forms of the jet, as well as the transition from one form to another, are investigated.

6. Gap wind LLJ diagnosis in the Lut valley

The LLJ in the Lut valley is a gap wind, defined as a channeled airflow that accelerates under a gap-parallel pressure gradient (Overland and Walter 1981). The elongated Lut valley confines the air and channels the airflow southward over its gently sloping floor. In order to understand the gap LLJ dynamics better, it is necessary to examine individual terms in the momentum equations for the along-jet and the cross-jet components of winds. These momentum terms are calculated in the COAMPS terrain-following coordinate:

\[
\frac{\partial u}{\partial t} = -u \frac{\partial u}{\partial x} - v \frac{\partial u}{\partial y} - w \frac{\partial u}{\partial z} + f v + D_x - C_x \theta \left( \frac{\partial \pi'}{\partial x} + \frac{\partial \sigma}{\partial x} \frac{\partial \pi'}{\partial \sigma} \right)
\]

and

\[
\frac{\partial v}{\partial t} = -u \frac{\partial v}{\partial x} - v \frac{\partial v}{\partial y} - w \frac{\partial v}{\partial z} - f u + D_y - C_y \theta \left( \frac{\partial \pi'}{\partial y} + \frac{\partial \sigma}{\partial y} \frac{\partial \pi'}{\partial \sigma} \right)
\]

where the vertical coordinate \( \sigma = z(z_s - z_i)(z_s - z_i) \), \( z_s \) is the depth of the model domain, and \( z_i \) the terrain height. In (1) and (2), \( x \) is the cross-jet direction, \( y \) the direction parallel, but opposite to the jet; \( u, v \) and \( w \) are the components of winds in \( x, y, \) and \( \sigma \); \( f \) is the Coriolis force parameter; \( D_x \) and \( D_y \) represent the effects of horizontal and vertical turbulent mixing for \( x \) and \( y \) components. The first three terms on the right-hand side of (1) and (2) are the momentum advection, followed by the Coriolis force and the friction force due to turbulent mixing. The last term is the pressure gradient, in which
$C_p$ is the specific heat at constant pressure, $\theta_v$ virtual potential temperature; and $\pi'$ is the perturbation of Exner function \[ \pi' = (p/p_0)^{\gamma/c_p}, \] with $p_0$ a constant reference pressure and $R_d$ the gas constant for dry air. The computation is made using the 12-h forecast data for 12, 13, and 14 February during which the jet occurs.

Figure 14 shows the horizontal force and wind vectors in the Lut valley and part of the dry lake 200 m above the ground at 1800 UTC 12 February. It can be seen that the acceleration (local tendency plus advection) at the valley entrance is toward the east because of the large eastward component of pressure gradient there. The acceleration vector eventually turns to the south as the jet progresses southward along the valley. The along-jet acceleration reaches a maximum at 31°N, that is, the middle of the valley where wind speeds also reach a maximum (also see Fig. 10). In other studies, gap winds increase along a flat channel and reach a maximum at the exit of the channel (e.g., Overland and Walter 1981). As mentioned above, the Lut valley has sloping terrain, so the jet accelerates down the valley and reaches its maximum in the middle of the valley before it begins to move upslope with the conversion of kinetic energy to potential energy.

In order to closely examine the force structure within the jet, we choose a location in the middle of the valley at (31°N, 58°E, i.e., point I in Fig. 5) for a detailed dynamical analysis. The time series of the along-jet momentum terms of (1) at this central point are shown in Fig. 15 at the average height of the jet core (i.e., 321 m above the ground). The pressure gradient, advection, and friction terms are the dominant forces in the along-jet direction. The balance (or imbalance) among them controls the intensity and persistence of the jet. The pressure gradient is always down the valley (negative values) until 14 February. It is the strongest in the early stage as the leading edge of the cold air surge passes the location. The advection is mainly in the along-jet direction since the wind components in the cross-jet and vertical direction are all very small. The advection is down the valley for most of the period, except at 0000 and 0300 UTC of 13 February and 0000 UTC of 14 February (local early morning time). As northerly momentum is advected to the location (a maximum at 1800 UTC on 12 February), the local acceleration is enhanced, especially at the beginning of the event in combination with the pressure gradient. It becomes positive only in early morning local time, indicating the influence of the nocturnal drainage that is stronger downstream of point I. The friction opposes...
FIG. 11. Wind vectors 10 m above the ground using the 12-h forecast of 27-km grid valid at 1200 UTC 13 Feb and observed wind barbs (a full tick = 10 m s^{-1}). The shaded background is topography.

the pressure gradient and advection, and is strong during daytime due to the effect of thermally generated turbulent mixing. The friction force also reaches extremes at local midnight of 13 and 14 February (e.g., 2100 UTC 12 and 13 February), since the jet becomes intense at night, as do the wind shear near the jet core and the mechanically generated turbulence. As expected, the along-jet Coriolis force is insignificant because the cross-valley flow is blocked by the mountains on the sides of the valley.

The force balance in the cross-jet direction at the same location (Fig. 16) is nearly in geostrophic balance between the pressure gradient, generated by local sloping cold air isentropes (Fig. 13a) and by the large-scale circulation, and the Coriolis force associated with the northerly jet. This balance is consistent with the result of the scale analysis made by Overland (1984) and Xu et al. (2000) for gap winds. Overland showed that when the width of a gap (l) is much less than its length (L) and the cross-gap Rossby number $R_L (=Vl/fL^2)$, where $V$ is the magnitude of the along-gap flow, is much less than 1, the mass field in the gap adjusts so that the cross-gap pressure gradient balances the Coriolis force associated with the along-gap wind component. In our case, $l = 150$ km, $L = 600$ km, $f = 10^{-5}$, and $V = 3$ m s^{-1}, so $R_L = 0.1$. At night the jet is stronger, as is the Coriolis force, but advection joins the pressure gradient to maintain the momentum balance against the cross-jet Coriolis force.

The momentum budget calculations above at point I (Fig. 15) have revealed that the component of the pressure gradient in the along-jet direction is a dominant driving force of the LLJ in the valley. It is of great interest in this dynamical study to further identify the contribution of each physical mechanism to this driving force. Following Mahrt (1982) and Xu et al. (2000), the
along-jet pressure gradient force $\mathbf{P}_{\text{total}}$ can be decomposed into three components:

$$
\mathbf{P}_{\text{total}} = \mathbf{P}_{\text{large}} - g(\Delta \theta/\theta) \sin \alpha - g(\Delta \theta/\theta) \cos \alpha \left( \partial h/\partial y \right),
$$

where $y$ is the direction parallel (but opposite) to the jet, $\alpha$ the terrain slope, $h$ the depth of the cold layer, $g$ gravity force; $\Delta \theta$ the bulk potential temperature difference between the air at the jet center and the air above the cold pool, and $\theta$ the average potential temperature for the bulk layer. On the right-hand side of (3), the first term $\mathbf{P}_{\text{large}}$ represents the large-scale pressure gradient resulting from the low pressure in the Persian Gulf and the Gulf of Oman in the south and the high pressure over the Iran Plateau in the north. The second term is caused by cold, stably stratified air over sloping terrain in the valley, and the third is due to the variation in cold-layer depth in the along-jet direction. The last two terms are also given short notations $\mathbf{P}_{\text{slope}}$ and $\mathbf{P}_{\text{depth}}$ respectively, and (3) is rewritten as

$$
\mathbf{P}_{\text{total}} = \mathbf{P}_{\text{large}} + \mathbf{P}_{\text{slope}} + \mathbf{P}_{\text{depth}} + \mathbf{P}_{\text{error}},
$$

where $\mathbf{P}_{\text{error}}$ is an implicit term representing the computational uncertainty in the evaluation of (4).

The pressure gradients in (4) are evaluated at point I with a gentle slope ($\alpha = 0.05^\circ$) in the middle of the valley. Here, $\mathbf{P}_{\text{large}}$ is computed from the 81-km-grid pressure field in the free atmosphere above the cold jet (e.g., 1671 m above the ground) to avoid the influence of local topography, and $\mathbf{P}_{\text{total}}$ from the 27-km grid at the average jet-core elevation (321 m above the ground). To eliminate the uncertainty in direct estimation of cold pool depth from the modeled temperature field, $\mathbf{P}_{\text{depth}}$ is estimated as a residual term from (4).
all the terms in (4) along with $\Delta \theta/\theta$ at three typical times of night and day. During the first night, $PG_{\text{depth}}$ contributes the most to $PG_{\text{total}}$, since the cold layer is deeper in the upper valley as the cold front just intrudes into the valley, creating a down-jet pressure gradient. By the next daytime, the cold air has occupied the valley. Turbulent mixing reduces the static stability and makes less variation in cold-layer depth, so $PG_{\text{depth}}$ becomes small and trivial. Term $PG_{\text{slope}}$ contributes to $PG_{\text{total}}$ at all times since point I is on the descending slope. It is larger at night and smaller during daytime caused by the diurnal variation in the static stability of the cold air; $PG_{\text{large}}$ is consistently negative, meaning high pressure in the north and low in the south, but it is smaller than $PG_{\text{slope}}$. Thus the LLJ in the Lut valley is not only supported by the large-scale circulation, but also by locally generated mesoscale forcing due to the sloping terrain and the cold air advection.

7. Downslope wind analysis in the dry lake

The configuration of the topography surrounding the Lut valley and the dry lake makes the LLJ dynamics change from a gap flow in an elongated channel to a downslope flow on a leeside slope when the airstream crosses over the west–east mountain dividing the Lut valley and the dry lake. The mechanism inducing the
downslope flow and the dynamic transition in the LLJ are analyzed in this section.

It is clear from Fig. 10 that strong downslope winds occur on the steep northern slope of the dry lake basin. This event lasts for three days with maximum speeds of 20–28 m s$^{-1}$ (e.g., Fig. 9). Figure 17 is a similar cross section but of potential temperature. Both the wind and temperature fields display some favorable meteorological conditions for the development of downslope winds in the dry lake according to the observational studies of Colson (1954) and Brinkmann (1974). These are (i) the wind is directed across the mountain range and the wind speed at the mountain crest exceeds a value 7–15 m s$^{-1}$ and (ii) the upstream temperature profile in the Lut valley exhibits enhanced static stability extending southward to the mountaintop.

There are basically three different mechanisms proposed in the literature to account for strong downslope winds (Durran 1986, 1991). The first mechanism is based on the shallow-water hydrostatic theory (Long 1954). This theory states that strong winds occur along the lee slope, followed by a subsequent hydraulic jump, when the fluid undergoes a dynamic transition from subcritical upstream to supercritical flow over the mountain crest (Durran 1986, 1991; Durran and Klemp 1987). This framework uses the Froude number (Fr), which is the ratio of the fluid velocity to the horizontal propagation speed of the longest gravity wave, to define subcritical flow ($Fr < 1$) and supercritical flow ($Fr > 1$). The second is based on a theory of internal gravity wave amplification (Klemp and Lilly 1975, 1978), in which downslope windstorms are produced by large-amplitude vertically propagating waves reflected by variation of vertical static stability or by the critical levels of flow reversal. The third is based on a theory of wave breaking amplification (Clark and Peltier 1984; Smith 1985). Growing mountain waves can break under certain circumstances in the upper levels and create a region of strong turbulent mixing or wind reversal that acts as a critical layer to trap wave energy. Durran and Klemp (1986) and Durran (1991) presented numerical evidence suggesting that the second mechanism of wave propagation plays only a minor role in most downslope windstorms, and that there is a fundamental similarity between the downslope winds underneath wave breaking critical levels and within shallow-water hydraulic jumps. Both types of stratified fluid systems undergo a dynamic transition from subcritical to supercritical status.

We focus on one particular time 0000 UTC 14 February to investigate the downslope winds in the dry lake. This time is in the later stage of the jet occurrence and the jet is close to steady state. Figure 17 clearly shows a layered temperature structure. A stably stratified cold layer of approximate 2-km thickness is located in the lower atmosphere ($\theta > 296$ K), and a less stably stratified layer is aloft. There is a well-mixed turbulent region with low Richardson number, 2–3 km above the mountain and upstream, between the two layers. The well-mixed temperature field and the weak and reversed cross-mountain winds (not shown) in this region char-

![Fig. 17. Vertical cross section of potential temperature (K) contours of the 12-h forecast of 27-km grid at 0000 UTC 14 Feb along the centerline of the Lut valley and the dry lake (Fig. 5). Thick contour (296 K) is the isentrope dividing different static stability layers. Asterisks (*) mark the locations of the Froude number calculations. The shaded area indicates cold air has been blocked by the mountain. The lower coordinate is distance (km) and the vertical altitude (km).](image-url)
acterize a critical layer. Underneath the critical layer, the isentropes from the upstream Lut valley descend on the lee slope until they reach the mountain base, then rise again downstream over the dry lake in an abrupt transition that bears striking visual resemblance to a hydraulic jump.

Following Smith’s (1985) nonlinear theoretical study and the numerical studies by Durran and Klemp (1986) and Durran (1991), some quantities are computed to qualitatively characterize the downslope winds over the dry lake. These quantities are the depth and static stability of upstream cold air inflow, the mountain height, and the Froude number along streamlines across the mountain. In Fig. 17, the Lut valley is chosen as the place that holds the less-disturbed upstream inflow conditions for the downslope winds. The depth of the stratified cold layer \( H_0 \) in the valley is about 2 km deep. The mountain height \( h_m \) on the windward side is about 500 m. In Smith’s theory, for every given upstream-inflow depth \( H_0 \) combining with the given upstream static stability and winds, there is one corresponding maximum mountain height that allows the upstream air mass to cross without blocking. According to Smith’s analytical results (Figs. 2 and 5 in Smith 1985), our calculations suggest that a mountain with \( h_m = 500 \) m would block the low-level cold air in the valley for the given \( H_0 \) (2000 m) in this case. By trial and error, we find that a hypothetical mountain height of 200 m on the windward side above the blocking is the proper value to match Smith’s theory. That means we have to assume that the cold air within 300 m above the ground is blocked in the valley, which is the shaded area in Fig. 17, while the air aloft can be transported to the dry lake across the elevated 200-m high mountain. Therefore, the inflow depth of the cold layer \( H_0 \) is reduced to 1700 m just above the shading area in Fig. 17. Using the vertically averaged stability parameter \( N_0 = \left[ (g/\theta)(\partial \theta / \partial z) \right]^{1/2} = 0.014 \) s\(^{-1}\) and the horizontal wind speed \( U_0 = 8 \) m s\(^{-1}\) in this layer, results in a vertical wavenumber \( k = N_0/U_0 = 0.00175 \) m\(^{-1}\) and a vertical wavelength \( L_z = 2\pi/k = 3600 \) m. These wave parameters are used in the following analysis.

Following Smith, we obtain the normalized inflow depth \( H_z k = 5.7 \pi/6 \) and the normalized mountain height \( h_z k = 0.35 \). According to Smith’s theory, for each given upstream altitude of critical level \( (H_z k) \), there exists a maximum allowed mountain height \( (h_z k) \) for the airflow to transit from subcritical to supercritical condition. Based on the curves in Figs. 1 and 5 of Smith, our estimated values of \( H_z k \) and \( h_z k \) are within the parameter range of downslope winds. According to Durran and Klemp (1986), when the upstream critical level or the depth of stratified inflow is located within 0.25 to 0.75 of the vertical wavelength \( (H_0/L_z = 0.47 \) in our case) and the normalized mountain height exceeds a threshold of 0.2 \( (h_z k = 0.35 \) in our case), intensive wave amplification between the surface and the downstream critical level becomes possible to enhance the downslope winds.

We adapt Smith’s dividing streamline concept in this downslope analysis. A dividing streamline or a dividing isentrope symbolically represents a critical level that separates the stratified transitional flow below and the less stratified, turbulent weak winds above, so it is actually an analog to the upper boundary of a shallow water layer in a hydraulic system. In Fig. 17, the 296-K isentrope is the best representation of such a dividing streamline. The Froude number defined by Durran and Klemp (1987) based on Smith’s theory for a nonlinear, continuously stratified system

\[
F = \frac{\sqrt{2U_0u_h}}{N_0\delta} \frac{1}{\sin\left(\frac{1}{2\delta}\right)}
\]

is calculated along the dividing isentrope, where \( u_h \) is the wind speed near the surface or the base of the stratified cold layer and \( \delta \) is the depth of the cold layer underneath the dividing isentrope. The Froude number is calculated for the five vertical columns below the dividing isentrope marked with asterisks (*) in Fig. 17. Two of them are upstream from the mountain crest, one is exactly over the mountain crest, and the other two are downstream over the leeside slope. The Froude numbers at these points along the dividing isentrope from the upstream to the downstream at 0000 UTC 14 February are 0.40, 0.42, 1.00, 1.51, and 0.62. The Froude number changes from \( F < 1.0 \) (subcritical) to \( F > 1.0 \) (supercritical) as the stratified flow traverses the mountain and the transition occurs at the mountaintop with \( F = 1.0 \). At steady state, as demonstrated by Long (1954) and Durran and Klemp (1986), \( F = 1 \) only at the mountain crest in both shallow-water and statically stratified systems. The stratified flow is subcritical upstream, where potential energy begins to be converted to kinetic energy, and the flow starts to accelerate as the \( \theta \) surface descends. Crossing the mountain crest, the flow has gained sufficient velocity and continues to accelerate becoming supercritical down the leeside slope. When the supercritical flow reaches its lowest level, it jumps up abruptly with a conversion of kinetic energy to potential energy, recovering to the ambient altitude. As a result, the Froude number becomes less than 1.0 as shown at the rightmost point above. Therefore, the strong downslope winds in the dry lake are developed in such an analog to the transition from subcritical to supercritical in conventional hydraulic framework. Even though Smith (1985), Durran and Klemp (1987), and Durran (1991) developed their nonlinear hydraulic and breaking wave theories in 2D spaces, their theories appear to be valid for this 3D model simulation of dynamical study, in which both wind shear and stratification vary continuously in the vertical and horizontal over complex topography.
8. LLJ forecasting discussion

In the above LLJ dynamic analyses, we demonstrate that the large-scale pressure gradient along the jet direction is one of the primary driving forces (mainly $PG_{\text{large}} + PG_{\text{slope}} + PG_{\text{depth}}$) in the valley, where it initiates and maintains the gap wind LLJ. It also produces the cross-mountain wind component at the mountain crest required for the downslope LLJ in the dry lake. The relationship of the large-scale pressure gradient to the downslope LLJ can be seen in the Froude number definition (5) involving the upstream wind $U_o$. Therefore, it might be a useful operational predictor for such LLJ events. Developing a predictive capability would be valuable to operational forecasting since an LLJ in the Lut Desert valley and the dry lake may cause hazardous operating conditions due to turbulence and dust lifting.

To explore this possibility, we could correlate the observed synoptic pressure gradient with the observed wind speeds in the valley, but there are insufficient wind observations in that area. Instead, we compare the observed synoptic pressure gradient with the modeled synoptic pressure gradient, and then, having found good agreement, check for a correlation between the modeled synoptic pressure gradient and the modeled winds. For the observed synoptic pressure gradient, the difference of sea level pressure is calculated between two surface stations: Tobat Heydarich (35.3°N, 59.2°E), on the eastern edge of the Kavir Desert basin and north of the Lut Desert valley, and Chah Bahar (25.4°N, 60.8°E), south of the dry lake and on the northern shore of the Gulf of Oman. The two locations are approximately 1000 km apart and marked as points E and F in Fig. 5. These stations are chosen because they contain a more complete time series of observations than others in the study area, the distance matches the along-jet spatial scale, and their orientation parallels the jet and the hypothesized relevant synoptic pressure gradient. The 12–21-h numerical forecasts of sea level pressure are bilinearly interpolated to these two points at 3-h intervals for the first step of this analysis, the time series of the observed and modeled synoptic pressure differences are shown in Fig. 18 and reveal excellent agreement in both magnitude and timing. For the second step, the 12–21-h modeled along-jet winds at point J in the dry lake (Fig. 5), at the average jet-core height (200 m above the ground), are also plotted in Fig. 18. Point J is closer to the Gulf of Oman, so the winds at the location are more important than in the Lut valley. In order to display better the correlation with pressure, the winds in the down-jet direction (northerly) are expressed with positive signs.

We see that the LLJ begins when the pressure difference exceeds 10 mb. The variation in pressure difference and its relationship to the jet speed is explained as follows. Recall that the high pressure in the Iran Plateau is established on 12 February after the passage of a deep 500-mb trough, while the mountain lee troughing occurs in the Gulf of Oman and the Persian Gulf. As a result, the north–south pressure gradient dramatically increases as shown in Fig. 18 on 10 and 11 February. During these two days, the winds are strong at night, weak during the day, indicating dominant diurnal forcing. Late on 12 February, the high pressure in the north and the lee trough in the gulf are both intensified (Figs. 3 and 4), so the pressure gradient further increases. The strongest gradient is achieved on 13 and 14 February. During this time, the winds are consistently high and the diurnal variation is reduced. On 15 February, a warm upper-level ridge moves over Iran (Fig. 2d) and the pressure gradient decreases. The winds weaken and a strong diurnal variation reappears.

We conclude that a pressure difference of 10–11 mb is conducive to the formation of the LLJ in the Lut valley and the dry lake. This pressure gradient may be determined easily from operational surface reports or global analyses and forecasts. However, this paper has shown that the synoptic pressure gradient is only one of the factors that determine the formation and intensity of the LLJ in this area. The dynamical analyses of gap and downslope winds in the previous sections demonstrate that there are some other important factors: a low-level cold air supply and the static stability of the cold air. These two together produce terrain-related pressure gradients ($PG_{\text{depth}}$ and $PG_{\text{slope}}$) to enhance the gap wind LLJ in the valley. This cold air supply and its strong static stability also create an upstream, elevated inversion over the mountain crest, which is one of the necessary conditions for the development of the downslope LLJ in the dry lake. Both the upstream depth and static stability of the inversion affect the intensity and duration of the leeside winds. The Froude number definition (5) reveals one of the connections between the upstream cold air characteristics and the downslope LLJ. On the other hand, this study is a single case simulation and is certainly not entirely sufficient to obtain all the factors.
that contribute to the configuration of such LLJ events. Many more cases and sensitivity studies are needed in the future.

9. Summary

A low-level northerly jet in the Lut Desert valley and the Jaz Murian dry lake of Iran has been identified using the mesoscale model COAMPS and through dynamical analyses. The jet carries cold, dry air from the upwind plateau Kavir Desert basin south to the Gulf of Oman and persists for three days. Our studies have shown that the jet is a result of several factors operating at different scales. It occurs after the synoptic high pressure builds in the Iran Plateau and the coastal mountain lee troughing takes place in the Gulf of Oman and the Persian Gulf. Such a synoptic circulation pattern supports the LLJ with a north–south pressure gradient as background forcing. Meanwhile the LLJ occurs when a low-level cold air layer intrudes southward into the Lut valley and the dry lake, which provides the LLJ with a stably stratified air mass. Under these favorable meteorological conditions, the elongated Lut valley produces convergent, channeling gap flow and the steep slope of the dry lake initiates strong cold pool downslope motion. Such a combination of meteorological conditions and complex topographic forcing makes this LLJ different from most other LLJs in other parts of the world.

For the gap wind LLJ in the Lut valley, it is found that in the along-jet direction, both the pressure gradient and momentum advection contribute to maintain the jet. As the jet intensifies, these forces are largely offset by friction. The balance (or imbalance) among the pressure gradient, advection, and frictional forces controls the intensity and persistence of the jet. In the cross-jet direction, the prevailing force balance is nearly geostrophic, that is, between the mesoscale pressure gradient and the Coriolis force associated with the northerly jet. The along-jet pressure gradient, as a driving force to the LLJ, consists of three components, among which the components resulting from the sloping terrain and cold air characteristics are also responsible for the gap wind evolution.

For the downslope LLJ in the dry lake, it is found that there is a dynamic transition in the airflow occurring at the mountain crest when the northerly winds traverse from the Lut valley to the dry lake. The transition is from subcritical on the windward side to supercritical on the lee side. The downslope jet is found to be a striking analog to a conventional hydraulic system with a subsequent hydraulic jump farther downstream. The results of theoretical and numerical studies by Smith (1985), Durran and Klemp (1986), and Durran (1991) are well applied to this real case in explaining the physical mechanisms of the downslope LLJ. Several major parameters, such as the depth, static stability, upstream cold air inflow, and the height and slope of the mountain, interact with one another to control the LLJ structure, strength, and duration in the dry lake.

COAMPS is shown to accurately predict the large-scale circulation and pressure gradient force that initiate and support the LLJ. This jet may mobilize and transport dust when it sweeps over the Lut Desert valley and the Jaz Murian dry lake, restricting visibility and degrading electromagnetic and optical propagation wherever it passes, especially as it arrives in the Gulf of Oman. A study related to these processes is under way. The correlation between the modeled pressure gradient and the wind speed suggests that a difference of 10 mb or more over a 1000-km horizontal distance may be one of the conditions conducive to the formation of the LLJ. This index implies some useful applications of COAMPS, other numerical models or surface observations to forecast or diagnose such an event in this remote area. Many more case studies are needed to explore the possibility of LLJ prediction by defining and verifying the robustness of predictive rules.

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