Validation of the Coupled NCEP Mesoscale Spectral Model and an Advanced Land Surface Model over the Hawaiian Islands. Part II: A High Wind Event*

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(Manuscript received 30 March 2004, in final form 22 March 2005)

ABSTRACT

A high wind event (14–15 February 2001) over the Hawaiian Islands associated with a cold front is simulated using the National Centers for Environmental Prediction (NCEP) Mesoscale Spectral Model (MSM) coupled with an advanced land surface model (LSM). During this period, a strong high pressure cell moved to the northeast of the Hawaiian Islands following the passage of the cold front. The cell then merged with the semipermanent subtropical high and resulted in windy conditions across the state of Hawaii. Analyses of soundings from Lihue on Kauai and Hilo on the Big Island reveal a mean-state critical level below 400 hPa, a strong cross-barrier flow (~13 m s⁻¹), and the presence of a trade wind inversion.

The MSM–LSM predicts downslope windstorms on the lee sides of mountains or ridges with tops beneath the trade wind inversion and within ocean channels between islands. In the case of high mountains with a peak height above the trade wind inversion, weak winds are simulated on the lee side. Around the corners of the islands and in gaps between mountains, gap winds and downslope windstorms are both important for the development of localized leeside windstorms.

The localized windstorms over the Hawaiian Islands develop as a result of interactions between large-scale airflow and the complex local topography. Since the terrain is not adequately resolved by the 10-km RSM–LSM, it is no surprise that these windstorms are better simulated by the high-resolution nonhydrostatic MSM–LSM than the 10-km RSM–LSM.

1. Introduction

Localized high winds frequently occur over the Hawaiian Islands during the cool season (November–April) due to combined effects from large-scale synoptic forcing and complex local terrain (Schroeder 1977). There are generally two types of synoptic disturbances that bring high winds to the Hawaiian Islands: cold fronts and kona storms (Blumenstock and Price 1967). Heavy rains and strong southerly winds frequently occur in the prefrontal zone; whereas strong, gusty north-easterly trade winds dominate after the frontal passage as the postfrontal high pressure system merges with the semipermanent subtropical high and centers just to the north of the Hawaiian Islands (Schroeder 1993). Moist southerly winds related to a kona storm (Simpson 1952; Ramage 1962) may persist for an extended period of time if such a system develops to the west of the Hawaiian Islands.

Local terrain plays a critical role in determining the locations of high winds (Ramage et al. 1977; Schroeder 1977). For the eight major islands in the Hawaiian chain (Hawaii, Maui, Kauai, Oahu, Molokai, Lanai, Kahoolawe, and Niihau), the terrain is mountainous and complex. Noteworthy is the fact that 50% of the land is above 600 m mean sea level (MSL). The summit elevations of Kauai and Oahu are from 1 to 1.5 km. The eastern islands of Maui and Hawaii consist of massive mountains with summit elevations of 3–4 km. Lee wind maxima are frequently recorded on Kauai and Oahu...
The northern and southern corners of the major Hawaiian Islands are places with frequent occurrences of localized high winds. Gaps between mountains and channels between islands are also prominent places of localized high winds (Patzert 1969; Smith and Grubisic 1993; Chen and Nash 1994; Lumpkin 1998).

Forecasting the location and magnitude of high winds is challenging in Hawaii due to sparse surface and upper-air observations in the mid-Pacific. With 10-km horizontal resolution (Fig. 1a), the operational Regional Spectral Model (RSM) (Juang and Kanamitsu 1994) is unable to resolve fine flow structures and pinpoint the locations where high winds are expected to occur. In early 2003, the nonhydrostatic version of the RSM (referred to as the Mesoscale Spectral Model, MSM) developed at the National Centers for Environmental Prediction (NCEP; Juang 2000) was coupled with the modified Oregon State University Land Surface Model (LSM; Zhang et al. 2005). The MSM–LSM was set up for three subregions of the state of Hawaii at high resolutions (≥3 km).
(Figs. 1b–d). In this study, the coupled MSM–LSM is used to simulate a high wind event over the Hawaiian Islands that occurred during 14–15 February 2001 in association with a cold front passage. The extent that the localized high winds over the island chain could be predicted using a high-resolution model is studied. The 10-km RSM–LSM simulations are presented alongside the MSM–LSM simulations for comparison. Furthermore, the possible mechanisms for the development of localized high winds at various locations over the Hawaiian Islands will also be examined. Section 2 contains a brief overview of gap winds, downslope windstorms, and airflow over an isolated island. Synoptic conditions in association with this high wind event are discussed in section 3. Section 4 validates the MSM–LSM and RSM–LSM predicted surface winds using observations. Model-simulated flow patterns over the Hawaiian Islands are examined in section 5. Conclusions and discussion are given in section 6.

2. An overview of gap winds and downslope windstorms

a. Gap winds

Gap winds (Reed 1931) are driven by the imposed gap-parallel pressure gradient. They are in approximate ageostrophic equilibrium between the inertia term and the pressure gradient force. Overland and Walter (1981) specify this balance of forces in the along-gap direction as

\[
\frac{d}{dx} \left( \frac{u^2}{2} \right) = -\frac{1}{\rho} \frac{\partial \rho}{\partial x},
\]

where \( u \) is the velocity, \( \rho \) is the density, and \( \frac{\partial \rho}{\partial x} \) is the gap-parallel pressure gradient. In integrated form, this gives the Bernoulli equation

\[
\frac{u^2}{2} = \frac{u_0^2}{2} - \frac{\Delta \rho}{\rho},
\]

where \( u_0 \) is the initial velocity and \( \Delta \rho \) is the pressure difference. Equation (2) can be used to estimate \( u \) at any point in the gap. Qualitatively, wind speeds increase along the gap and reach their maximum near the exit. However, the effects of surface drag and vertical entrainment can reduce the acceleration in the along-gap direction by almost 50% (Overland 1984; Lackmann and Overland 1989). Bell and Bosart (1988) show that friction can strongly offset sustained accelerations during Appalachian cold-air damming events. Mass et al. (1995) take into account a three-way balance between inertia, friction, and the imposed pressure gradient and arrive at a superior relationship to explain downgap acceleration:

\[
u^2(x) = \left[ \frac{u^2(0)}{k} - \frac{P_x}{k} e^{-2\Delta x} + \frac{P_x}{k} \right],
\]

where \( u(0) \) is the initial wind speed, \( u(x) \) is the wind speed some distance \( x \) from the starting point along the gap, \( P_x \) is the along-gap pressure gradient force, and \( k \) is a friction coefficient that is a function of surface roughness, stability, and boundary layer depth. Gap winds are recorded in many areas such as the Strait of Juan de Fuca in Washington State (Reed 1931; Overland and Walter 1981), the Columbia River Gorge bordering Washington and Oregon (Cameron and Carpenter 1936), and the Shelikof Straits in Alaska (Lackmann and Overland 1989).

b. Downslope windstorms

Observations of the strongest Colorado windstorms with peak gusts in the 50–60 m s\(^{-1}\) have been collected by aircraft (Lilly and Zipser 1972) and Doppler lidar (Neiman et al. 1988). Brinkmann (1974) examines 20 Boulder windstorms and notes that favorable large-scale conditions for the development of windstorms are a stable layer or inversion level above mountaintop and relatively strong winds at that level. Klemp and Lilly (1975) suggest that strong downslope winds occur when the atmosphere has a multilayer structure that produces a constructive superposition (reinforcement) of vertically propagating mountain waves reflected by variations in vertical static stability. Klemp and Lilly (1978) further note that reflections could occur due to the presence of critical levels (i.e., levels of zero wind or flow reversal). In a series of papers, Clark and Peltier (1977, 1984), Peltier and Clark (1979), and Clark and Farley (1984) show that even without a critical level in the ambient flow, a growing wave may break and generate a “self-induced” critical level.

Colman and Dierking (1992) identify three necessary criteria for strong downslope winds (Taku winds) to occur in southeast Alaska as a manifestation of amplified mountain waves. These include 1) an inversion at or just above ridgetop, somewhere between 1.5 and 2 km MSL; 2) strong cross-barrier flow near ridgetop, typically 15–20 m s\(^{-1}\) in geostrophic wind speed; and 3) cross-barrier flow decreasing with height to a critical level somewhere between 3 and 5.5 km MSL. Dierking (1998) documents a mountain wave windstorm that occurred in the Taku River valley near Juneau, Alaska. Mass and Albright (1985) note substantial wave activity above the lee slopes of the Washington Cascade Mountains during an intense windstorm event. Durran (1986)
shows that strong winds will occur along the lee slope when the fluid undergoes a transition from subcritical flow upstream to supercritical flow over the mountain, and a hydraulic jump in the lee. Smith (1985) and Durran (1986) show that the transition to supercritical flow is sensitive to the height of an elevated inversion. Removal of a crest-level inversion prevents the development of strong downslope winds. Smith and Sun (1987) develop generalized hydraulic solutions pertaining to downslope winds that show qualitative agreement with observations.

Most of the theoretical studies of the downslope windstorms have focused on the case of flow over infinitely long mountain ranges with no variations in the direction parallel to the ridge axis. Clark et al. (1994) perform three-dimensional simulations of a severe Boulder windstorm with realistic orography using time-dependent inflow boundary conditions and compare their results with Doppler lidar observations. Colle and Mass (1998a,b) emphasize the role of the three-dimensional topography of the Washington Cascades on the leeside windstorm and the sensitivity of windstorms to critical-level height, cross-barrier pressure gradient, crest-level stability, and the magnitude of the cross-barrier flow. The most destructive winds occurred primarily along the steepest foothills of Stampede Gap’s western terminus where gap winds and mountain wave dynamics both play a role.

c. Flow over a subtropical island

The island terrain for all major Hawaiian Islands is three-dimensional. In the case of circular 3D mountains, the airflow is qualitatively different from 2D flow for the same flow parameters because of horizontal streamline splitting (Miranda and James 1992; Li and Chen 1998). Other dynamic processes such as vortex shedding downstream of the island (Nickerson and Dias 1981; Schär and Smith 1993; Sun and Chern 1993) also occur in a 3D flow. For Froude number \( \text{Fr} \gg 1 \) (\( \text{Fr} = U/Nh \), where \( U \), \( N \), and \( h \) are upstream wind speed, Brunt–Väisälä frequency, and mountain height, respectively), the airflow approaches the three-dimensional potential flow solution (Smith 1980). Based on the linear theory of nonrotating fluid, for a bell-shaped mountain, stagnation occurs at the surface and aloft as the \( \text{Fr} \) passes below 0.77 (Smith 1989, 1990). As the \( \text{Fr} \) approaches 0, the airflow is determined by the horizontal potential flow solution (Drazin 1961).

Early simulations of island flow response using realistic orography were made by Lavoie (1974) for the island of Oahu and Nickerson (1979) for the island of Hawaii. Lavoie (1974) simulates a weak hydraulic jump above the leeside slopes. Flow splitting and stagnation on the windward side of island of Hawaii as a result of island blocking were simulated by Smolarkiewicz et al. (1988) and Smolarkiewicz and Rotunno (1989). For \( \text{Fr} \) below 0.5, the flow is essentially nonlinear. In addition, the island flow response is also sensitive to the diabatic heating associated with clouds and precipitation (Chen and Feng 2001) and the diurnal heating cycle at the surface (Feng and Chen 2001).

Using aircraft data gathered during the Hawaiian Rainband Project (HaRP; 11 July–24 August 1990), Smith and Grubišić (1993) identify hydraulic jumps within the regions of accelerated trades to the north and south of the island of Hawaii (the Big Island) where a supercritical flow is suddenly brought back to a subcritical state via an abrupt transition in which streamlines jump upward, and horizontal velocity decreases. The jump is also discernible as an abrupt decrease of potential temperature and increase of relative humidity. They also noticed jumps on the leeward slopes with downward air motion immediately downstream of the crest. These jumps were observed under normal trade wind weather. Mountain wave activities in the vicinity of the Hawaiian Islands have been documented (Burroughs and Larson 1979).

The Hawaiian Islands have various sizes (\(<140 \) km), shapes, and mountain heights (500–4000 m). Each island is under different \( \text{Fr} \) flow regimes under the same trade wind conditions. They have irregular shapes with gaps between mountains and variations in heights along the mountain ridges. The high winds in various parts of the state and the adjacent waters under the same strong trade wind conditions (\( >14–15 \) m s\(^{-1}\)) may be related to gap winds, downslope windstorms, or a combination of the two depending on the local terrain. In this study, the high-resolution MSM model that adequately depicts the island terrain with full model physics is used to identify the locations and mechanisms for the development of the localized high winds for the 14–15 February 2001 case.

3. Synoptic-scale conditions

During the period of 14–15 February 2001, a strong high pressure cell moved to the north of the Hawaiian Islands following the passage of a cold front (Fig. 2a). This high cell reached its maximum strength of 1038 hPa by 1400 HST (Hawaiian standard time; UTC = 10 h + HST) 14 February 2001 after merging with the semipermanent subtropical high (Fig. 2b). At 500 hPa (Fig. 3), a NE–SW-oriented trough was located northeast of the Hawaiian Islands. The strong surface high produced windy conditions (trade winds in excess of 15 m s\(^{-1}\) with gusts in excess of 22 m s\(^{-1}\)) across the state
of Hawaii, with trees knocked down by high winds and power disrupted in many island communities. Most of the power failures and much of the damage occurred on the lee sides of mountain ranges on Kauai and Oahu as well as on the lee side of gaps between mountain ranges. Gale warnings were issued by the National Weather Service (NWS) at 0400 HST 14 February for all channels in Hawaiian coastal waters. In addition to localized high winds, rain showers were reported throughout the state of Hawaii except on the lee sides of high mountains of Maui and the Big Island. The Big Island received the heaviest rainfall, with more than

Fig. 2. Surface analyses for 1400 HST on (a) 13 and (b) 14 Feb 2001 (adapted from the subjective analyses by forecasters at the NWS Forecast Office in Honolulu). Isobars are every 4 hPa.
121 mm of rainfall recorded on the windward lower slopes in the 24-h period ending at 0000 HST 15 February.

Atmospheric soundings are taken routinely at Lihue on Kauai and at Hilo on the Big Island. Figure 4a shows the sounding from 1400 HST 14 February at Lihue. A weak trade wind inversion is identifiable between 800 and 700 hPa (2–3 km), which is just above the mountain ridges on Kauai and Oahu. A critical layer is located between 500 and 400 hPa (5.5–7.5 km). The cross-barrier wind speed near the ridgetop (1–1.5 km) is about 13 m s⁻¹. The northeasterly trades are the strongest between the surface and 2 km MSL. At Hilo (Fig. 4b), a weak trade wind inversion is identifiable between 700 and 650 hPa (3–3.5 km) with a critical level located between 600 and 500 hPa (4–6 km). The trade wind inversion at Hilo is much weaker than that at Lihue, likely due to the midlevel trough that lies just ahead of the Big Island (see Fig. 3b). Notice that the low-level winds at Hilo (Fig. 4b) are slightly weaker than at Lihue (Fig. 4a). This is likely because the low-level winds at Hilo are strongly affected by island blocking and the diurnal heating cycle due to the massive and high mountains on the Big Island (Yang and Chen 2003).

The Lihue and Hilo soundings at 1400 HST indicate that the three necessary criteria for strong leeside downslope winds to occur as a manifestation of amplified mountain waves (Colman and Dierking 1992) are satisfied for Kauai and Oahu and for small mountains over the island of Maui and the Big Island. Specifically, 1) the trade wind inversion is situated above the ridgetops, 2) the cross-barrier flow near ridgetop is strong (—10–15 m s⁻¹), and 3) a mean-state critical level is located below 400 hPa. Colle and Mass (1998b) perform three-dimensional idealized simulations of windstorms along the western side of the Washington Cascade Mountains using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (Penn State–NCAR) Mesoscale Model (MM5) at 3-km horizontal resolution. Their results show that damaging winds are favored along the Cascade foothills for strong cross-barrier flow cases (>13 m s⁻¹) when an environmental critical level exists below 400 hPa. From the Lihue and Hilo soundings (Fig. 4), the Froude number for Kauai and Oahu, and for low mountains and ridges of the island of Maui and the Big Island, is about 1 (U —10–15 m s⁻¹, N ~0.01 s⁻¹, and h ~1–1.5 km). Wave breaking aloft is likely to occur with amplifying mountain waves above the leeside slopes (Smith 1989; Smolarkiewicz et al. 1988).

4. Verifications of model-forecasted surface winds

In this section, the forecasted surface winds from the RSM–LSM and MSM–LSM are validated at 10 surface sites over the Hawaiian Islands where wind observations are available. The 10 surface sites are Kahuku, Bellows, Wheeler Air Force Station (AFS), and Waialua on Oahu (Fig. 1c); Hana, Haleakala, and Maalaea Bay on Maui (Fig. 1b); and Honokaa, Kawaihae, and the Upolu airport (AP) on the Big Island (Fig. 1b). The validation period runs from 0000 HST 14 February through 0000 HST 15 February 2001.

a. Oahu

Both models produce a gradual increase in the surface wind speeds during the day, which is consistent
with observations (Fig. 5). The observed strong winds at Kahuku (11–12 m s\(^{-1}\)) (Fig. 5a) and Bellows (11–14 m s\(^{-1}\)) (Fig. 5c), and the relatively weaker winds (2–6 m s\(^{-1}\)) at Wheeler AFS and Waialua within the valley between the Koolau and Waianae mountain ranges (Figs. 5e and 5g), are reasonably well predicted by the MSM–LSM. In contrast, the RSM–LSM underestimates the strong winds at Bellows by about 2 m s\(^{-1}\) and overestimates the weak winds at Wheeler AFS and Waialua by about 2 and 6 m s\(^{-1}\), respectively.

### b. Maui

On Maui (Fig. 6), both models reproduce the observed wind speeds at Hana (Fig. 6a). At Haleakala...
Fig. 5. Observed, and RSM–LSM and MSM–LSM simulated 10-m (left) wind speeds (m s\(^{-1}\)) and (right) wind directions (°) at (a), (b) Kahuku, (c), (d) Bellows, (e), (f) Wheeler AFS, and (g), (h) Waialua on Oahu, during the period of 0000 HST 14 Feb–0000 HST 15 Feb 2001. Solid, dotted, and dashed lines refer to the observations, RSM–LSM, and MSM–LSM simulations, respectively. Stars indicate the magnitudes of reported wind gusts (m s\(^{-1}\)). Refer to Figs. 1a and 1c for the locations of these surface sites. The 10-km RSM–LSM and 1.5-km MSM–LSM runs were initialized at 1400 HST 13 Feb 2001.

(Fig. 6c), the MSM–LSM forecasted wind speed is rather consistent with the observations whereas the RSM–LSM forecasts are persistently 3–4 m s\(^{-1}\) higher than the observed. The terrain height of this site in the RSM–LSM domain is 1 km, ~2 km lower than the actual height. Atmospheric soundings at Lihue and Hilo (Figs. 4a and 4b) have indicated that the easterly trade winds are strongest between the surface and 2 km but decrease upward. Related to orographic amplification of trade wind flow through the Central Valley (Daniels and Schroeder 1978), the gusty wind speeds at Maalaea Bay are in excess of 15 m s\(^{-1}\) throughout the period. This is better resolved by the MSM–LSM than the RSM–LSM. The MSM–LSM simulated wind directions are consistent with observations at Maalaea Bay (Fig. 6f), but appreciable discrepancies are noted at Hana and Haleakala (Figs. 6b and 6d), likely due to the transient nature of the winds there that the model fails to capture.

c. The Big Island

On the Big Island (Fig. 7), both models capture the strongest wind speed in the afternoon hours of 14 February at Honokaa and Upolu (Figs. 7a and 7e). The wind pattern reported at Kawaihae (Fig. 7c) is peculiar in that the gusty wind speeds are much higher than the 2-min-average wind speeds. This is likely related to the fact that this surface site is located in a bay area on the
lee side of the Kohala Mountains, which experiences significant fluctuations in winds (Tyson 1968; Schroeder 1981). The RSM–LSM and MSM–LSM forecasted wind speeds at this site are much higher than the observed 2-min-average wind speeds; however, they are rather consistent with the observed high gusty wind speeds there. The predicted surface wind directions from both models are generally consistent with observations at Honokaa and Upolu (Figs. 7b and 7f). However, large discrepancies occur at Kawaihae (Fig. 7d) due to fluctuations in winds.

d. Forecasting errors

Based on observations and the RSM–LSM and MSM–LSM simulations, forecasting errors of 10-m wind speed (m s$^{-1}$) and 10-m wind direction (°) for the period of 0000 HST 14 February–0000 HST 15 February 2001 are computed at the 10 surface sites. The forecasting error is defined as

$$fe = \frac{1}{N} \sum_{i=1}^{N} |A_o^i - A_m^i|,$$

where $A$ refers to either 10-m wind speed or 10-m wind direction; $N$ is the sample size ($N = 25$); subscripts $o$ and $m$, respectively, denote observations and model simulations; and superscript $i$ ranges from 1 to 25.

The results (Table 1) indicate lower forecasting errors for the MSM–LSM than the RSM–LSM in terms of 10-m wind speeds at most of the surface sites. At Wheeler AFS, Waialua, and Haleakala, large forecasting errors are noted for the RSM–LSM simulated 10-m wind speeds, related to the fact that the local terrain is not well represented by the RSM–LSM as mentioned before. The forecasting errors in the 10-m wind direction in the MSM–LSM are not much different from those of the RSM–LSM since the dominant wind direction for most times of the day is from the east-northeast both in the MSM–LSM and RSM–LSM simulations (Figs. 5–7).

5. Horizontal and vertical structure of the flow field

In this section, the horizontal and vertical distributions of the MSM–LSM forecasted flow field within
each nested domain are examined for downslope windstorms and gap winds through channels and mountains and compared with the RSM–LSM forecasts. The temporal evolution of the forecasted upstream surface winds along 154°W by the 10-km RSM–LSM (Fig. 8) shows strong winds in the late afternoon and early evening hours of 14 February 2001, with a maximum around 1800 HST near 20.8°N. Therefore, we will focus our discussion on the simulated wind fields at this time.

Figure 8 also reveals that the zonal wind component is dominant (~95%) in the horizontal wind field.

The horizontal map of 10-m winds forecasted by the

<p>| Table 1. Forecasting errors of 10-m wind speed (m s⁻¹) and 10-m wind direction (°) for the period of 0000 HST 14 Feb–0000 HST 15 Feb 2001, computed from observations and the RSM–LSM and MSM–LSM simulations at the 10 surface sites. Computation of the forecasting error is given by Eq. (4) in the text. |
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10-km RSM–LSM valid at 1800 HST 14 February is shown in Fig. 9. General features include 1) weaker winds on the windward sides of Maui and the Big Island due to island blocking, 2) wakes on the lee sides of Maui and the Big Island, 3) two vortices with weak (2–4 m s⁻¹) return flow in between on the lee side of the Big Island, and 4) strong winds (≥16 m s⁻¹) at the southern and northern corners of major islands as well as within the Alenuihaha Channel. Wind patterns on Kauai and Oahu are relatively uniform due to the coarse representation of the islands’ terrain by the 10-km model grid.

a. Kauai domain

For the nested MSM–LSM Kauai domain with 1.5-km horizontal resolution, relatively weak winds of 6 m s⁻¹ are evident on the windward side with strong winds (≥16 m s⁻¹) along the lee slopes (Fig. 10a). The wake of the island is located between its western coast and the island of Niihau with the wind speed as low as 2 m s⁻¹. Strong winds (≥16 m s⁻¹) also prevail on the northwestern and southern corners of Kauai, consistent with damage reports of tiles blown off and tree limbs downed in those areas.

We constructed the longitude–height cross section along 22.13°N (the solid line in Fig. 10a) showing the MSM–LSM forecasted zonal wind and equivalent potential temperature (Fig. 10b), and pressure vertical velocity (Fig. 10c). This cross section features a mountain with a peak height of 1.2 km, with a gentle eastern slope and a sharp western slope. This peak height is about 0.1–0.2 km lower than the actual height. Strong downslope winds (≥24 m s⁻¹) are evident along the western slopes with downward motion reaching 25 Pa s⁻¹ (Fig. 10c). These downslope flows extend past the coastline and, when combined with the strong winds deflected along the northwestern corner of the island (see Fig. 10a), form a low-level easterly jet with the strongest winds of 24 m s⁻¹ located about 1 km above the surface. This low-level jet extends farther downstream. The eastern slope of the mountain is characterized by decelerating, rising flows with surface wind speeds below 6 m s⁻¹. The critical level is located around 6 km.

Orographic lifting of the easterly trade wind flow is evident along the windward (eastern) slopes. Upstream of the island, tight vertical gradients of equivalent potential temperature representing the trade wind inversion layer are present just below the 3000-m level. In the midtroposphere above the 3000-m level, the vertical gradients of equivalent potential temperature are rather small (Fig. 10b). The 324-K equivalent potential temperature surface at about the 3000-m level marks the top of the trade wind inversion layer. It tilts upward above the windward slope and reaches its highest level before the crest. There is a slight increase in wind speed before the air reaches the ridgetop as in subcritical shallow-water flow. Above the leeside slope, a hydraulic jump is evident. After reaching the highest level before the crest, the constant equivalent potential temperature surfaces within the trade wind layer (the layer between the surface and the 324-K surface) tilt significantly.
downward with high wind speed (>24 m s⁻¹) within the layer. The thickness of the trade wind layer reduces significantly above the leeside slope. The air in this layer accelerates down the slope as in supercritical shallow-water flow. The 324-K surface drops below the 1600-m level at the foothills on the lee side and recovers to above the 2000-m level about 15–20 km offshore.

Along the same latitude (22.13°N) in the 10-km RSM–LSM domain (Figs. 10d and 10e), amplification of the trade winds is evident toward the lee side of the mountain ridge. In this cross section, the peak height of the mountain is 0.6 km, ~0.7 km lower than the actual height. The downward tilt of the constant equivalent potential surfaces is less significant as compared with the MSM–LSM. With a lower mountain height in the RSM–LSM as compared with the MSM–LSM, the gravity wave–induced acceleration above the leeside slope is less. Thus, orographic enhancement of the trade

**Fig. 10.** The MSM–LSM forecasted (a) 10-m winds (m s⁻¹) over the Kauai domain, (b) longitude–height cross section along MKA–MKB showing zonal winds (solid lines) and equivalent potential temperature (dashed lines), and (c) the pressure vertical velocity constructed from the MSM–LSM 28-h forecasts valid at 1800 HST 14 Feb. (d), (e) As in (b), (c), respectively, except that they are along RKA–RKB and constructed from the RSM–LSM 28-h forecasts. Isotachs are every 2 m s⁻¹ in (a) and 4 m s⁻¹ in (b) and (d). Isentropes are every 2 K. Pressure vertical velocity is every 5 Pa s⁻¹ in (c) and 1 Pa s⁻¹ in (e). The wind barbs in (a) are plotted every three grid points in both directions. The 1.5-km MSM–LSM and 10-km RSM–LSM simulations were initialized at 1400 HST 13 Feb 2001. The locations of the cross sections, MKA–MKB and RKA–RKB, are shown in Figs. 10a and 9, respectively.
winds through mountain wave amplification is only partially resolved by the RSM–LSM.

This high wind case was also run using the original version of the MSM for the Kauai domain (as well as for the Oahu and Molokai–Maui–Hawaii domains, not shown). Relatively weak winds of 12 m s\(^{-1}\) are evident on the windward side with strong winds (\(\geq 28\) m s\(^{-1}\)) along the lee slopes (Fig. 11a). The MSM–LSM also simulates relatively weak winds on the windward side with strong winds along the lee slopes (Fig. 10a). However, the MSM simulated surface wind speeds over land are twice as strong as the MSM–LSM simulated surface wind speeds, related to small surface roughness lengths in the MSM (Zhang et al. 2005). The MSM simulated surface winds over land are also smoother than the MSM–LSM simulations, due to the specification of homogeneous surface properties in the MSM. In the longitude–height cross section along 22.13\(^\circ\)N, the MSM forecasted flow pattern (Figs. 11b and 11c) is rather similar to that in the MSM–LSM (Figs. 10b and 10c) except that the surface wind speeds over land in the MSM are larger than in the MSM–LSM.

b. Oahu domain

For the nested MSM–LSM Oahu domain with 1.5-km horizontal resolution, relatively weaker winds (<10 m s\(^{-1}\)) are found on the windward slopes of the Koolau and Waianae mountain ranges on Oahu (Fig. 12a). Except along the leeside coasts of the Waianae and Koolau mountain ranges, strong winds (>14 m s\(^{-1}\)) are found in the coastal areas with the strongest (>16 m s\(^{-1}\)) immediately off the northern and southern coasts of the island. Surface winds are relatively weaker in the island interior due to surface friction.

The longitude–height cross section along MOA–MOB (21.45\(^\circ\)N) (Figs. 12b and 12c) is characterized by two mountain ranges: the Koolau Range in the east and the Waianae Range in the west, with a similar peak height of 0.6 km in the model. This height is about 0.2–0.3 km lower than the actual mountain height. Strong downslope winds (>24 m s\(^{-1}\)) are evident above the lee slopes of both mountain ranges (Figs. 12b and 12c). This is consistent with reports that recorded damages mainly on the lee sides of both mountain ranges. The strongest winds with speeds in excess of 24 m s\(^{-1}\) are located above the surface. Upstream of the island, there are two weak inversion layers (~2000 and ~3500 m). The amplification of winds on both lee slopes occurs mainly below 2000 m. Weak hydraulic jumps are evident on the lee sides of both mountain ranges. Figure 12c also shows that the upward motion along the windward slope of the Koolau mountain range is larger than along the windward slope of the Waianae moun-

tain range (10 versus 5 Pa s\(^{-1}\)) likely due to the contribution from abundant moisture along the windward slope of the Koolau mountain range. At the time when strong trades reach the Waianae mountain range, the moisture content is due less to the removal of water vapor on the windward slope of the Koolau mountain range by precipitation (Schroeder et al. 1977). Our sen-
sitivity test with the Koolau mountain range removed (not shown) reveals significant moisture content along the windward slope of the Waianae Range with vertical motion comparable in magnitude and vertical extent to that along the windward slope of the Koolau mountain range in the control run.

In contrast, in the RSM–LSM domain (Fig. 1a), the island of Oahu is represented as one mountain with a peak height of 0.3 km. This height is about 0.6 km lower than the actual height. Thus, the orographic effect is not properly resolved by the RSM–LSM. Along the same latitude (21.45°N) in the 10-km RSM–LSM domain (Figs. 12d and 12e), strong winds (~20 m s⁻¹) are predicted toward the lee side of the mountain ridge with relatively uniform flow immediately above the mountain surface. The isentropic surfaces are also nearly horizontal. The vertical motion (1–2 Pa s⁻¹) is more than one order of magnitude smaller in the RSM–LSM

Fig. 12. Same as in Fig. 10 except for the Oahu domain and along MOA-MOB and ROA-ROB. The locations of MOA-MOB and ROA-ROB are shown in Figs. 12a and 9, respectively.
than in the MSM–LSM. It is apparent that amplification of trade wind flow on the lee slopes is not resolved by the RSM–LSM model.

c. Molokai–Maui–Hawaii domain

This model domain (Fig. 1b) contains several significant geographic features. There are two massive volcanoes, Mauna Kea and Mauna Loa, on the Big Island with peak heights greater than 4.0 km. Another small mountain range, the Kohala Mountains, is located over the northern part of the island with a peak height of 1.5 km. The Waimea Saddle is situated between Mauna Kea and the Kohala Mountains. In the MSM–LSM model with 3-km horizontal resolution, these mountains are \(~0.2–0.3\) km lower than the actual heights. Haleakala Mountain on Maui has a peak height of 2.3 km in the model, which is \(~0.7\) km lower than the actual height. The western Maui mountains have a peak height of 0.8 km in the model, \(~0.7\) km lower than the actual height. Between Maui and the Big Island lies the Alenuihaha Channel.

Figure 13a shows the 10-m winds predicted by the 3-km MSM–LSM for the Molokai–Maui–Hawaii domain valid at 1800 HST 14 February. Salient features associated with the Big Island include (a) decelerating and splitting flow along the windward side, (b) accelerating winds to the north and south of the island, (c) strong winds downstream of the Waimea Saddle (Waimea jet), and (d) a wake on the lee side of the island consisting of two elongated counterrotating eddies that give rise to a wide region (\(~60\) km) of reverse flow along the wake axis. These flow features are generally consistent with the schematic diagram by Smith and Grubišić (1993) for the Big Island with one exception: a small Kohala wake observed by Smith and Grubišić (1993) is not identified in the MSM–LSM simulations. This is because Smith and Grubišić (1993) examine the flow features under normal trade wind conditions when weak winds are frequently recorded on the lee side of the Kohala Mountains (Schroeder 1981). For our high wind case, favorable large-scale conditions result in strong downslope winds through amplification of mountain waves on the lee side of the Kohala Mountains.

Along the Central Valley between Haleakala Mountain and the western mountains on Maui, relatively weaker winds (\(~8\) m s\(^{-1}\)) are predicted on the windward side with strong winds (\(~20\) m s\(^{-1}\)) on the leeward side (Fig. 13a). This flow pattern agrees qualitatively with Daniels and Schroeder’s (1978) findings based on low-level wind observations. The weaker winds on the windward side of the Central Valley are due to the island blocking, whereas the strong winds along the leeward side of the Central Valley result primarily from the combined channeling effect and the impact of subsidence from the surrounding mountains. A sensitivity test involving doubling the mountain height produces relatively weak winds (\(<10\) m s\(^{-1}\)) along the lee side of the Central Valley (not shown).

Strong winds are also predicted within the oceanic
channels between the islands. A narrow wake on the lee side of Haleakala Mountain is evident (Fig. 13a). This narrow wake delineates strong winds to the south of the island and strong winds along the leeward side of the Central Valley to the north. In contrast, the 10-km RSM–LSM resolves a broad wake on the lee side of Maui without predicting strong winds along the leeward side of the Central Valley (Fig. 9). This is likely due to the fact that the combined channeling effect and the impact of the surrounding mountains along the Central Valley are not properly resolved by the RSM–LSM due to the lower (≈0.2 km) mountain height on west Maui in the 10-km grid (see Fig. 1a).

Longitude–height cross sections are constructed from the MSM–LSM simulations to investigate the interaction of airflow with the complex terrain. The positions of these cross sections are shown in Fig. 13b. The cross section along the Alenuihaha Channel is represented by MCA–MCB. MWA–MWB crosses the Waimea Saddle, MLA–MLB crosses a high mountain (Mauna Loa). The cross section, MSA–MSB, examines the airflow over the southern corner of the Big Island.

Across the Alenuihaha Channel (Figs. 14a and 14b), the airflow below 3 km starts to accelerate and descend in the entrance region and continues the acceleration and descent for nearly 100 km downstream. The strongest wind of 26 m s\(^{-1}\) is located at 0.8 km MSL. This flow pattern is consistent with Patzert’s (1969) results based on cruise observations and Smith and Grubišić’s (1993) findings based on aircraft observations. Upstream of this channel, tight vertical gradients of equivalent potential temperature are discernible below the 4000-m level. The equivalent potential temperature surfaces below the 4000-m level tilt considerably downward when they reach the channel exit with high wind speeds within the layer. A moderate hydraulic jump is evident at the exit region through an abrupt transition in which the equivalent potential temperature surfaces jump up-
ward, and horizontal velocity decreases. The tightening of the equivalent potential temperature surfaces notably decreases at the channel exit region (Fig. 14a). In the 10-km RSM–LSM simulations (Figs. 14c and 14d), strong winds of 24 m s\(^{-1}\) are evident in the exit region with a weak hydraulic jump. However, the downward tilt of the equivalent potential temperature surfaces is less pronounced when compared with the MSM–LSM.

Like gap winds, channel winds are driven by the imposed channel-parallel pressure gradient (Overland and Walter 1981). Figure 15 shows a close correspondence between the temporal evolution of pressure difference and zonal wind difference across the channel. Since friction is rather small over the ocean, we can apply Eq. (2) to estimate the channel exit wind speeds. For the period of maximum winds (∼1800 HST 14 February), with the entrance wind speeds of 16 m s\(^{-1}\) (Fig. 14a) and the pressure difference across the channel at 2.75 hPa (Fig. 15), using a density of 1.225 kg m\(^{-3}\), the channel exit wind speeds are approximately 26.6 m s\(^{-1}\). This translates to a cross-channel increase in wind speeds of 10.6 m s\(^{-1}\), which is about 4 m s\(^{-1}\) higher than the model-simulated 6 m s\(^{-1}\) (Fig. 15).

The existence of large mountains on either side of the channel adds complexity to the mechanisms of these strong channel winds. As noted earlier, a hydraulic jump is predicted by the MSM–LSM at the exit region of the channel. A sensitivity run that doubles the height of the terrain was conducted to investigate the role of the surrounding terrain. The results (Figs. 16a and 16b) show a slightly stronger gap flow (~26 m s\(^{-1}\)) when compared to the control run (Figs. 14a and 14b), due to the enhanced channel-parallel pressure gradient because of increased mountain height (not shown). The relatively stronger gap flow also expands in both the horizontal and vertical directions when compared to the control run. Compared to the control run (Figs. 14a and 14b), a weaker hydraulic jump is discernible at the exit region of the channel (Figs. 16a and 16b). This change in the channel exit flow response after the terrain is doubled appears to point to the impact of the surrounding mountains on the circulations within the channel. It may be inferred that the strong channel exit hydraulic response in the control run (Figs. 14a and 14b) is due to the terrain flow perturbations in the lee that were carried over to the nearby channel.

Ramage (1979) shows that the Waimea Saddle is one of the windiest sites on the Big Island. Across the Waimea Saddle (Fig. 17a), relatively weaker winds

![Fig. 15. Hourly surface pressure differences (hPa, solid lines) and zonal wind speed differences (m s\(^{-1}\), dashed lines) between the entrance region (C2) and the exit region (C1) of the Alenuihaha Channel during 0000 HST 14 Feb–0000 HST 15 Feb predicted by the 3-km MSM–LSM. The locations, C1 and C2, are shown in Fig. 13b. The MSM–LSM was initialized at 1400 HST 13 Feb 2001.](image-url)

![Fig. 16. Same as in Figs. 14a and 14b except that the terrain height is doubled.](image-url)
(<8 m s\(^{-1}\)) are predicted along the windward side with strong winds (>24 m s\(^{-1}\)) on the lee side. The strongest winds of 30 m s\(^{-1}\) are located above the lee slopes. The downward motion in association with these strongest winds exceeds 25 Pa s\(^{-1}\) (Fig. 17b). A low-level easterly jet of 22 m s\(^{-1}\) exists farther downstream. A moderate hydraulic jump is evident above the lee slopes of the Waimea Saddle. The 10-km RSM–LSM predicts strongest winds (~24 m s\(^{-1}\)) along the leeside upper slopes of the Waimea Saddle (Figs. 17c and 17d). A weak hydraulic jump is also identified in the RSM–LSM simulations.

Gap winds and downslope winds are also known to occur simultaneously (Colman and Dierking 1992). This appears to be the case for the strong winds downstream of the Waimea Saddle (Waimea jet). Surface pressure differences across the saddle are on the order of 3–6 Pa (100 km)\(^{-1}\) in the MSM–LSM simulations (Fig. 18), with the zonal wind differences closely following the pressure pattern. Equations (2) and (3) can be applied to estimate the gap exit velocity. For a period of near maximum winds (~1800 HST 14 February), with the entrance wind speeds at 12 m s\(^{-1}\) (Fig. 17a) and the pressure difference across the gap (~80 km) at 6 hPa, using a density of 1.225 kg m\(^{-3}\) and k value of 1.7 × 10\(^{-5}\) m\(^{-1}\) (Mass et al. 1995), we would get an exit velocity of 34 m s\(^{-1}\) and 20 m s\(^{-1}\) from

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**Fig. 17.** Same as in Fig. 14 except for the cross sections along the Waimea Saddle: MWA–MWB and RWA–RWB. Pressure vertical velocity is (b) every 5 Pa s\(^{-1}\) and (d) 1 Pa s\(^{-1}\).

**Fig. 18.** Same as in Fig. 15 except for the Waimea Saddle between W1 and W2. The positions W1 and W2 are shown in Fig. 13b.
Eqs. (2) and (3), respectively. The model simulations (Fig. 18) show along-gap differences of 10 m s\(^{-1}\). Thus, an overestimation by Eq. (2) and a slight underestimation by Eq. (3) are indicated. It is clear that surface drag is important.

Across the Mauna Loa volcano (Figs. 19a and 19b), weak westerly (upslope) winds (1–2 m s\(^{-1}\)) are evident along the lee slopes of the mountain. Since strong downslope winds above the lee slopes are trapped below the trade wind inversion (see Fig. 10b), for mountains with heights above the trade wind inversion (about 3 km), strong downslope winds are absent in the lee. Furthermore, the winds aloft at 4 km are relatively weak (<5 m s\(^{-1}\)) (Fig. 4b). Thus, neither downslope windstorms nor hydraulic jumps are present in the model simulations. Farther downstream of the mountain, a return flow of 4–6 m s\(^{-1}\) is found between the surface and 1 km MSL. This return flow originates near 157.2°W, about 100 km west of the leeside coast, and extends eastward toward the mountain with an approximate north–south extent of 60 km at low levels (see Fig. 13a). Lumpkin (1998) and Xie et al. (2001) also suggest an origination point of the return flow at around 157°W. Smith and Grubišić (1993) observe a broad region of reverse flow (5–10 m s\(^{-1}\)) in the lee of the island, spanning more than 80 km in the crosswind direction under normal trade wind conditions. Nickerson and Dias (1981) estimate the north–south extent of the return flow around 60 km based on aircraft observations in June 1980. The MSM–LSM captures the return flow pattern reasonably well. In the 10-km RSM–LSM simulations, the horizontal extent of the return flow is less (Figs. 19c and 19d) when compared to the MSM–LSM simulations (Figs. 19a and 19b).

Strong winds (24 m s\(^{-1}\)) are predicted above the lee slope along the southern corner of the Big Island by the MSM–LSM (Figs. 20a and 20b). The airflow descends along the lee slope and then abruptly ascends in a jump-like feature. The critical level is identified at 8 km MSL with less wave activities above it. In contrast, the strong winds (~20 m s\(^{-1}\)) forecasted by the RSM–LSM are located above the mountain ridge, with an associated weak hydraulic jump (Figs. 20c and 20d). The magnitude of the vertical motion in the RSM–LSM simulations is about half of that predicted by the MSM–LSM. The terrain height in this cross section is 0.3 km in the

**Fig. 19.** Same as in Fig. 14 except for the cross sections along Mauna Loa: MLA–MLB and RLA–RLB. Pressure vertical velocity is (b) every 2 Pa s\(^{-1}\) and (d) 1 Pa s\(^{-1}\).
RSM–LSM, about 0.2 km lower than the actual height. Thus, the orographic effect is only partially resolved by the RSM–LSM.

Gap winds may also contribute to the strong winds associated with the southern corner of the Big Island (Fig. 20a). Figure 21 shows a close correspondence between the simulated temporal evolution of the pressure difference and the zonal wind difference across the southern corner, indicating characteristics of gap winds (Overland and Walter 1981; Mass et al. 1995).

In summary, downslope windstorms are predicted by the MSM–LSM along the lee slopes of Kauai and Oahu. These windstorms also display features of hydraulic jumps. At the exit region of the Alenuihaha Channel, strong winds are forecasted with characteristics of gap winds (Reed 1931) and a hydraulic jump. The model also resolves the Waimea jet (Ramage 1979) with the strongest wind speed of 30 m s⁻¹ above the lee slopes of the Waimea Saddle. A hydraulic jump is forecasted in association with the Waimea jet and the strong winds over the southern corner of the Big Island.

In contrast, the 10-km RSM–LSM predicts uniform trade wind flow for Kauai and Oahu, without large spatial variability in wind distributions related to complex local terrain.

Fig. 20. Same as in Fig. 14 except for the cross sections along the South Corner of the Big Island: MSA–MSB and RSA–RSB. Pressure vertical velocity is (b) every 5 Pa s⁻¹ and (d) 1 Pa s⁻¹.

Fig. 21. Same as in Fig. 15 except for the South Corner between S1 and S2. The positions S1 and S2 are shown in Fig. 13b.
6. Conclusions and discussion

A localized high wind event (14–15 February 2001) over the Hawaiian Islands associated with a cold front is simulated using the coupled MSM–LSM. It occurred when a strong high pressure cell moved to the northeast of the Hawaiian Islands following the passage of a cold front and merged with the semipermanent subtropical high, which resulted in windy conditions across the state of Hawaii. Analyses of the routine atmospheric soundings at Lihue on Kauai and Hilo on the Big Island reveal a mean-state critical level below 400 hPa and strong cross-barrier flows (~13 m s\(^{-1}\)), and a trade wind inversion. These large-scale conditions are favorable for the development of wave-induced downslope windstorms in the lee.

Comparisons between observations and model simulations over 10 surface sites show that the MSM–LSM reproduces the spatial distribution and magnitude of the surface wind better than the RSM–LSM. This is largely due to the much improved representation of the local complex terrain by the high-resolution MSM–LSM. Since the windstorms over the Hawaiian Islands develop as a result of interactions between large-scale airflow and the complex local topography, and the terrain is not adequately resolved by the 10-km RSM–LSM, it is not surprising that these windstorms are better simulated by the MSM–LSM than the 10-km RSM–LSM.

The MSM–LSM forecasts reproduce the general flow features in association with the local complex terrain. These include 1) decelerating and deflecting flow in the area immediately upstream of the islands, 2) wakes on the lee sides of major islands, 3) strong winds around the northern and southern corners of the islands, 4) strong winds within the oceanic channels between islands and within the gaps between mountains, and 5) strong winds on the lee sides of mountains or ridges with tops below the inversion.

Downslope windstorms are forecasted by the MSM–LSM on the lee sides of mountains, and these windstorms display features of a hydraulic jump and are trapped below the trade wind inversion. The top of the trade wind inversion layer is lifted on the windward side and reaches the highest level before reaching the crest. It tilts significantly downward above the leeside slope with a rapid decrease in the depth of the trade wind layer (the layer between the surface and the top of the trade wind inversion). Since downslope windstorms are trapped below the trade wind inversion, they are not simulated in the lee of high mountains that extend above the inversion.

Localized strong winds are also forecasted in gaps between mountains and the northern and southern corners of the islands with ridge tops below the trade wind inversion. In these regions, gap winds and downslope wind storms are both important for the development of localized windstorms in the lee. These localized windstorms share many similarities with the windstorms that frequently occur on the western lee side of the Washington Cascades (Colle and Mass 1998a,b) and the Taku windstorms in Alaska (Colman and Dierking 1992). Strong winds are also forecasted at the exit region of the Alenuihaha Channel with characteristics of gap winds and a hydraulic jump.

Without the LSM, the main features of the localized windstorms are also simulated by the MSM. However, the wind speed at the surface is overestimated by the MSM because the surface roughness is interpolated from the gridpoint values of the Global Forecast System (GFS) model. Most of the GFS grids over the Hawaiian Islands are ocean points with small roughness length.

For this case, a mean-state critical level below 400 hPa, strong cross-barrier flows, and a trade wind inversion are likely the main large-scale factors for the development of windstorms. The presence of a critical level as well as a trade wind inversion are common features in the postfrontal region during the cool season (November–April) season. The key element appears to be strong cross-barrier flows that depend on the strength of the migratory high that merges with the semipermanent subtropical high. Since amplifications of winds above the lee slopes are trapped below the inversion, these windstorms mainly occur within ocean channels between islands and on the lee sides of mountains or ridges with tops beneath the inversion. Due to the lack of high-resolution observational data over the Hawaiian Islands, especially on the lee sides of mountains, the transient evolution of the storm development is still yet to be examined. It is apparent that the deployment of low-level wind profilers over the Hawaiian Islands is highly desirable for the study and monitoring of the frequently observed localized high wind events.

Acknowledgments. The authors wish to thank Diane Henderson for editing the text. Three anonymous reviewers are thanked for their very constructive comments, which have helped to improve the manuscript substantially. The model simulations were carried out on a DEC AlphaServer ES40, which was funded by the National Weather Service Pacific Region (NWSPR). This work has been supported by NOAA through the
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