

Numerical Simulation of the Maui Vortex in the Trade Winds

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Abstract

On the island of Maui, the dispersion of air pollutants associated with the field burning of biomass is complicated because the persistent Maui Vortex in the lee of Haleakala tends to trap smoke making pollution concentration worse in the central valley. In the present study we describe the spatially as well as temporally continuous short-term climatology of the airflow in the central valley of Maui during a summer month under trade wind conditions, simulated by a high-resolution mesoscale model initialized and forced by large-scale objective analyses. These simulations are compared with station observations. Under persistent trade wind conditions, an eddy would form in the central valley, and would remain steady and stationary regardless of the time of the day with little indications of vortex shedding even if the Froude number $Fr = \frac{U}{NH} < 0.4$, where U is the uniform flow speed, H the height of the obstacle, and N the Brunt-Väisälä frequency. However, vortex shedding could occur if the West Maui mountains were removed. Two major factors contributing to formation, maintenance, and steadiness of the Maui Vortex, solar heating and the accelerated northerly flow over the valley resulting from deflection of the trades by West Maui, are discussed. Suggestions are made for preferred synoptic conditions for field burning.

1. Introduction

The island of Maui, Hawaii, located approximately at 21°N and 156°W within the tropical belt, consists of two shield volcanos, the West Maui Mountains (elevation 1764 m) on the west and Haleakala (3055 m) on the east with a connecting isthmus (Fig. 1). Maui, the second largest of the Hawaiian Islands, spans approximately 76 km east-west and 50 km north-south with an area of 1887 km^2 . The broad, flat isthmus (central valley) between the West Maui Mountains and Haleakala is cultivated for sugar at the north end, and is a growing tourist destination at the south end (Armstrong,

1983).

It is common practice in Hawaiian sugar plantations to burn the crop prior to harvesting, primarily to separate the leaves from the sugar-laden stalks. The fires turn the leafy material to ash and kick up silicates, the primary constituent of surface soil, which are transported downwind in large smoke plumes. The smoke reduces visibility and poses a potential respiratory health hazard (Daniels, 1980; Schroeder, 1993). Sugarcane burning is therefore regulated by a state health agency. Fires are also set maliciously in the uncultivated vegetation. When wildfires occur, they react primarily to weather effects, thriving in high winds and low humidity. Wildfire has been a recurrent problem on the east

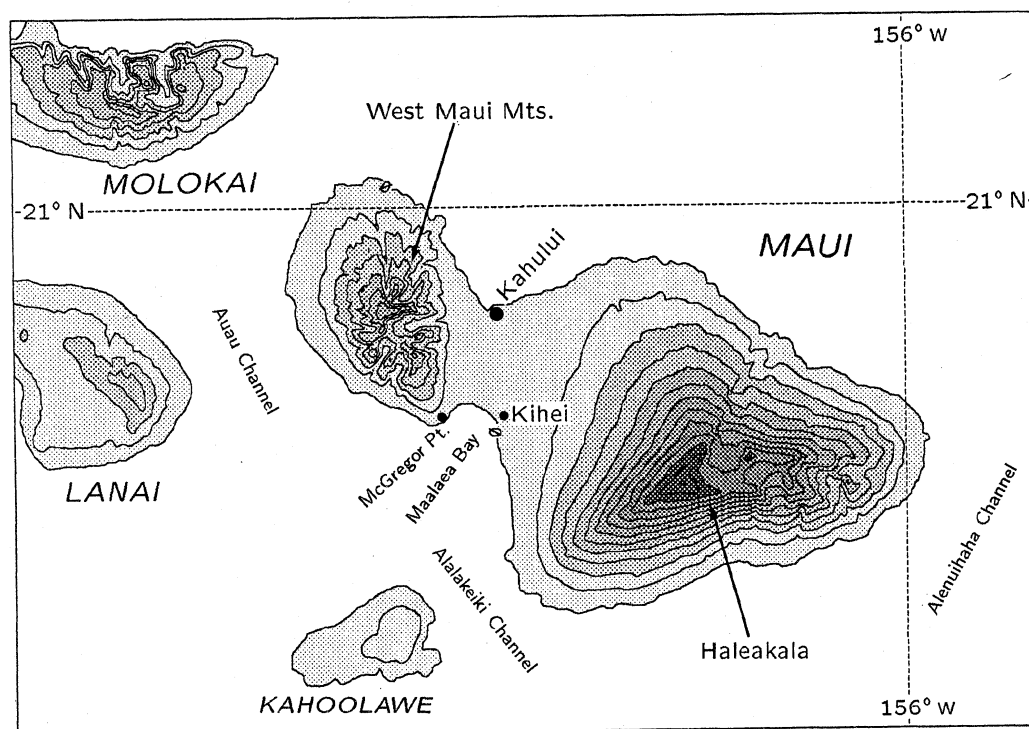


Fig. 1. Topography and geographical locations for Maui and the surrounding area. The contour interval of topography is 250 m.

and south side of the isthmus, its magnitude varying with the weather.

During the summer months the trade wind situation dominates the synoptic-scale flow patterns over the Hawaiian Islands in which a high pressure cell is situated either to the north of the Islands to bring northeast winds (Type 1a) or to the northeast of the Islands to bring southeast winds (Type 1b), as classified by Yeh *et al.* (1951). The combined percentage frequencies of these two “anticyclonic” types of flow patterns are approximately 86 per cent for June and 97 per cent for both July and August. An example of this trade wind situation is illustrated in Fig. 2 using the National Meteorological Center’s (NMC) large-scale objective analysis for 1400 LST 04 July 1988.

In the presence of persistent trade winds, diurnal variations due to mesoscale processes generally dominate Hawaii’s local weather (Daniels and Schroeder, 1978). In the case of Maui, the subsidence inversion at about 2 km above mean sea level (MSL) (Rogers, *et al.*, 1993) provides an effective upper lid locally and the central valley becomes a venturi due to the mechanical effects of Haleakala and West Maui deflecting northeasterly trades and channeling the surface airflow through the isthmus. This simple flow pattern in the central valley is complicated by the interaction of deflected trade winds with the southwesterly sea breeze/anabatic winds and the downslope mountain-valley winds (Leopold, 1949; Lyons,

1979). In fact, observations indicate the existence of a cyclonic eddy motion, known locally as the “Maui Vortex,” over the central valley at ~ 1000 m MSL, as shown in Fig. 3 (Leopold, 1949). The dominant northeasterly trade wind situation during the summer months seems to be the most favorable synoptic condition for generation of the Maui Vortex. As many works based on observations have contributed to our understanding of the modulation of the surface airflow by the diurnal heating cycle for the island of Hawaii (Garrett, 1980; Chen and Nash, 1994; Chen and Wang, 1995; Carbone *et al.*, 1995; Wang and Chen, 1995), similar studies for Maui should shed light on generation mechanisms for the Maui Vortex which appears to form when the deflected trades interact with sea breezes and mountain-valley winds.

The Maui Vortex is a cause for concern during the sugar harvest season (March to December), when it can entrain the ash and smoke from burning cane fields (Schroeder, 1993). While the trade winds generally act as natural ventilation that maintain better air quality at most locations in Hawaii, the Maui Vortex traps smoke from agricultural burns, tending to make pollution concentration worse in the central valley. To make the situation more serious, the Maui Vortex appears to be stationary and is not shed downstream periodically, unlike the observed cases of mesoscale vortices in the wake of large islands described by Etling (1990). The vortex may

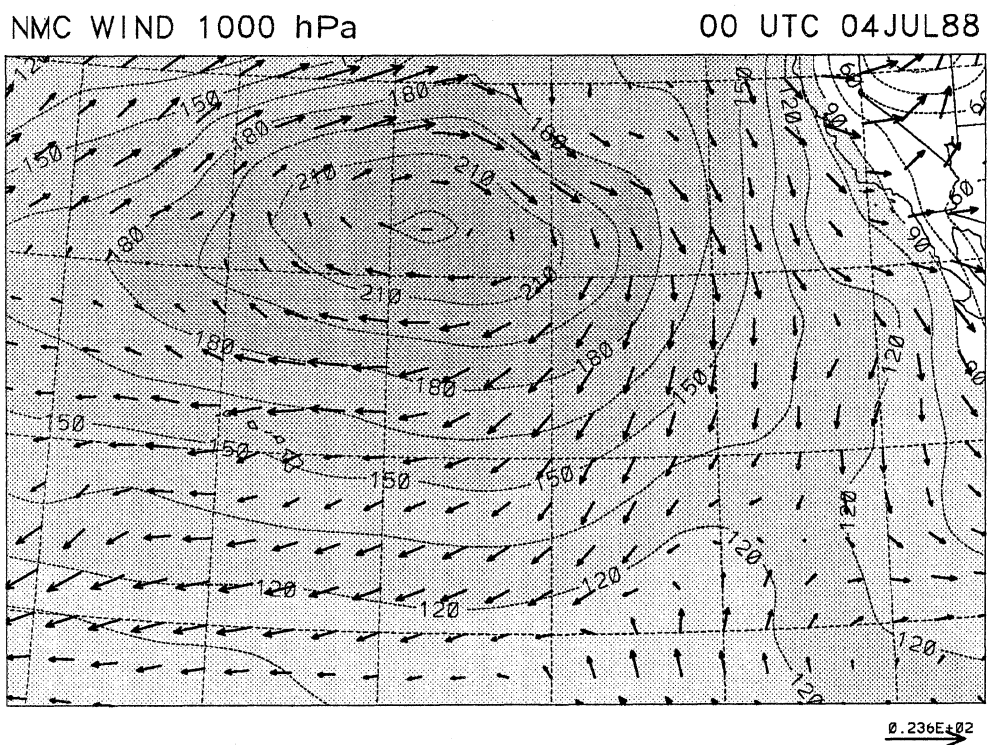


Fig. 2. An example of NMC’s objective analysis over the central Pacific Ocean on a $2.5^{\circ} \times 2.5^{\circ}$ grid in which an “anticyclonic” cell is situated to the north of Hawaii to bring northeast trade winds. The 1000-hPa fields of wind vectors and geopotential heights drawn with a contour interval of 15 m are shown for 1400 LST (0000 UTC) 04 July 1988.

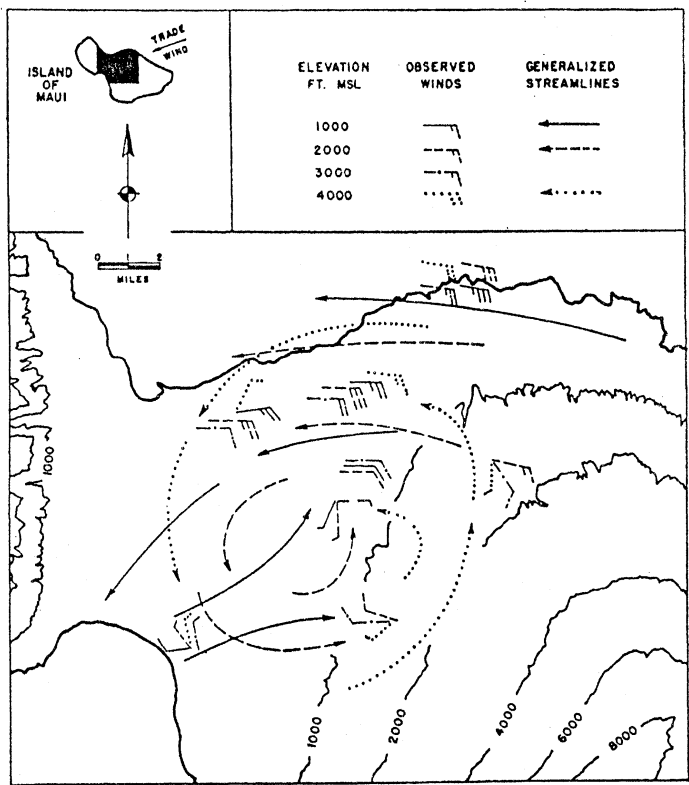


Fig. 3. Composite streamlines over the central valley of Maui at 1000-, 2000-, and 4000-ft levels based on pilot-balloon observations at 1200 LST estimated from the field experiment in July 1948 (reproduced from Leopold, 1949).

also influence the behavior and control of wildfires which occasionally occur in the isthmus and on the western flanks of Haleakala.

Although a considerable amount of ground-truth observations has been accumulated, including the data from the observation network maintained by the Hawaiian sugar industry and those from cooperative stations and the first-order station in Kahului, it is still inadequate for developing long-term statistics at a sufficient number of points that would describe climatology in the central valley of Maui. For example, in the field program conducted by Daniels and Schroeder (1978), the presence of the Maui Vortex could only be inferred in the analysis because the Vortex lay just south of the analyzed field. As described later, the Hawaiian sugar industry observation network mostly covers the northern and western portions of the perimeter of the region where the eddy is located but leaves the eastern and southern parts of the region without adequate measurements.

On the other hand, a variety of attempts has been made to model air flow over the islands of Hawaii and Oahu (Lavoie, 1974; Nickerson, 1979; Smolarkiewicz *et al.*, 1988; Rasmussen *et al.*, 1989; Ueyoshi and Han, 1991; Rasmussen and Smolarkiewicz, 1993; Reisnerx and Smolarkiewicz, 1994). These works have contributed significantly to our understanding of the effects of topographic barriers on air flow structure and precipitation over and around these islands by filling the spatial as well as temporal gaps in observational data analyses. We can expect similar contributions from model study of local air flows over Maui, which so far has not been seriously attempted. Particularly, in the light of inadequate available data, model study should be the most feasible way of gaining some insight into generation mechanisms for the Maui Vortex and understanding the spatial and temporal behavior of the smoke from the field burning and occasional wild fires.

In this study we attempt to fill the spatial as well as temporal gaps in observations in Maui by relating large-scale objective analysis to local meso- α scale circulations with the use of a numerical model. In particular, we describe a short-term climatology of the surface wind flow patterns in the central valley for the month of July 1988 developed from a high-resolution mesoscale model with the use of large-scale objective analyses as external forcing.

The numerical model and the experimental design adopted are described in the next section. In Section 3 we attempt to gain some insight into the characteristics of the Maui Vortex by analyzing the simulation results and comparing them with available observations. We investigate in Section 4 whether shedding of the Maui Vortex is possible. Pollution concentration in the central valley would be less serious if shedding of the Maui Vortex occurs intermittently.

Further discussions on various aspects of the model results, including better choices of the field-burning days, are presented in Section 5. This model study is summarized in Section 6.

2. Model description and design of the simulations

The simulations described in this article were performed with the Regional Atmospheric Modeling System (RAMS) developed at Colorado State University. The basic model structure is described in Tripoli and Cotton (1982). More recent model developments are described in Tremback *et al.* (1986), Cotton *et al.* (1988), and Pielke *et al.* (1992). In the vertical, a sigma- z terrain-following coordinate is used in the model version utilized here. In the horizontal, grids are mapped on the earth's surface using a polar stereographic projection. Two-way interactive grid nesting in RAMS allows local fine-mesh grids to resolve micro-scale atmospheric systems while simultaneously modeling the large-scale environment of the systems on a coarser grid. RAMS also allows us to use a non-hydrostatic mode for model integrations with a grid size on the order of a few kilometers. Temperature and moisture fluxes at the surface are determined from the surface energy balance, which includes both short- and long-wave fluxes (Chen and Cotton, 1983), latent and sensible fluxes, and sub-surface heat conduction from a soil temperature model (Tremback and Kessler, 1985). A radiation scheme described by Mahrer and Pielke (1977) is used which considers the influence of water vapor, ozone, and carbon dioxide on short- and long-wave radiative transfer. A modified Kuo-type cumulus parameterization is incorporated in the model, along with a bulk microphysical parameterization of precipitation processes, outlined in detail in Flatau *et al.* (1989). Smagorinsky's deformation eddy viscosity is used as the turbulence scheme (Tripoli and Cotton, 1982).

In this study, doubly-nested model domains are used. The outer model domain consists of a rectangular 440×400 km² area encompassing the entire islands of Hawaii with a 45×41 10-km grid mesh (Fig. 4). The inner model domain, shown in Fig. 5a, consists of a 112×92 km² area covering the island of Maui with a 57×47 2-km grid mesh. In the vertical, the model atmosphere is divided into 23 layers between the ground and 18.5 km with grid spacing of about 80 m near the surface stretching to approximately 2 km near the model top. There are 11 soil levels with the deepest level at 50 cm. Vegetation differences were not considered; the soil type was assumed to be clay loam everywhere. The "wall on top" boundary condition is used at the model top by setting the vertical velocity to zero. Model topography was derived from a 3-s (~ 80 m) data set acquired from the National Cartographical Center

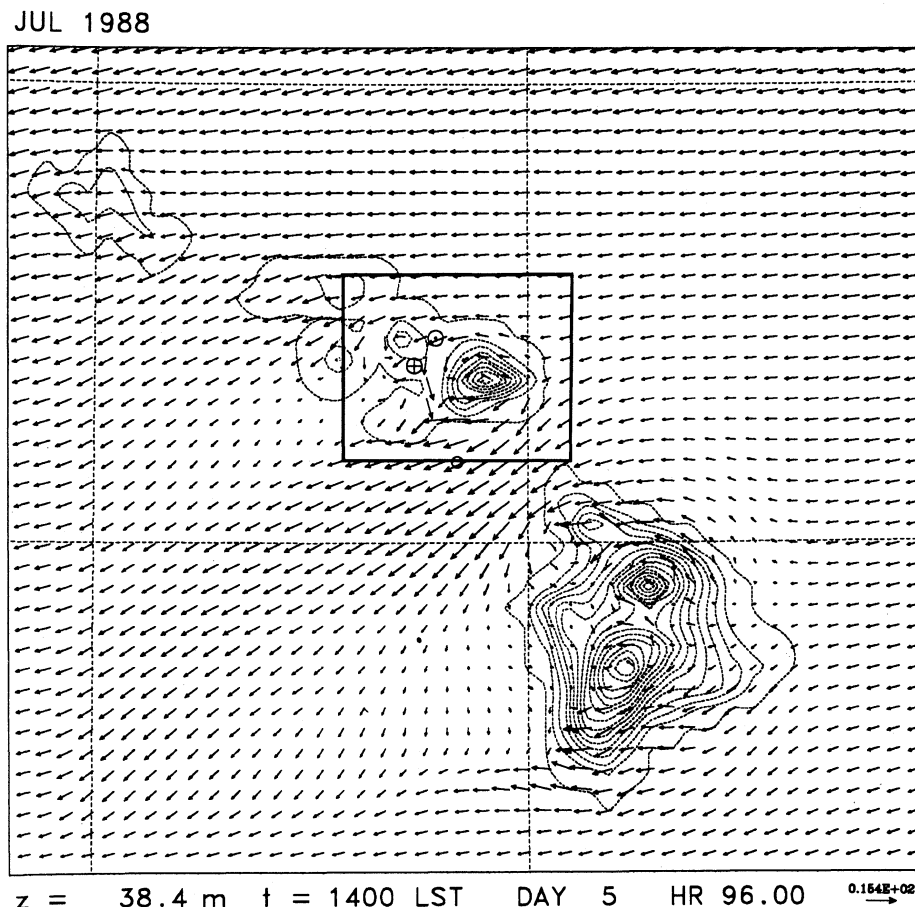


Fig. 4. Outer model domain and grid points utilized. The coarse grid points with a grid length of 10 km are located at the origins of individual wind vectors showing the model flow nearest to the surface (~ 38 m above the surface) for 1400 LST 04 July 1988. The rectangular area indicates the inner model domain (see Fig. 5). The contour interval of topography is 250 m.

with a silhouette averaging scheme that preserves realistic topography heights.

NMC's large-scale objective analyses were used to initialize and externally force the model. NMC's analyses are generated twice daily for 0000 UTC and 1200 UTC on a global $2.5^\circ \times 2.5^\circ$ grid mesh network. The data archived are virtual temperature, wind velocity components, and geopotential height at 11 mandatory pressure levels with the highest level at 70 hPa. Relative humidity is only archived for 6 levels at 300 hPa and below (see Roads *et al.*, 1995). The procedures of initialization and specification of the lateral boundary conditions used in this study are similar to that described in Ueyoshi and Roads (1993). Initially all gridpoints were assigned NMC's objective analysis for 0000 UTC 01 July 1988, interpolated onto both the inner and outer model grids.

The simulation was then integrated continuously for one model-month in a non-hydrostatic mode. During the integrations, two successive analyses were spatially and temporally interpolated and assigned to the lateral boundaries of the outer model grid as external forcing.

3. Results

Hourly instantaneous fields of various meteorological parameters at all model levels were archived for the entire 31 model-day period. Time series of wind vectors, speeds, and streamlines at selected hours were visually examined for the presence of the model Maui Vortex. Shown in Fig. 5 are typical near-surface wind vector fields developed over the island of Maui and its vicinity after midnight and in the afternoon during the month-long simulation. The so-called "Maui Vortex" detected in these instantaneous fields favorably compares with the observed eddy illustrated earlier in Fig. 3. Mean fields of model winds are described and compared with observations in Sections 3.1 and 3.2, respectively, while we examine in Sections 3.3 and 3.4, respectively, the effects of two major factors on generation and maintenance of the vortex, the southward deflection of the trades by West Maui and the surface heating. We found only a couple of model days which had failed to generate the Maui Vortex. The model results and the synoptic situations for these

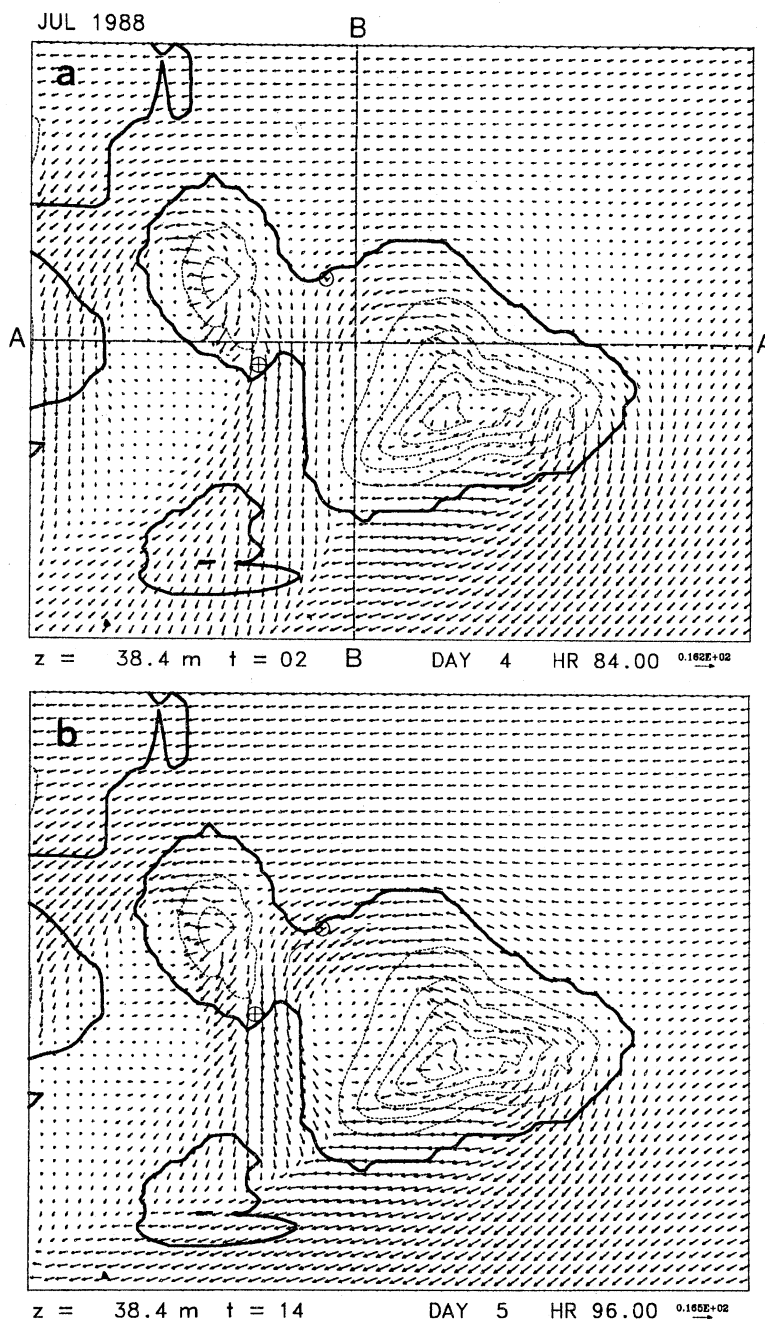


Fig. 5. (a) Wind-vector field depicting the near-surface flow for 0200 LST 04 July 1988 in the inner model domain (see Fig. 4). The inner grid points with a grid length of 2 km are located at the origins of individual wind vectors. Cross sections referred to in the text were taken along the lines A—A' and B—B'. (b) Wind-vector field depicting the near-surface flow for 1400 LST 04 July 1988. Mountain height contour intervals are 500 m.

“no-vortex” days are examined in Section 3.5.

3.1 30-day mean fields

Mean wind speeds and streamlines at the near-surface level (~ 38 m above the surface) in the central valley are shown in Fig. 6 at 6-h intervals. The averages were taken for the 30-day hourly model results excluding the initial 24-h period. The model streamlines compare favorably with the average fields esti-

mated from the 7–26 August 1976 observations by Daniels and Schroeder (1978).

At night, a cyclonic eddy is found in the eastern region of the valley at the foot of Haleakala, while the daytime center of the vortex is located on the mid-slope. At night, downslope flow from Haleakala is evident at higher elevations in the eastern portion of the valley and over the northern slope of Haleakala. The wind speeds remain less than 2 ms^{-1}

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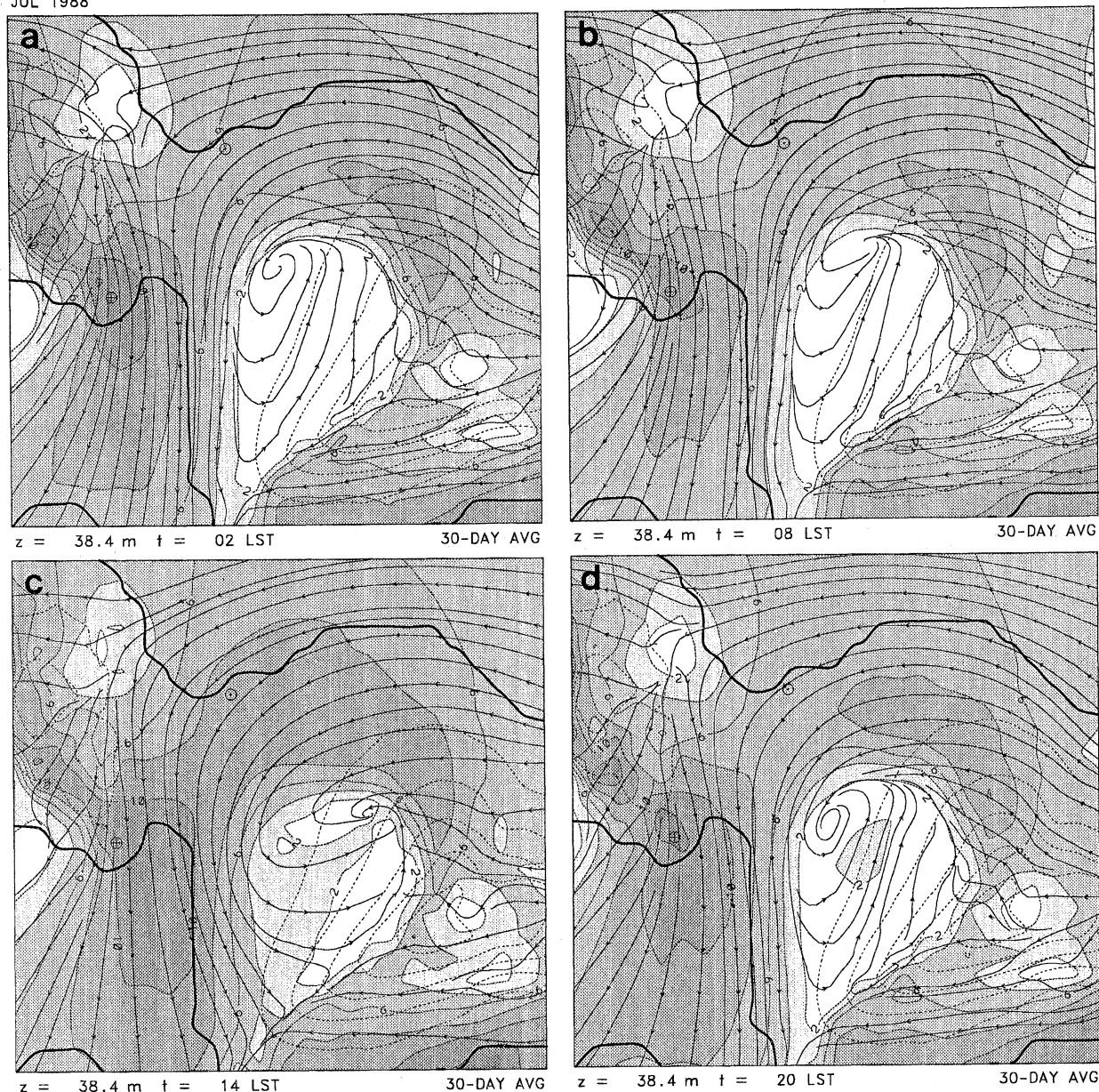


Fig. 6. 30-day mean wind speeds and streamlines at the near-surface level ($\sim 38 \text{ m}$ above the surface) in the central valley at 6-h intervals: (a) 0200 LST (nighttime), (b) 0800 LST (early morning), (c) 1400 LST (daytime), and (d) 2000 LST (early evening). The contours are drawn at intervals of 2 ms^{-1} for wind speeds (solid) and 500 m for mountain heights (dotted).

in the eddy during the most of the night. During the day, upslope flow dominates over the lee side of Haleakala, with its speeds increasing up to $\sim 4 \text{ ms}^{-1}$. Diurnal variations in surface wind are also apparent in the northern and southern flanks of the eddy. Wind speeds increase in the afternoon over wide areas within the path of the deflected trades and accelerate up to 12 ms^{-1} over Maalaea Bay, south of the isthmus.

Relative vorticity patterns corresponding to the flow fields in Fig. 6 are shown in Fig. 7. In both

daytime and nighttime cases, the mean relative vorticity is positive along the eddy flow path on all sides although it is much weaker along the southern side of the vortex where the southern branch of the split trades generates a shear-line with the eastward flow in the lee slope. Relative vorticity is generally stronger during the daytime than at night. Especially along the northern flank of the vortex, the daytime relative vorticity is twice as strong as that at night ($16 \times 10^{-4} \text{ s}^{-1}$ vs. $8 \times 10^{-4} \text{ s}^{-1}$). The cyclonic motion in the daytime flow can be clearly

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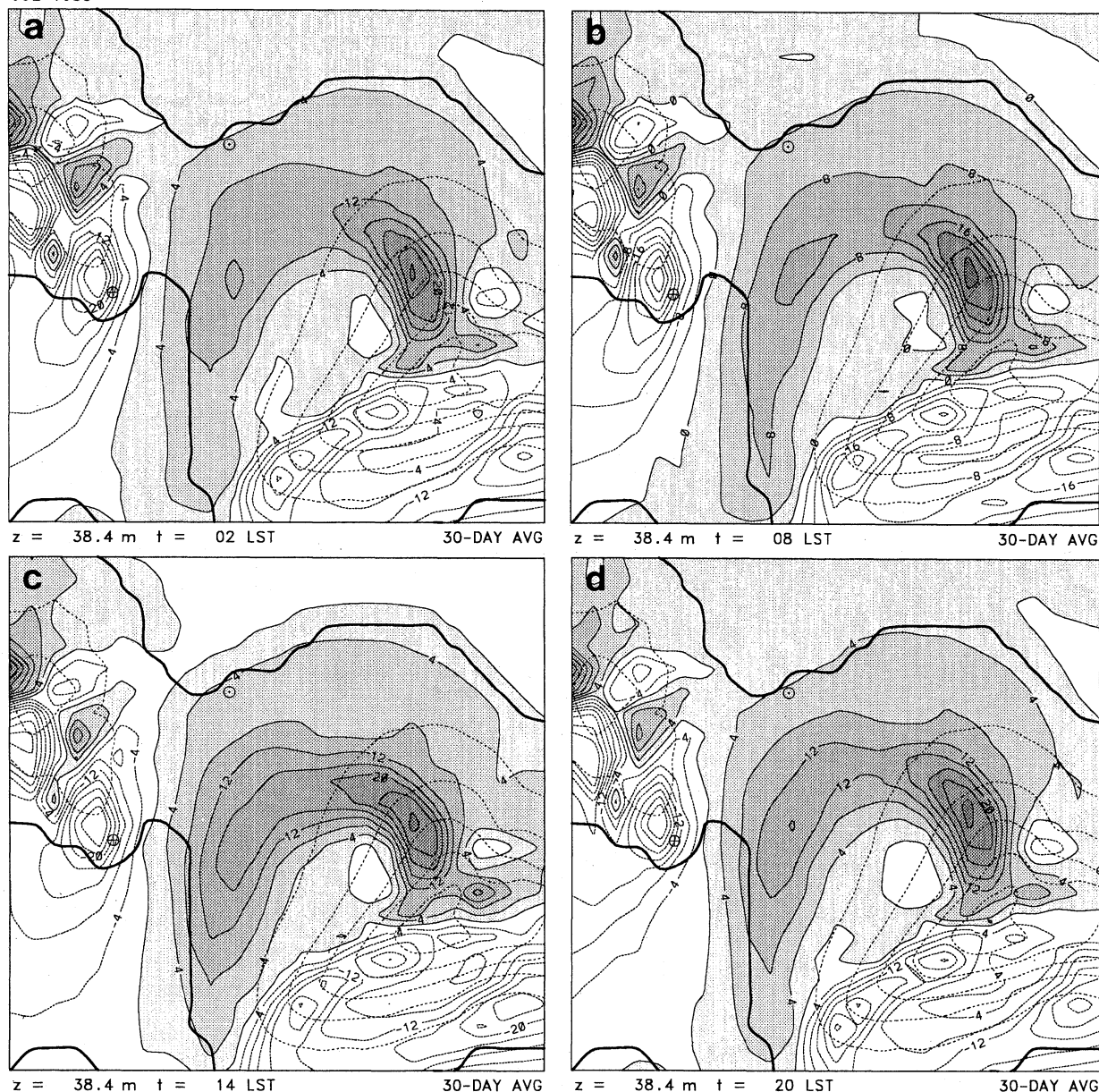


Fig. 7. As in Fig. 6, except for the corresponding relative vorticity. Positive relative vorticity is shaded. The contours are drawn at an interval of $4 \times 10^{-4} \text{ s}^{-1}$.

recognized in Fig. 6c and also in Fig. 10c shown later. This flow behavior is contrary to the nocturnal Melbourne eddy described by McGregor and Kimura (1989), which vanishes as the daytime heating increases.

The east-west and north-south cross sections of relative vorticity are shown in Fig. 8 for 0200 and 1400 LST along the lines approximately passing through the center of the vortex as depicted in Fig. 5a. Relative vorticity near the ground is generally stronger during the daytime than at the night. In particular, we note weakening of the nighttime relative vorticity near the ground at middle altitudes.

The vertical extent of positive relative vorticity both in day and night cases is up to about 2 km MSL, where the magnitudes of the relative vorticity are on the order of $4 \times 10^{-4} \text{ s}^{-1}$. It should be noted here that the maximum elevation of the model Haleakala is about 2850 m. However, mean streamline patterns on the model coordinate levels (not shown) reveal that a discernible vortex exists up to an altitude of about 1100 m MSL. In Leopold's (1949) daytime field observations, the vortex was also estimated to exist even at 1200 m MSL (see Fig. 3). For comparison, Smith and Grubišić (1993) noted that the depth of the main part of the wake in the lee of

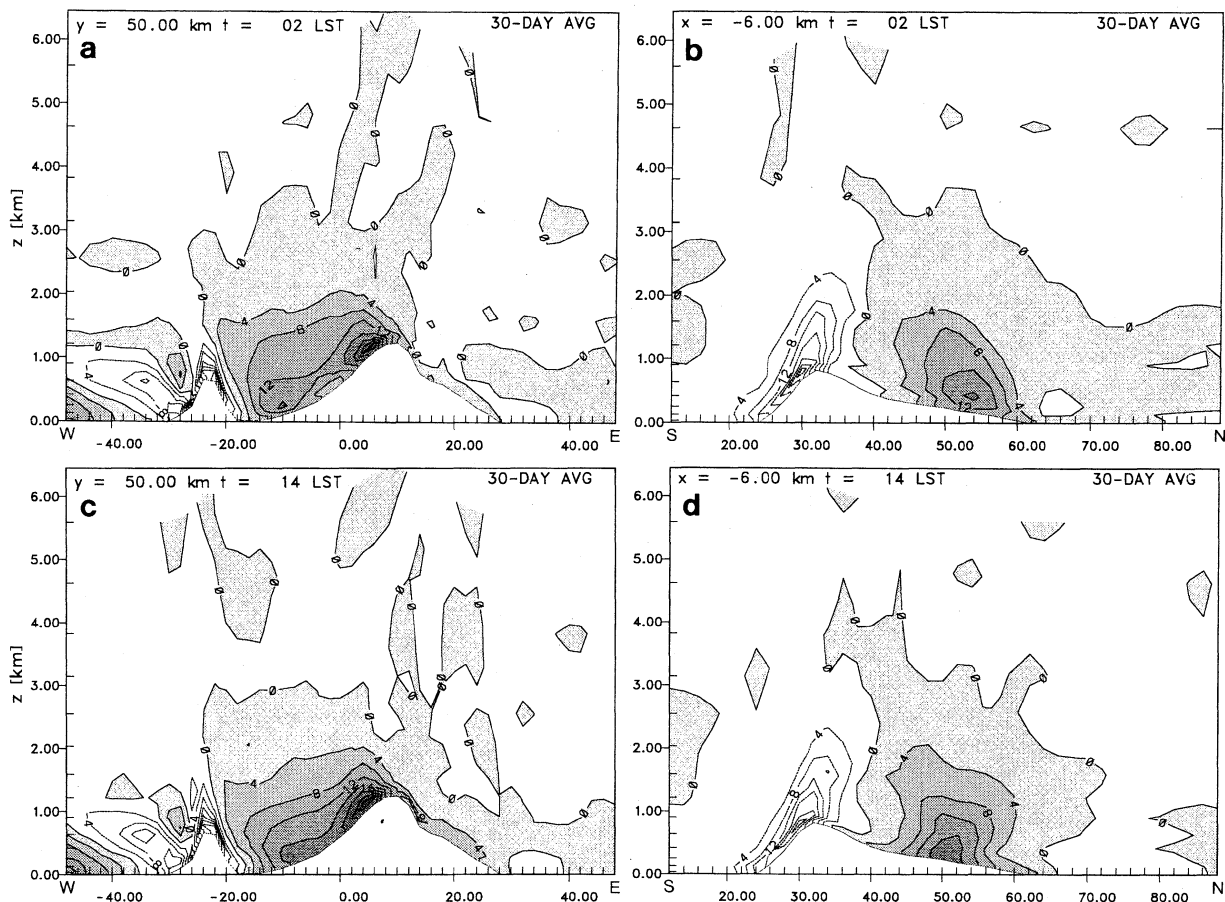


Fig. 8. Cross sections of 30-day mean relative vorticity at 0200 LST and 1400 LST. (a) and (c): along the east-west (A—A') line viewed from south toward north, and (b) and (d): along the north-south (B—B') line viewed from east toward west. Both lines approximately pass through the center of the vortex as depicted in Fig. 5a. The contour interval is the same as in Fig. 7.

the island of Hawaii during the Hawaiian Rainband Project (HaRP) was up to a height of 1700 m, while Kohala's wake in Hawaii was somewhat shallower than that.

The corresponding cross sections of *horizontal* wind vectors and speeds are illustrated in Fig. 9. Acceleration of the trades deflected southward by West Maui and channeled through the central valley is seen to reach an altitude of up to about 1000 m in the isthmus in both daytime and nighttime east-west cross sections. The north-northeasterly near-surface flow over the lower end of the leeward slope of Haleakala becomes stronger in the afternoon (Fig. 9c), which may be attributed to a sea breeze effect due to turbulent vertical mixing in the boundary layer created by island heating (Daniels and Schroeder, 1978; Pielke, 1984). In the north-south vertical plane, the winds are much weak across the center of the vortex. During the night, an area with the wind speeds less than 2 ms^{-1} extends up to the 2500 m level (Fig. 9b). In the afternoon, this area

of the weak winds recedes below the 2000 m level, while near the surface the westerly sea breeze of up to 4 ms^{-1} moves up the western slope of Haleakala (Fig. 9d), attaining a maximum depth of 500 m above the ground at around 1600 LST (not shown).

3.2 Comparison with station data

Figure 10 shows the 30-day mean near-surface model wind vector fields over the central valley at 6-h intervals. Also plotted are the 30-day mean observed wind vectors at ten observation stations in the valley maintained by Hawaiian Commerce and Sugar Company (HCS). Although measurements were generally taken at the usual "anemometer" level, it should be noted that there are some irregularities in the environment of individual HCS stations. For example, the wind directions at station No. 711 are generally more biased to the east compared to those at the neighboring station No. 717. Station No. 711 is rather exceptional in that its observation equipment was, in fact, set up at the top of a towering structure about 30 m high, which may ex-

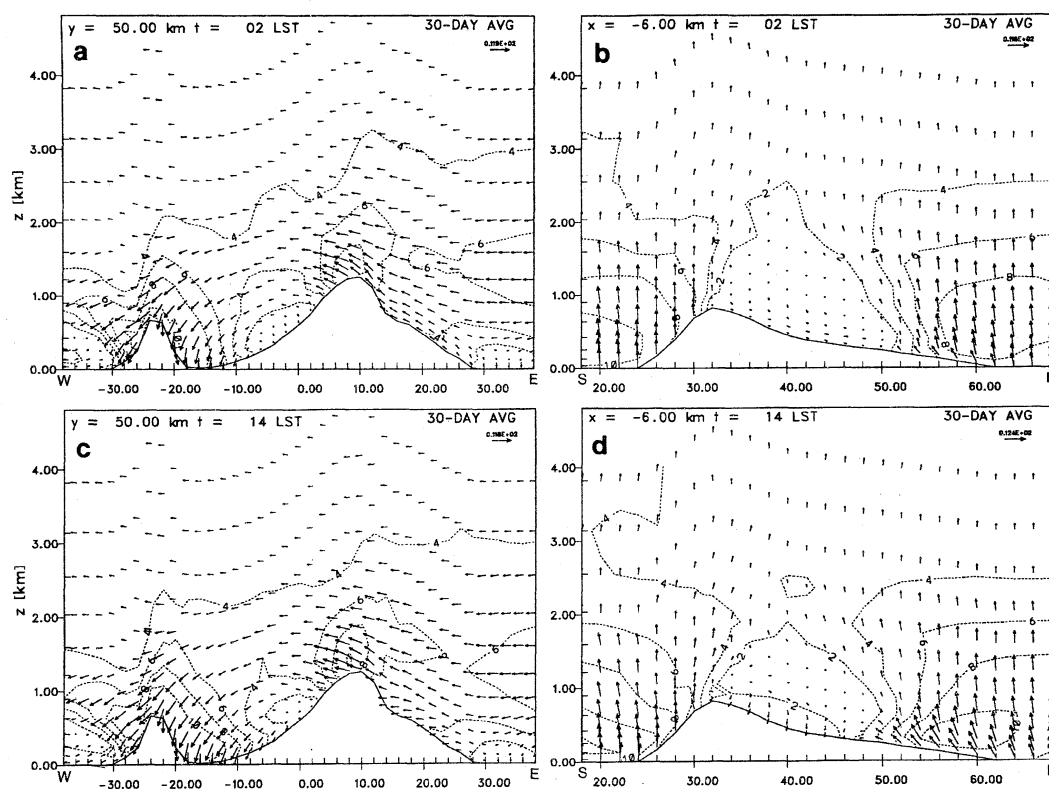


Fig. 9. As in Fig. 8, except for the corresponding *horizontal* wind vectors and speeds. For wind vectors the top and bottom sections of the frame of the panels (a) and (c) correspond to north and south directions, respectively, and the right and left sections of the frame correspond to east and west directions, respectively, so that a downward vector means northerly flow. In the panels (b) and (d) the top and bottom sections of the frame correspond to west and east, respectively, while the right and left sections correspond to north and south directions, respectively, so that a downward vector means westerly flow. The contours are drawn at an interval of 2 ms^{-1} for wind speeds (dotted).

plain the above differences. The model wind vectors plotted in Fig. 10, on the other hand, are considered to represent the winds at the mid-level ($\sim 38 \text{ m}$ above the surface) of the model layer closest to the surface of about 80 m thick. Taking into account these characteristics in the comparison, it appears that the agreement of the model wind speeds and directions with observations are generally good.

For close comparisons, time series of mean observed wind speeds and directions were plotted in Fig. 11 for the 1st-order station situated in Kahului (marked by the symbol * in Fig. 10) and 5 individual HSC stations. Also plotted are the model mean wind speeds and directions for a model grid point closest to each station that was selected subjectively. In the July 1988 dataset we had acquired for the 1st-order station No. 147 in Kahului, observations for 2100 LST to 0400 LST were missing. During the afternoon, the mean model wind at this station is about 20 per cent weaker than actually observed although the model wind directions are realistic. The 30-day mean 1000-hPa wind speeds at a NMC large-

scale analysis point over the Pacific Ocean upwind of Maui, where the effects of the islands are not likely to influence the analysis, are also shown in Fig. 11 by the symbols $\odot$ for 0200 LST (1200 UTC) and $\oplus$ for 1400 LST (0000 UTC). They indicate that the model wind speed at the Kahului gridpoint for 1400 LST is closer to the mean large-scale analysis value than to observation.

For HCS stations, we note that the model wind speeds are in general closer to observations during the daytime hours. The model speeds tend to vary less than observations do during the course of 24 hours and remain larger than observations during the night. Also we note that at grid points near the stations such as 711 (and 110, 201, 602, and 717, not shown) the mean model speeds are close to the mean analysis speeds. These stations are located in the midst of the main flow of deflected trades. The model seems to be unable to respond adequately to diurnal changes in the surface conditions at these grid points. It appears that the influence of steady trade winds dominates the flow regime at these grid

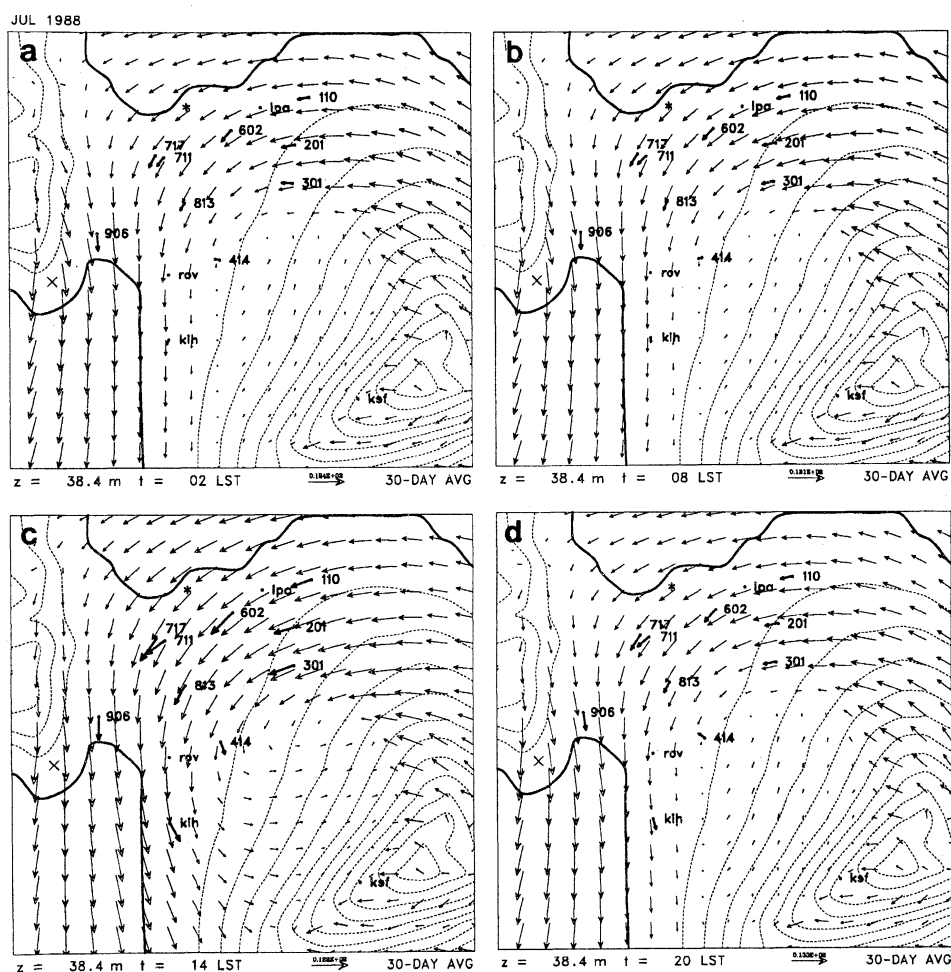


Fig. 10. As in Fig. 6, but for the near-surface model wind vectors (thin arrows). Also plotted are the 30-day mean observed wind vectors (thick arrows) at 10 observation stations in the valley maintained by Hawaiian Commerce and Sugar Company (HCS). Three other stations that are part of the HCS network (but for which data were not available) are also indicated [stations at Kula state forest (ksf), lower Paia (lpa), and Rover (rof)]. The mountain height contour interval is 250 m.

points due to insufficient spatial grid resolutions (the lowest model grid box is 80 m deep vertically and 2 km wide horizontally) and less detailed surface parameter specifications. The size of the model domains may also have an influence on the model results. We speculate that it may be necessary to use a vertical layer thickness on the order of 10–20 m or less to more realistically model the near-surface flow in this region dominated by the trades. Additionally, more detailed specifications of the surface characteristics such as roughness length, vegetation, and soil types would be required.

On the other hand, at grid points near the HCS stations, such as 301, 414, and Kihei (kih) (along with 813, not shown), the model mean wind speeds maintain relatively larger diurnal variations and are closer to mean observations. These stations are situated relatively inland away from the main path of

the deflected trades and close to the center of the Maui Vortex where the wind speeds are relatively weak and the direct influence of the deflected trades on the flow is much less. At these points the agreement between the model and observations can be quite good, as the case of the station 301 illustrates.

Both model and observed mean near-surface winds at the HCS station 906 near McGregor Point (shown in Fig. 10 by the symbol $\times$) in the leeward end of the central valley are, in general, stronger and more persistent and uniform compared with those at observation stations in the northern end of the valley such as the 1st order station at Kahului. Wind directions are also more persistent in the southern end of valley. Field observations by Daniels and Schroeder (1978) also exhibit similar features.

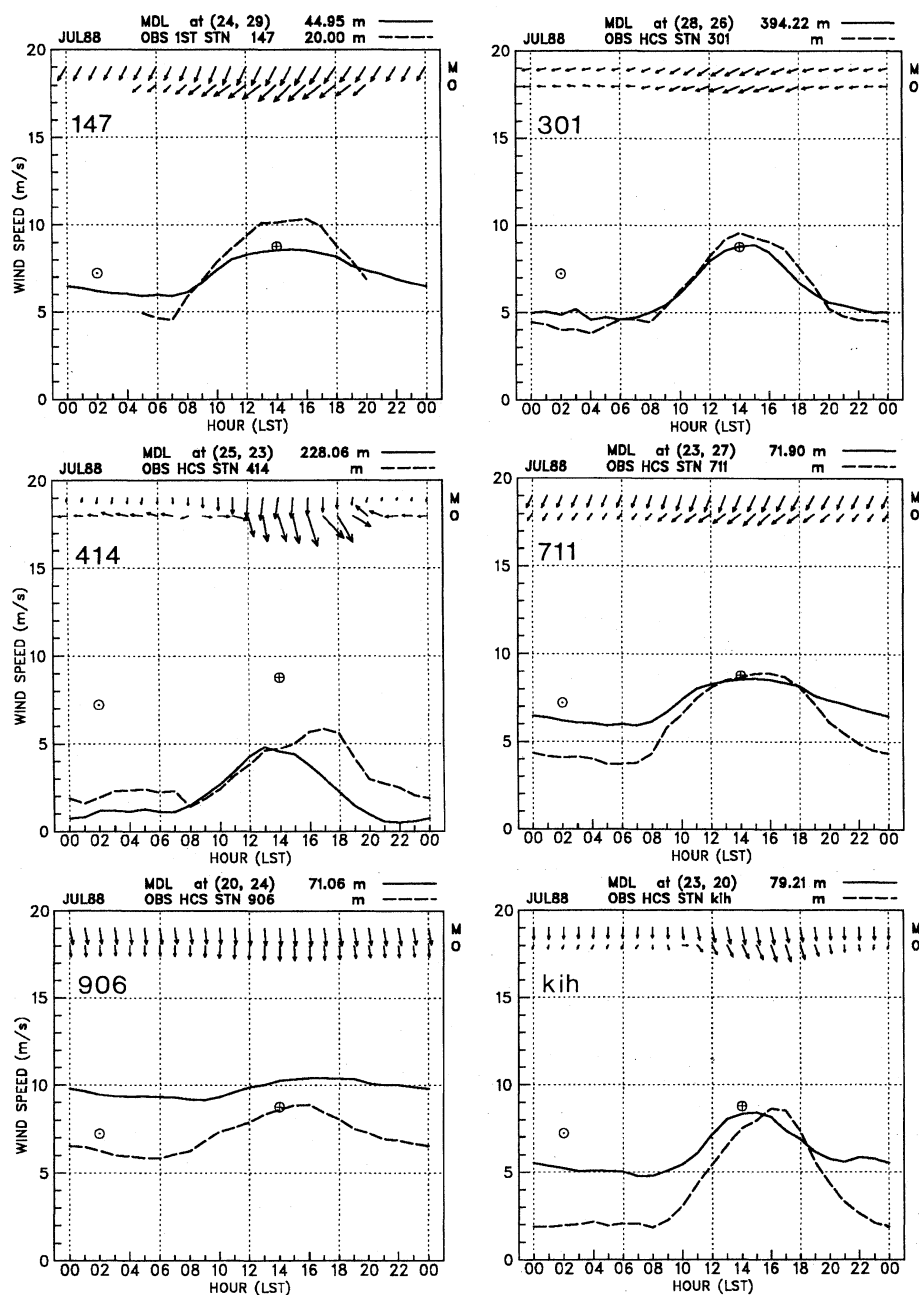


Fig. 11. 30-day mean hourly surface wind speeds and directions for the 1st-order station No. 147 in Kahului and 5 designated HCS stations in the valley (broken; $O \equiv$ observation), along with those for a grid point nearest each station (solid; $M \equiv$ model). 30-day mean 1000-hPa wind speeds at a NMC large-scale analysis point over the Pacific Ocean upwind of Maui are shown by the symbols $\odot$ for 0200 LST and $\oplus$ for 1400 LST (see the text).

3.3 The effects of solar radiation on the eddy formation

The effects of solar radiation on the formation of the Maui Vortex were tested in a separate simulation initialized after sunset. To mimic the external forcing without direct effects of solar heating, the large-scale analysis for 0200 LST 01 July 1988 was used as the initial model state for 2000 LST and also as the fixed lateral boundary conditions during

the 12-h integrations. Although downslope flow became dominant on the eastern slope of the valley within several hours, a weak eddy-like motion was still found in the valley at 0600 LST after 10 h of surface cooling (Fig. 12a). Because the starting time of simulation was set at 2000 LST, the surface heating would not have caused generation of the eddy initially. The streamline pattern for 0800 LST at the end of the 12-h integrations, on the other hand,

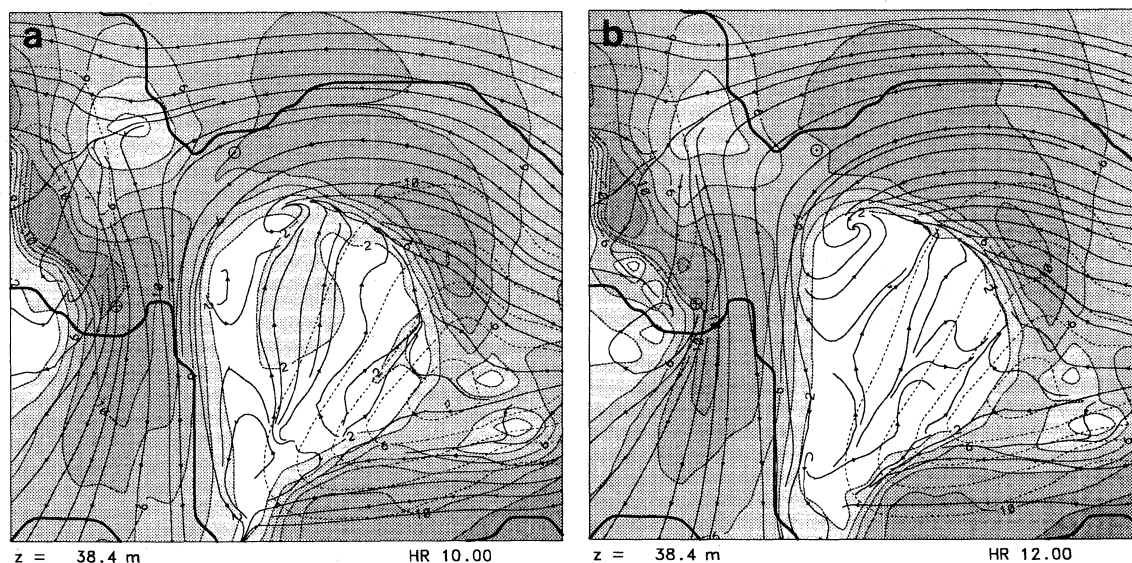


Fig. 12. As in Fig. 6, but for instantaneous wind speeds and streamlines showing the effects of cooling: (a) before sunrise and (b) after sunrise. See the text for the initial and lateral boundary conditions used.

clearly shows the formation of a new vortex in the central valley which must be initiated by the solar heating (Fig. 12b). Therefore, if an eddy-like circulation formed during the cooling period it must result from some different driving force than surface heating. It appears that the mechanical effects of West Maui in deflecting the easterly trades southward play a key role in maintaining vorticity against the destructive effects of downslope flow at night. During the night, the northerly flow through the isthmus plays the rôle of a wall in a sense and prevents the downslope flow from cascading toward the western shore, effectively creating a damming situation with cyclonic motion in the stagnant flow, even if there were no eddy motion resulting from the preceding daytime surface heating to begin with. We speculate that this cyclonic motion is formed by feeding, from the outside, positive vorticity generated in the boundary layer through the entrainment of the northerly flow on the west into the stagnant flow in the east. This argument follows Chopra's (1973) suggestion that viscosity plays an important role in the formation of a vortex street behind a bluff body which requires the Reynolds number as a governing parameter (Smith and Grubišić, 1993). If cyclonic motion existed during the preceding afternoon, vorticity fed by the northerly flow would help the vortex sustain its structure longer against the nighttime downslope flow, as the 30-day mean streamline patterns shown in Fig. 6 demonstrate.

3.4 The effects of West Maui on the vortex formation

Will a stationary vortex form in part due to solar heating, if West Maui were removed thereby eliminating the mechanically forced northerly flow through the isthmus? This question was examined by performing an experiment with a modified topography. In this experiment we used a topography with West Maui flattened to a uniform plain at 20 m MSL, along with the islands downwind (see Fig. 15 for the modified topography). The effects of surface heating and cooling of the original West Maui region therefore still exist in a modified form. The model was integrated for 6 model-days, starting at 0000 UTC 01 July 1988. Expanded views of the 5-day mean streamlines and wind speeds (excluding the initial 24 h) in the central valley with a flattened West Maui are shown in Fig. 13. At night the downslope flow dominates the lee slopes of Haleakala (Fig. 13a). Without West Maui, the northern branch of the trades split by Haleakala would flow through toward the southwest over the flattened West Maui. This northeasterly flow turns toward the south only after it reaches Auau Channel. Without the damming effect of the northerly flow through the isthmus, the nighttime downslope flow on the leeward side of Haleakala does not form a vortex over the "valley." As the solar heating starts, however, an eddy-like motion begins to form in the "valley," and during the daytime the northeasterly flow that goes over the flattened West Maui turns east over the channel and moves up along the western slope of Haleakala (Figs. 13b and 13c). The

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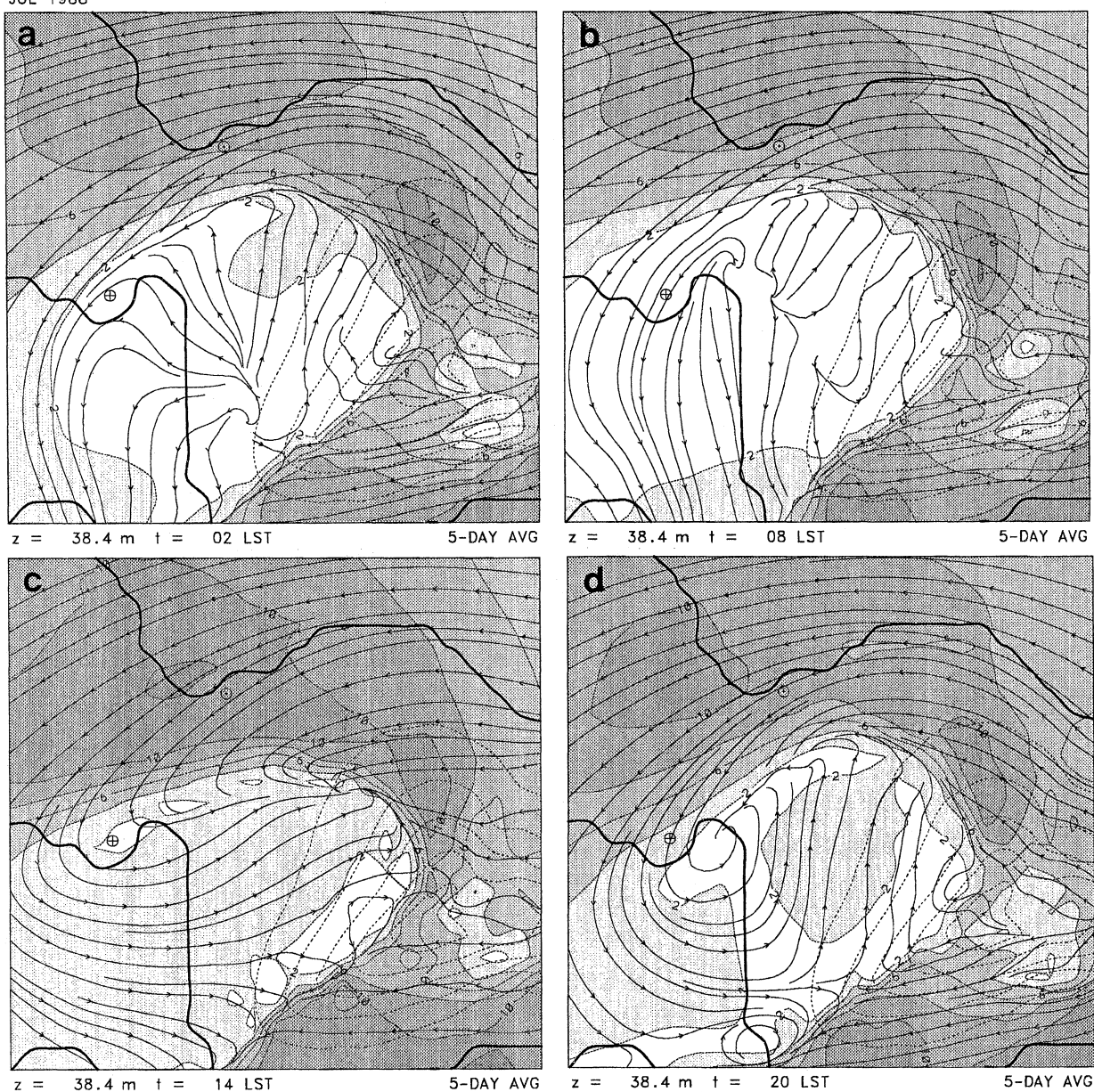


Fig. 13. As in Fig. 6, but for the 5-day mean fields of speeds and streamlines for the case with a topography of West Maui flattened to a uniform plane of 20 m MSL.

result is the formation of an elongated vortex expanding from the middle of the valley to Maalaea Bay. As the effects of surface heating diminishes after sunset, the flow over the lee slope of Haleakala turns from its westerly direction counterclockwise to become parallel and downslope relative to the mountain height contours. During this transition period in the early evening, a vortex will form which is centered in Maalaea Bay off Kihei (Figs. 13d). Downslope flow dominates the lee slope as cooling intensifies, and eventually wipes out the vortex (Fig. 13a). The results of this experiment indicate that even without the blocking effect of West Maui a vortex can be generated in the lee of Haleakala because of the surface heating, which acts as the driving force.

However, the vortex is transient and lacks the steadiness of the Maui Vortex found in the simulation with West Maui.

3.5 No-vortex days

Out of 31 model days, only two days failed to generate the Maui Vortex. Those days were 24 to 25 July 1988 local time. The large-scale objective analysis indicates that at 0200 LST (1200 UTC) 24 July 1988 there was a weak surface low pressure center situated over the islands accompanied by a distinct cyclonic motion at the 500 hPa level and centered just north of the islands, while a strong high surface pressure core was located northwest of the islands. As the near-surface synoptic-scale flow became more

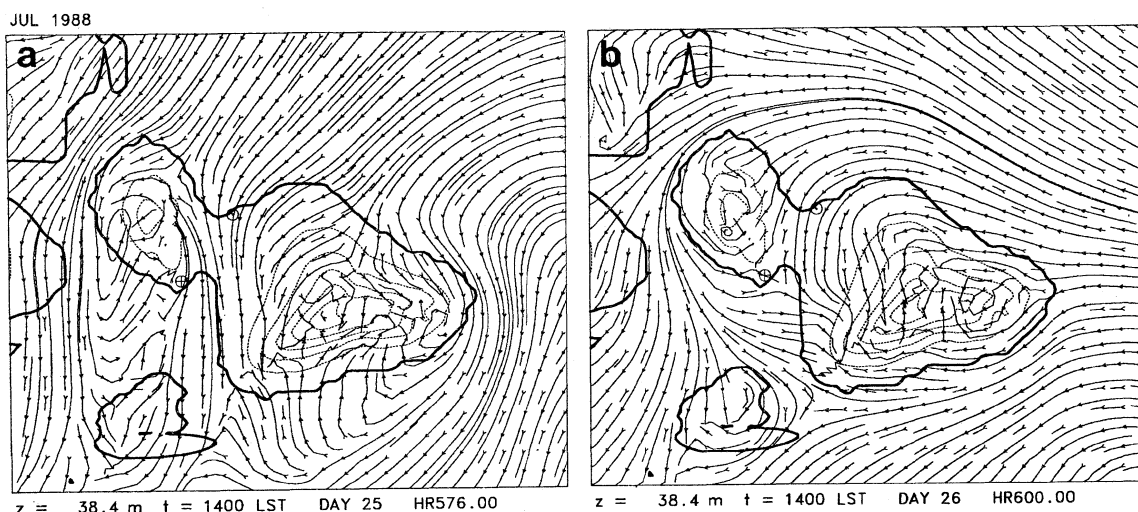


Fig. 14. Near-surface streamline patterns for 1400 LST on the “no-vortex” days of (a) 24 July, and (b) 25 July 1988.

northeasterly during the next 12 h, the Maui Vortex began to fade away. The model streamline pattern in Fig. 14a shows the northeasterly surface flow passing over the central valley without being deflected much by either West Maui or Haleakala, thereby making the leeward slopes (western slopes) of Haleakala on the “vortex” days become the windward slopes on this “no-vortex” day. During the following 48 h, as the surface trough along with the depression at 500 hPa level moved westward, easterly to east-southeasterly environmental flows near the surface and southeasterly flows at higher levels became dominant, causing West Maui to be in the “shadow” of Haleakala. This synoptic condition created the northerly near-surface flow passing through the valley and the flow going around West Maui (Fig. 14b). In this flow situation, the Maui Vortex did not form, and it is apparent that this no-vortex situation had resulted from these surface as well as upper-level synoptic disturbances over the islands that the rest of the model simulation period did not experience.

4. Vortex shedding

Mesoscale atmospheric vortex shedding in the lee of large islands has been documented by satellite pictures of cloud patterns at the top of the mixed layer for various geographical locations (see, *e.g.*, Chopra and Hubert, 1965; Chopra, 1973; Jensen and Agee, 1978; Etling, 1989 and 1990). In Hawaii, Nickerson and Dias (1981) first documented wake vortices resembling a von Kármán vortex street behind the island of Hawaii using the data collected during the Hawaiian Mesoscale Energy and Climate (HAMEC) project. A field observation of wake vortices behind this island was also conducted during the HaRP project (Smith and Grubišić, 1993). Recent studies in an attempt to advance our understanding of

mountain wakes in Hawaii include Smolarkiewicz and Rotunno (1989), Schär and Smith (1993a and 1993b), and Sun and Chern (1994). In Maui, the low-level cloud motions that suggested the existence of the Maui Vortex were first filmed in the 1940s (Schroeder, 1993). The only *in situ* observation of the Maui Vortex that has been reported was conducted by Leopold (1949) (see Fig. 3). However, shedding of the Maui Vortex has not been documented either by satellite or field observations.

Development of a mesoscale vortex street is influenced by stratification as well as rotation (Etling, 1990), the effects of which can be characterized, respectively, by the Froude number

$$Fr = \frac{U}{NH}, \quad (1)$$

and the Rossby number

$$Ro = \frac{U}{fD}, \quad (2)$$

where U is the uniform flow speed, H the height of the obstacle, $N = \left(\frac{g}{\theta} \frac{\partial \theta}{\partial z}\right)^{1/2}$ is the Brunt-Väisälä frequency, θ the potential temperature, f the Coriolis parameter, and D the base diameter of the obstacle. Etling (1990) noted that mesoscale atmospheric vortex shedding from large islands has been found for $Fr < 0.4$ and for Ro in the range of 2 to 4, implying that shedding has been observed in the situations in which strongly stratified flow is influenced by moderate background rotation ($Fr < Ro$).

The 30-day mean values of model wind speed U and potential temperature lapse rate $\partial \theta / \partial z$ in the lowest 1 km of the undisturbed trades upstream of Maui for July 1988 were estimated to be about $7\text{--}9 \text{ ms}^{-1}$ and $3.0\text{--}4.0^\circ \text{K km}^{-1}$, respectively. The mean Fr and Ro on the windward side of Haleakala were thus estimated to be 0.21 to 0.31, with $H =$

Table 1. Geometrical data (φ = latitude, H = height, D = base diameter, H/D = aspect ratio) and flow characteristics (U, R_o, Fr) for cases of mesoscale vortex shedding from large islands (reproduced from Etling, 1990).

Island	$\varphi(^{\circ})$	$H(\text{km})$	$D(\text{km})$	H/D	$U(\text{ms}^{-1})$	R_o	Fr
Jan Mayen	71	2.2	20	0.11	10	3.6	0.36
Aleutian	55	2.1	28	0.08	7	2.1	0.20
Madeira	33	1.8	34	0.08	5	2.0	0.19
Cheju	33	1.9	43	0.05	8	2.5	0.31
Hawaii	20	4.2	120	0.04	10	1.6	0.20
Maui	21	3.0	43	0.07	7–9	3.3–4.2	0.21–0.31

2900 m, and 3.3 to 4.2, respectively. These values of Fr and R_o were subjectively obtained as U and $\partial\theta/\partial z$ and hence the stability, N , were not uniform with height. As shown in Table 1, the mean values of R_o in our 30-day simulation is at the upper limit of R_o for the observed cases of mesoscale vortex shedding from large islands (Etling, 1990), while Fr is within the observed range. For the above mean values of Fr (< 0.4) for Maui, we may therefore expect vortex shedding to occur for moderate values of R_o .

Examination of the 30-day hourly time series of the surface wind vector and streamline patterns, however, reveals that the Maui Vortex is stationary and there are no apparent indications of the so-called “vortex shedding” taking place. It appears that the same northerly flow through the isthmus that creates a damming situation with cyclonic motion in the stagnant flow at night also plays a role in preventing the vortex shedding from occurring by blocking the path along which the vortex would migrate westward toward the ocean.

4.1 Vortex shedding without West Maui

An hourly time series of the near-surface streamline patterns from the experiment with the flattened West Maui described in Section 3.4 was examined to find out whether vortex shedding was inhibited by these mountains. We found that, starting in the mid-afternoon of the 5th day, an anticyclonic eddy began to regularly drift away from the southern end of Haleakala while a cyclonic eddy began to form over the “central valley” (see Fig. 15). During the next 20 h or so through the morning of the 6th day, vortices of opposite rotations were seen to form and to be shed from both the northern and southern ends of Haleakala. During this period the trades were east-northeasterly to nearly easterly. Fr and R_o were estimated to be about 0.30 and 3.5, respectively, with $U = 8 \text{ ms}^{-1}$ and $\partial\theta/\partial z = 2.5^{\circ}\text{K km}^{-1}$. These parameters are within the ranges for the observed cases of atmospheric vortex shedding and indicate the possibility of vortex shedding of “the Maui Vortex” if the West Maui Mountains were replaced by a low flat land. Prior to the above period of shedding, Fr and R_o were also mostly within the

observed ranges given in Table 1, and rather sporadic and poorly organized shedding was detected in the near-surface flow patterns as eddies occasionally formed at northern and southern ends of Haleakala which did not show up in the mean flow fields depicted in Fig. 13.

By contrast, the corresponding flows in the experiment with unmodified topography, both before and during the period of shedding, behaved much like the patterns shown in Fig. 6 with a well-defined northerly flow and vortex over the isthmus and there was little indication of vortex shedding.

5. Discussion

The hypotheses proposed by Reisner and Smolarkiewicz (1994) state: 1) that for $\eta^* \sim Fr < 1$, where η^* represents a magnitude of thermal forcing which depends on a heating rate, along with U, H, D , and $\partial\theta/\partial z$, as defined by them, there is a domain in the $Fr - \eta^*$ régime in which upwind flow exhibits upwind stagnation during the entire diurnal heating cycle (see their Fig. 4), and 2) that thermal effects should be substantially more pronounced in the lee than over the upwind side of the mountain. The values of Fr and η^* for Maui estimated by the above authors based on HaRP observations, 0.17 and 0.13, respectively, would place Maui in the domain where stagnation and splitting of the flow upwind of the obstacle are expected. Streamlines and speeds of the 30-day mean flow past Maui shown in Fig. 16 indicate the existence of the stagnation and splitting of the surface flow upwind of Haleakala, even at the time of peak surface heating. Also apparent are the lee eddies in the wake of the West Maui Mountains. These distinctive features of low Froude number flows found in the results presented here are consistent with Reisner and Smolarkiewicz’s (1994) hypotheses and also with their model simulation results.

Figure 17 illustrates the cross sections of potential temperature and relative humidity for 0200 LST 14 July 1988 along the east-west line (A—A’) across the Maui Vortex as depicted in Fig. 5a. The corresponding field of wind speed (not shown) is similar to that illustrated in Fig. 9a, but with much stronger speeds

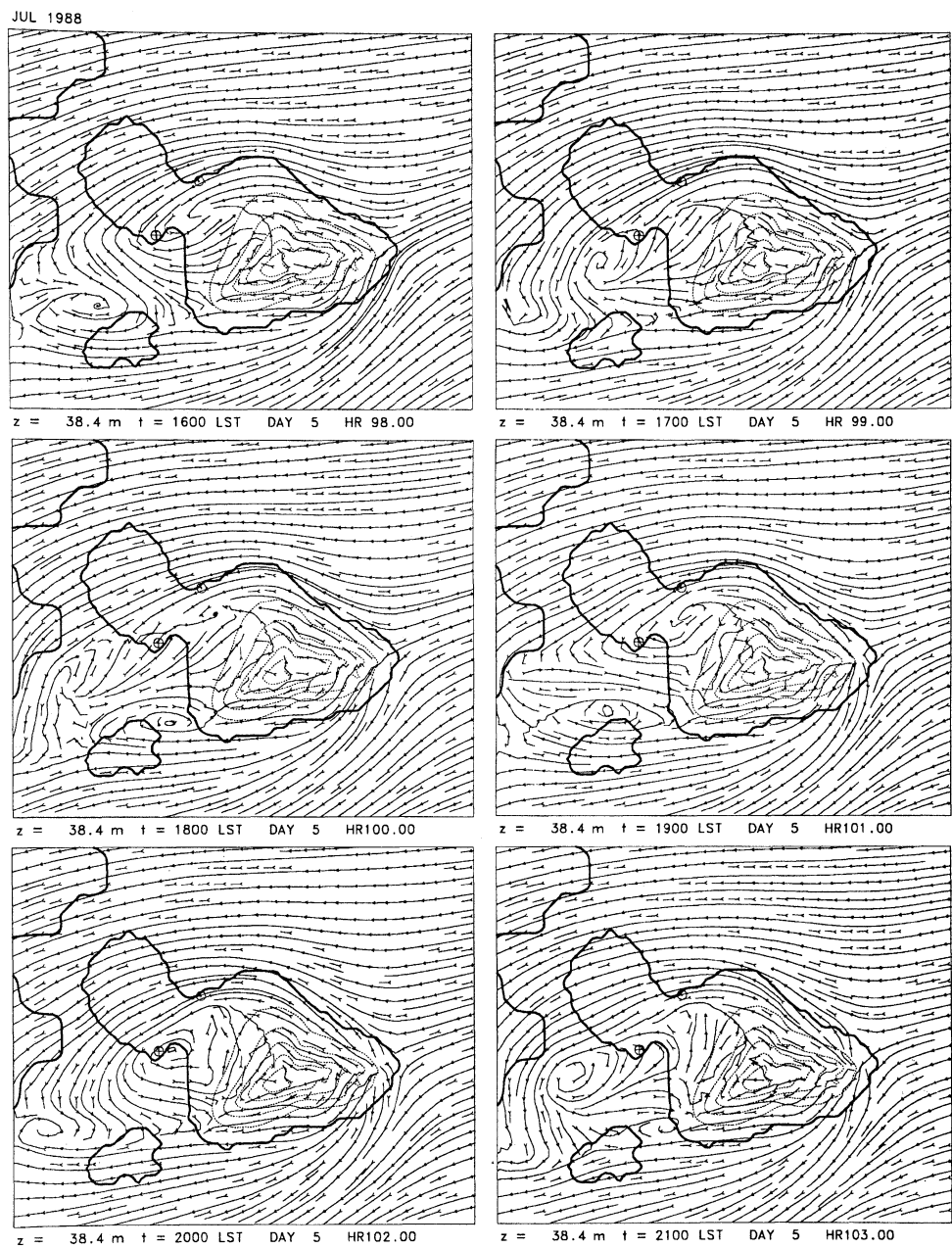


Fig. 15. Hourly near-surface streamline patterns for 1600 LST to 2100 LST 04 July 1988, showing vortex shedding in the wake of Haleakala in the case with West Maui flattened to a uniform plain at 20 m MSL.

reaching over 14 ms^{-1} above the east mountain ridge and less than 2 ms^{-1} along the mid-slope on the lee of Haleakala. We note hydraulic jump-like discontinuities in these quantities over and immediately downstream of the obstacle crests. This situation began to develop in the afternoon of 13 July 1988 and remained stationary through the next morning. Similar jumps existed in these quantities during most of the simulation period, and it appears that an abrupt increase downstream in potential temperature and decreases in wind speed and relative hu-

midity are indicative of hydraulic jumps associated with the Maui Vortex. It is suggested that the above situation may qualitatively correspond to a case of the lee jump described by Houghton and Kasahara (1968) (see their Figs. 3 and 14b) in which the lee jump remains stationary on the downstream side of the obstacle crest. Smith and Grubišić (1993) found similar jump-like discontinuities in these quantities associated with Hawaii's wake along the ENE to WSW flight tracks flown at 1800 m MSL over the northern and southern tips of Hawaii during HaRP.

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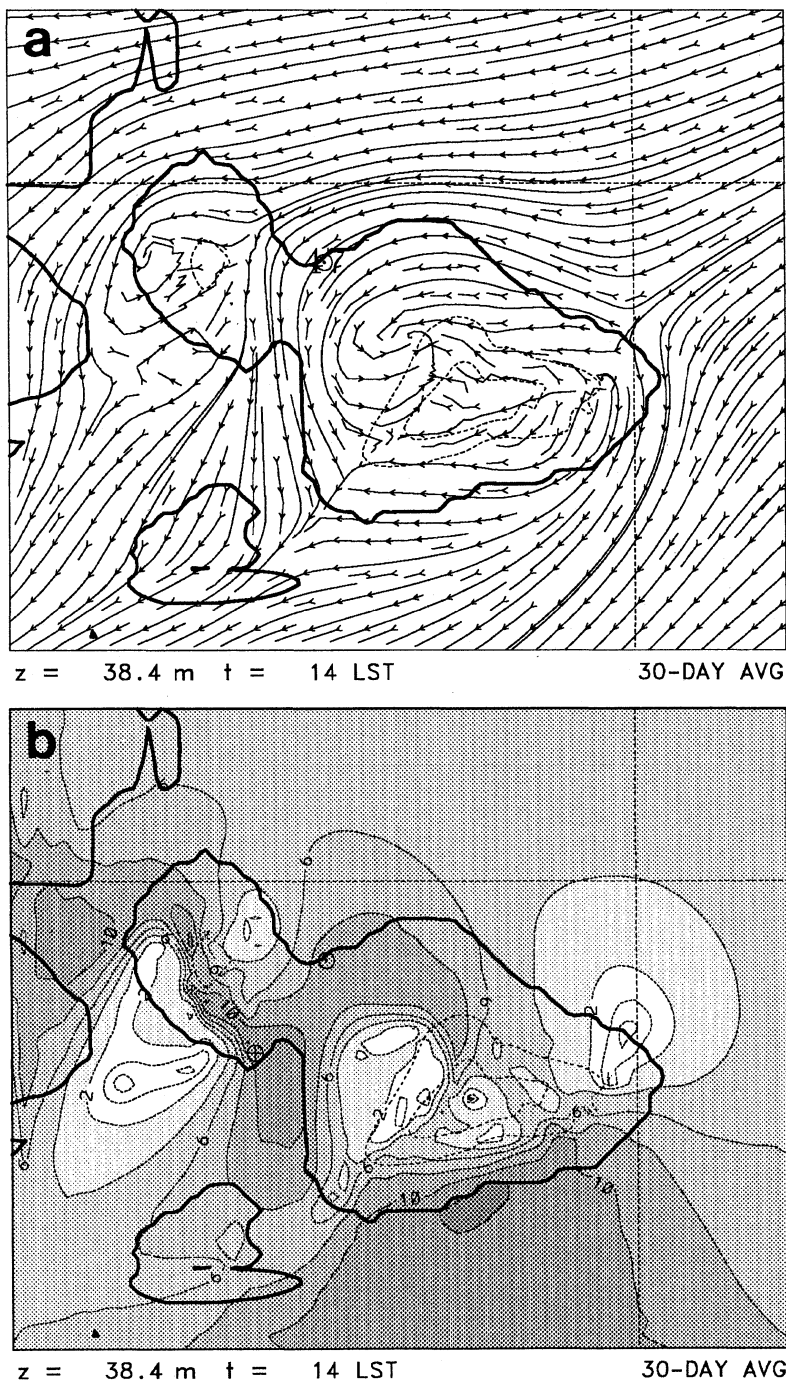


Fig. 16. 30-day mean near-surface patterns past Maui of (a) streamlines, and (b) wind speeds (contour intervals: 2 ms^{-1} ; additional contours at 1 ms^{-1} and 0.5 ms^{-1}) for 1400 LST, showing flow stagnation and splitting upwind of Haleakala and vortices downwind of West Maui. Mountain height contour intervals are 1000 m.

Measurements at this height indicated the intrusion from above of potentially warmer and drier air along the flight path. From a practical viewpoint, drying in the surface layer could decrease the moisture content in wildland fuels, thus increasing the potential of wildfire.

It is suggested from the results presented in Section 3.5 that field burning would have least effects on the air quality in the central valley on the “no-vortex” days similar to 24–25 July 1988, which were the result of a synoptic disturbance passing over Hawaii. The model results presented in previous sec-

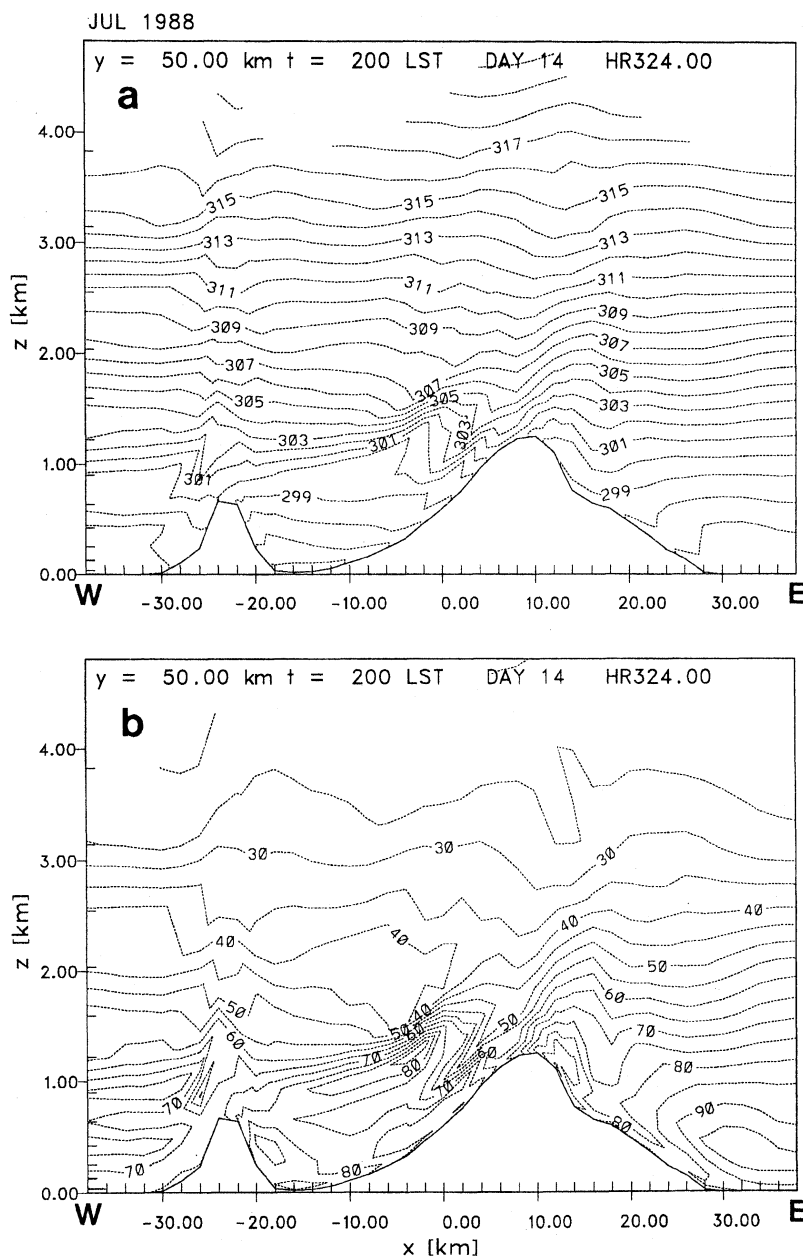


Fig. 17. Cross sections of: (a) potential temperature ($^{\circ}\text{K}$), and (b) relative humidity (%) for 0200 LST 14 July 1988, showing hydraulic jump-like structures, along the east-west line (A—A') across the Maui Vortex as depicted in Fig. 5a, viewed from south toward north.

tions, however, have indicated that under the east-northeast synoptic flow situation that persisted during July 1988 the Maui Vortex would be maintained, steady, and stationary in the central valley throughout the day and night, although its intensity may vary depending on the time of the day. The stationary Maui Vortex would make pollution concentration in the central valley even worse, since smoke from agricultural burns may be trapped in the vortex and may not be dispersed periodically toward the open ocean.

Climatologically, about 85 per cent of the July flow patterns consists of this anticyclonic situation (Type 1a). The rest of the trade wind situations, about 12 per cent in July, consist of a high-pressure cell to the northeast of the islands (Type 1b), bringing southeast winds over Maui. In the latter situation, West Maui would be in the “shadow” of Haleakala, and the Maui Vortex may not form. Therefore, the Type 1b situation may be a better choice for field burning in the limited sense that the eddy is less likely to form to trap pollutants. It is

not certain in this case whether the dominant surface flow in the valley will be northerly, since the southeasterly trades in a typical Type 1b situation may not produce the surface flow patterns in the valley similar to those in “no-vortex” days shown in Fig. 14. It should be noted that the situations aloft would also have significant effects on the regional surface flow patterns in any synoptic types of surface flow situations. In the Type 1a situation, it would be desirable to choose a day with a relatively stronger northerly component in the synoptic surface wind, since stronger northerly ambient flow would be likely to reduce the chance of formation of the persistent Maui Vortex.

In other types of synoptic flow pattern situations, “purely cyclonic situations” and “mixed situations” as defined by Yeh *et al.* (1951), the Maui Vortex is not likely to form. These situations, unfortunately, are not very frequent during the summer months. However, during the spring months the frequencies of these flow patterns range from 30 to 40 per cent. Additionally, the flow pattern of Type 1b has about a 15 per cent of frequency throughout the spring. Therefore, it is likely that a “good” burning day will be found within a time frame of, say, any one specific week. A similar statement can be made for the autumn and winter months during the burning season (March–December).

Numerical simulations of Hawaii’s wake by Smolarkiewicz *et al.* (1988) and Ueyoshi and Han (1991) both predict shedding eddies behind Hawaii. In the latter simulation, the Froude number upstream of Hawaii, calculated from the composite wind and temperature profiles fixed in time and space, was $Fr < 0.4$. On the other hand, analyses of the HaRP data indicate that the main wake of Hawaii consists of two elongated counter-rotating quasi-steady symmetric eddies with little indication of oscillating flow fields (Smith and Grubišić, 1993). The Froude number during HaRP ranged between 0.3 and 0.8 with an average value of $Fr \approx 0.5$. It is suggested that the above discrepancy between observations and simulations may be due to the simplified topography included in the models. One problem with the simulation by Ueyoshi and Han (1991) is that the island of Maui was not included in the model topography, which would have further accelerated the trade wind moving through Alenuihaha Channel between Maui and Hawaii. We speculate that this exclusion of Maui may have had a significant influence on simulated vortex shedding behind Hawaii because the accelerated northeasterly trade wind through the Channel might have deterred vortex shedding in a mechanism similar to that in the Maui Vortex case.

Extending the above argument, it is suggested that the eddy generated downwind of the Kohala Mountains in the northern peninsula of Hawaii may

be stationary, even if $Fr < 0.4$, if the accelerated northeasterly trade wind flowing through Alenuihaha Channel has similar effects on Kohala’s wake.

6. Summary

In this study we attempted to describe the spatially as well as temporally continuous short-term climatology of the airflow in the central valley of Maui during July 1988, a summer month under typical trade wind conditions, with the use of a high-resolution non-hydrostatic model initialized and forced by large-scale objective analyses. Our immediate objective was to simulate the so-called “Maui Vortex,” which influences air quality during the sugarcane burning season, and wildfire potential in the fire season.

Month-long (30-day) averages of near-surface model flows in terms of speeds and directions generally compare favorably with those of measurements taken at 10 HSC observation stations in the valley. Model speeds are, in general, close to observations during the daytime hours but remain larger than observations during the night. However, at the HSC stations situated relatively inland away from the main path of the deflected trades and close to the center of the Maui Vortex, the model speeds are in better agreement with observations. To improve the simulations even more, the model may require an even finer spatial resolution and more detailed specifications of the surface characteristics.

Under the persistent northeasterly trade wind conditions (Type 1a) during July 1988, the Froude number in the low-level flow upwind of Haleakala was generally less than 0.4, indicating that the conditions suitable for generation and possibly shedding of the vortex in the lee of Haleakala mostly existed in the ambient basic flow. The model results indicate that under the above condition the Maui Vortex would form, and it would remain steady and stationary in the central valley regardless of the time of the day with little indication of Maui-Vortex shedding, although its intensity may vary. Two major factors, solar heating and the accelerated northerly flow over the valley resulting from deflection of the trades by West Maui, appear to contribute to the formation and maintenance of the Maui Vortex. Without solar heating, an eddy-like circulation still forms in the valley under nighttime cooling conditions. It appears that the northerly flow feeds vorticity to the stagnant flow to the east and provides at the same time a damming effect to maintain the eddy in the valley. However, its intensity is much weaker compared to the eddy formed under the diurnal heating cycle. If West Maui were removed, thereby eliminating the intensified northerly flow through the valley, the cyclonic motion would disappear in the mean flow during the night but would come back in an elongated form in the lee of Haleakala as soon

as surface heating begins in the morning. Without West Maui, the eddy becomes transient and loses its steadiness. We also found that vortex shedding can occur in the wake of Haleakala with $Fr < 0.4$ if West Maui were removed from the model topography, indicating an important rôle of the northerly deflected flow in maintaining the steady, stationary vortex in the valley.

The Maui Vortex did not form during a two-day period of a synoptic disturbance over Hawaii which created the ambient flow situations in which the central valley became the "wind-ward side" of Heleakala or West Maui was in the "shadow" of Heleakala. However, an occasion like this instance would seldom occur during the summer months. For field burning, other types of flow situations including the southeasterly trade situation (Type 1b) may be preferable to Type 1a because the smoke-trapping eddy in the central valley should be less likely to form in this type than in the Type 1a situation.

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貿易風下におけるマウイ渦の数値実験

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マウイ島ではハレアカラ山の風下にマウイ渦が持続的に生じて、サトウキビなど収穫の際の野焼による大気汚染が中央谷平原でより複雑なものとなる。この論文では、ある夏の一ヵ月を例にとり時間・空間に連続して時間積分した結果から、貿易風下のマウイ島の中央谷平原における大気流のclimatologyを調べた。ここでは、大規模場の客観解析によって初期値と境界値を求めて、高精度のメソスケールモデルによってシミュレーションを行った。こうして求めた結果を観測点の観測値と比較した。持続する貿易風下において、中央谷平原で渦が発生した。フルード数 $Fr(=U/NH)$ 、 U は一様風の大きさ、 H は障害物の高さ、 N はブラント・パイサラ周波数が0.4よりも小さい場合でも渦の放出(vortex shedding)は起こらず、一日の時刻に関係なく形を変えずに動かなかった。しかしながら、西マウイ山をモデルから取り除くと、渦の放出が起こるようになった。マウイ渦の発生や持続およびその定常性に寄与する主要要素として、日射と、貿易風が西マウイによって曲げられて生ずる北からの加速された流れがあり、これらについて議論した。さらに野焼きに好都合な総観場の条件について提案を行った。