**Interactions of a mesoscale katabatic flow with a small crater basin to produce cold and warm air intrusions, flow bifurcations and a hydraulic jump**

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[presently 8137 words; 7500 word limit not counting abstract, figure captions or references)

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NOTES

1. Figures are pasted up composites of multiple figures. They need work.

2. "Comparison with radiosoundings at BASE revealed that the tethersonde at BASE had a cold bias of about -0.4 °C above about 70 m AGL". Has this bias been corrected or are we presenting uncorrected data?]

3. We have a word count problem. Thus, I have omitted anything to do with pressure observations. Also have reduced information on the wake.

4. Wind components reported along two axes (15° axis and 35°). This needs an explanation.

There were changes between present tense and past tense. I changed everything to present tense because the text started in present tense.

I have suggested a few cuts and I also think that the text could be shortened considerably by avoiding repetitions. This refers mostly to descriptions of the upstream flow and boundary layer, which are repeated multiple times throughout the manuscript.

Not sure about the separation of a bifurcation / WAI phases. I think WAIs are triggered/ intensify when the waves (CAI/WAI) merge. It would be nice to have some argument what triggers the wave / wave amplification. (Fig 7 d shows a weak bifurcation, I think)

Abstract (≤ 250 words, presently 246 words)

Observations from the Second Meteor Crater Experiment (METCRAX II) on the night of 19-20 October 2013 are used to investigate the interactions between a regional scale katabatic flow and Arizona's Meteor Crater. The crater is a 1.2 km wide, 170-m-deep, near-circular basin with a rim that extends 30-50 m above the surrounding plain. On clear, quiescent nights a southwesterly katabatic flow lifts the stable boundary layer (SBL) on the 1° tilted plain over the crater rim to produce a continuous inflow of negatively buoyant air that runs down the crater's inner sidewall. As the SBL deepens and the katabatic flow strengthens, a flow bifurcation forms above the rim with the negatively buoyant lower portion running down the inner sidewall to produce a hydraulic jump and the neutrally buoyant upper portion carried quasi-horizontally over the basin. Unsteady short waves form in this upper flow behind the rim. As the stable layer continues to deepen and a mesoscale downslope flow develops above the katabatic flow, a high-amplitude short wave develops in the lee of the rim bringing warm air from the elevated residual layer downward into the basin. Interactions between the continuous cold-air intrusion and the descending lee wave accelerates the flow down the slope to enhance the hydraulic jump, which then reaches vertically to merge with the rising air in the ascending portion of the lee wave. The strong winds penetrate to the basin floor, displacing and stirring the pre-existing, intensely stable, cold pool and creating warm air streaks.

Key words: Meteor Crater, Arizona, basin, katabatic flow, hydraulic flow, hydraulic jump, cold air intrusion, bifurcation, warm air intrusion, lee wave

**1. Introduction**

The first Meteor Crater Experiment (METCRAX I, Whiteman et al. 2008) was conducted in Arizona's Barringer Meteorite Crater (AKA Meteor Crater) in October 2006 to investigate the nocturnal buildup of stable boundary layers in the crater basin. This experiment resulted in the serendipitous discovery of an unexpected phenomenon - the occasional occurrence of strong turbulent nocturnal flows near the base of the upwind inner sidewall. While the observational network was not ideally placed to investigate these flows, initial analyses identified the phenomenon as a downslope windstorm-type flow (Adler et al. 2012) in which warm air descended into the lee of the upwind crater rim from aloft. A Second Meteor Crater Experiment (METCRAX II, Lehner et al. 2016a) was then designed to investigate this phenomenon and the related cold-air intrusions that come continuously over the crater rim from the southwest at night as a katabatic flow develops on the adjacent plain. METCRAX II was conducted in October 2013 and the strongest, best-observed downslope windstorm-type flow, which occurred during the 4th Intensive Observing Period (IOP4) on the night of 19-20 October 2013, is the topic of this paper. Scientific questions that are of special interest are the evolutionary phases of the flow development and the relationships between the changing atmospheric structure inside the crater and the upstream conditions.

 The paper is structured as follows: **Section 2** surveys literature on the current understanding of similar phenomena observed in other locations and in the Meteor Crater. The topography, ground cover, synoptic situation and experimental setup are described in **Section 3**. **Section 4** presents analyses, **Section 5** provides further detail on the most interesting features of the phenomenon and **Section 6** provides a summary and conclusions.

**2. Background**

As air flows past three-dimensional, isolated topography, the resulting interactions with the terrain can produce multiple flow responses, including a splitting of the flow around the obstacle; flow over the obstacle; the formation of mountain waves, wave breaking, and downslope windstorms over and in the lee of the obstacle; and the formation of wakes (e.g., Smith 1989). Strong, gusty downslope windstorms are one of the more high-impact phenomena associated with airflow over terrain. Downslope windstorms have been observed throughout the world in regions of complex terrain, for example, in the European Alps (e.g., Grisogono and Belušić 2009; Richner and Hächler 2012; Armi and Mayr 2007, 2015; Mayr and Armi 2008), in the Pyrenees (e.g., Bougeault et al. 1993), in the Rocky Mountains (e.g., Lilly 1978), and in the Sierra Nevada (e.g., Grubisić et al. 2008; Armi and Mayr 2011). Three different concepts have been proposed to explain their formation (Jackson et al. 2012). Hydraulic theory is based on an analogy between water flowing over obstacles and air flowing over mountains. The flow layer thins and accelerates as it transitions from subcritical upstream of a mountain to critical at the ridge and supercritical downstream (Long 1954, 1955; Houghton and Kasahara 1968; Smith 1985; Armi and Farmer 2002; Mayr and Armi 2010; Winters and Armi 2012, 2014). The second and third concepts are based on upward propagating mountain waves and a downward reflection of wave energy at a critical level that produces wave amplification. The two concepts differ in the origin of the critical layer. A critical layer can be present in the ambient flow, where the wind component perpendicular to the mountain range goes to zero (Klemp and Lilly 1975) or it can be self-induced as a result of wave breaking (Clark and Peltier 1977; Peltier and Clark 1979).

 The downslope windstorm literature has focused primarily on the interactions of synoptic-scale flows with mountain ridges to produce windstorms over the lee slopes. Such synoptically driven windstorms can occur any time of day. At the crater, in contrast, relatively strong gusty winds occur frequently on the upwind inner sidewall of the basin on clear or partly cloudy nights when the background synoptic flows are weak and katabatic flows are present on the adjacent plain. These winds do not occur every night and sometimes appear intermittently on a given night and, while weaker than typical downslope windstorms, share some of theircharacteristics.

 The occurrence of strong gusty winds on the lower inner upwind sidewall of the Meteor Crater was the motivation for two sets of idealized numerical simulations. Lehner et al. (2016b) simulated the effects of basin size, upstream flow, atmospheric stability and uniform background winds on flow regimes in a basin embedded on a 1° tilted plain using 2-dimensional simulations in which a katabatic flow developed naturally on the tilted plain. Flow responses were of four types: wakes over the upstream basin sidewall, sweeping of the basin atmosphere, waves over the basin, and stagnation over the upstream basin sidewall accompanied by a hydraulic jump. Rotunno and Lehner (2016) determined the two-dimensional flow regimes that might occur within a depression embedded in a smooth horizontal plain when a height-independent ambient flow was imposed on a two-layer atmosphere with a statically stable lower layer and a neutral layer aloft. Steady state flow solutions were presented in a regime diagram using three non-dimensional numbers that characterize the basin and the upstream flow parameters. The various flow regimes included sweeping and stagnating flow, lee waves, hydraulic jumps, internal wave breaking and upstream propagating waves.

 The present paper investigates the flow regimes observed in the METCRAX II observational network, extending previous observational and modeling investigations.

**3. Meteor Crater topography, observations, data conventions and synoptic weather**

*a. The Meteor Crater topography and ground cover*

The crater has been fully described in previous publications (e.g., Kring 2007; Whiteman et al. 2008, Lehner et al. 2016a), so that a succinct summary here will suffice. The near-circular crater with a diameter of 1200 m is 170 m deep and has a rim projecting 30-50 m above the extensive surrounding mesoscale plain. The plain descends from the Mogollon Rim, a line of high plateaus southwest of the crater, with a mean slope angle of 1° in the vicinity of the crater. The Meteor Crater and its surroundings are in a semiarid region with a sparse ground cover of shrubs and grasses. The main topographic features surrounding the crater are shown in **Fig. 1** where the crater is seen as a small pockmark on an extensive sloping plain.

*b. Measurement sites*

The meteorological measurements used in this paper and their locations are indicated in **Fig. 2**. Lehner et al. (2016a) provided a full listing of equipment, locations and observation frequencies, so we present a summary here. A 40-m tower at RIM had temperature, humidity and 3-D sonic wind measurements at 5-m height intervals. A 50-m tower at NEAR was equipped similarly, but with a 3-m height substituted for the 5-m height. Scanning Doppler lidars were located on the crater's north rim and floor. They were programmed to execute multiple scanning strategies; coplanar Range Height Indicator (RHI) scans along the 15-195° azimuth indicated in **Fig. 2** are featured in this paper. Full surface energy balance measurements, including four-component radiation measurements and ground, sensible and latent heat fluxes, were made at FLR, NEAR and FAR. Tethered balloon soundings were made at ~20-minute intervals from TS-B, TS-SW and TS-C to heights well above the crater rim, measuring temperature, humidity, and wind speed and direction. Battery changes sometimes resulted in larger time gaps between soundings. TS-C and TS-SW soundings were conducted synchronously. Additional sonic anemometer wind data came from the SSW4 and SSW2 masts on the upper and lower south-southwest sidewall. Lines of temperature data loggers (HOBO Pro, Onset Computer, Bourne MA) ran from the basin center up the sidewalls and out onto the adjacent plain to the northeast, southeast, south-southwest, southwest and west. These lines are designated NE, SE, SSW, SW, and W, respectively. The data logger on the rim on each line is designated by the line name with a prime (e.g., NE'). Automatic weather stations (AWSes) were located at various sites on the rim and the surrounding plain.

*c. Data conventions*

Several conventions will be employed in this paper. Times are Mountain Standard Time (MST) at the ends of any averaging intervals, unless otherwise stated. Astronomical sunset was at 1742 and sunrise was at 0635 MST. Elevations are in meters above mean sea level (m MSL) and heights are in meters above ground level (m AGL). Potential temperature (*θ*, °C) was calculated using *θ = T + γd (z - z0)*, where *T* is temperature (°C), *γd* = 0.0098°C m-1 is the dry adiabatic lapse rate, *z* is elevation, and *z0* = 1564 m is the crater floor elevation. Potential temperature is used instead of virtual potential temperature because of humidity measurement uncertainties in the variety of instruments used in the low humidity environment. In the remainder of this paper, without exception, the word *temperature* will be used for *potential temperature*.

*d. Weather conditions during IOP4*

IOP4 occurred during a period when a blocking ridge off the Pacific coast brought high pressure and clear, undisturbed weather to the region. The ground was dry, as the last rain (5.7 mm) had fallen on 10 October. The geopotential height gradient was weak above the region, supporting geostrophic winds from the northwest at 700 hPa. Three-hourly rawinsonde soundings at BASE upwind of the crater (**Fig. 3**) found a convective boundary layer (CBL) containing northeasterly upslope winds that extended to 2700 m MSL (1000 m AGL) at 1600 MST. The daytime incoming net radiation on the plain at the FAR and NEAR sites changed sign just before 1700 MST, followed soon after by a reversal of the surface sensible heat flux (**Fig. 4**). Sunset occurred at XYZ MST. A downward-directed surface sensible heat flux of about 15 W m-2 persisted throughout the night, producing a continuous cooling of the lower atmosphere and the upward growth of a surface-based stable boundary layer (SBL) containing downslope winds. It cooled continually and grew upward into the residual layer (RL) during the night to reach a strength of 12°C and a depth exceeding 100 m AGL after midnight. Downslope winds extended well above the shallow SBL into the RL where typical temperature gradients were 3°C km-1.

**4. Data analysis**

A sequence of transitional changes in temperature and wind structure was observed inside the crater basin during IOP4. These structural changes occurred in six phases and were driven by interactions between the crater and the boundary layer structure of the approaching flow over the adjacent plain. The analysis begins with an overview of the boundary layer evolution on the plain. The different structural phases inside the crater are then addressed individually and chronologically while relating them to the boundary layer structure of the approaching flow.

*a. Wind and temperature structure evolution above the plain*

Tethered balloon soundings at BASE, supplemented by winds from lidar velocity-azimuth display (VAD) scans above 1930 m MSL, provide an overview of the evolution of wind and temperature structure on the plain upwind of the crater (Fig. 5). In late afternoon an SBL began to form on the plain and grew upward into the afternoon RL, which contained upslope (northeasterly) winds. The SBL depth (Fig. 5a) grew at a rate of about 8.4 m h-1 to reach a depth of about 150 m AGL by 0600 MST. Above the SBL, the RL maintained a weak static stability and cooled slowly during the night. The development of the SBL over the inclined plain resulted in the formation of a southwesterly thermally-driven katabatic flow (Fig. 5b) that continued to strengthen and grow in depth to extend above the SBL, as is typical for katabatic flows (Zardi and Whiteman 2012). Wind directions veered with height from SSW at the surface to SW aloft. The downslope flow deepened and strengthened appreciably after 2230 MST when the thermally-driven katabatic flow merged with a southwesterly mesoscale flow [of unknown origin] that extended up to 2100 m ASL (400 m AGL) by 0100 MST. A northwesterly mesoscale flow developed above the plain after 0230 MST **(Fig. 5b)**, causing the downslope wind speeds to decrease at all elevations, and eventually confining the katabatic flow to elevations below 1760 m MSL (65 m AGL) after 0400 MST. Despite this confinement, the katabatic flow still extended above the crater rim. The increase and then decrease of downslope winds associated with the overlying mesoscale flows modulated the strength of the downslope flow in the RL, with a peak mean wind speed of ~5 m s-1 at around 0100 MST in the 80-120 m AGL layer. Similar mean speed variations were noted in the SBL on the NEAR and RIM towers and at the (varying) height of the wind speed maximum, with peak downslope speeds of about 6 m s-1 at 0130 MST **(Fig. 5c)**.

 Continuous data from sonic anemometers and temperature sensors at 5-m height intervals on the NEAR and RIM towers provided additional high accuracy, time-resolved data on the approaching boundary layer and its flow over the crater rim (**Fig. 6**), although the SBL and the height of the speed maxima eventually exceeded the tower heights (**Fig. 5**). On the plain at the NEAR tower, the 3-m level had already begun cooling by 1800 MST (**Fig. 6a**). The cooling progressed to higher tower levels as the SBL deepened, reaching the 50-m top of the tower at ~1915 MST. The 2-m AGL temperature 7.2 km south-southwest of the crater at FAR paralleled the temperature at 3-m AGL on the NEAR tower, highlighting the along-slope homogeneity of the temperature structure within the terrain-following SBL. Mean cooling rates at the lower tower levels were initially strong (~ 4°C/hour), but decayed to near zero between 2300 and 0230 MST. Substantial temperature oscillations occurred at the mid levels of the NEAR tower between 1900 and 2200 MST and at the upper tower levels between 2200 and 0230 MST.

 The daytime upslope winds reversed to down-slope at the lowest level of the NEAR tower at approximately 1830 MST (**Fig. 6b**). By 1930 MST, all levels of the tower had reversed to downslope. Downslope wind speeds then increased, with significant wind speed oscillations, as the height of the katabatic wind speed maximum climbed the tower, extending above the tower top by 2230 MST. Between 2230 and 0300 MST downslope winds at all levels on the tower remained fairly steady with a peak speed of 6 m s-1 at the top of the tower and with a strong positive wind shear in the lowest 3 tower levels (15 m).

 On the RIM tower, the temperature and downslope wind speed evolution was similar to that on the NEAR tower. The cold air from the growing SBL on the upwind plain first surmounted the rim at 1920 MST (**Fig. 6c**) and reached the 40-m top of the tower at 2000 MST. A sudden decrease in temperature at the top two levels of the tower occurred at about 0100 MST, significantly decreasing the stability in the tower layer. Temperatures were lower on the RIM tower than on the NEAR tower at the same elevation, as cold air from the plain SBL was lifted over the crater rim by the developing katabatic flow. The initial rapid rates of cooling and the leveling off after midnight were similar to the NEAR tower. Strong temperature oscillations at the lower tower levels before 2130 MST were similar to those at mid levels of the NEAR tower. These oscillations progressively moved to the higher levels of the RIM tower.

 Downslope winds (**Fig. 6d**) reached all levels of the RIM tower between 1930 and 2000 MST. They underwent a temporal progression similar to the NEAR tower, but with higher vertical wind shear at the lowest level, less shear at the higher levels and higher downslope wind speeds reaching 7 m s-1 at the top of the tower. Vertical wind speeds at the top of the RIM tower, initially positive, became negative at about 2000 MST and underwent a sinusoidal-like oscillation with peak negative speeds of about 0.6 - 0.8 m s-1 during the period from 2300 to 0230 MST and a return to near-zero speeds by 0430 MST (**Fig. 6e**).

 Temperature and downslope wind speed profiles on the RIM tower are shown at hourly intervals in **Figs. 6g and h** to supplement the time series portrayals in **Figs. 6a-f**. The temperature deficit on the 40 m tower reached 6-8°C by 2100 MST and maintained this deficit during the remainder of the night. The depth of the overflowing SBL exceeded the tower depth after 2300 MST. The tower temperature deficit reached 6-8°C by 2100 MST and remained at this level for the remainder of the night. Jet-like katabatic downslope wind speed maxima of about 3 m s-1 were present within the mid-levels of the tower during most of the night except at 0100 and 0200 MST when wind speeds increased throughout the tower depth.

*b. Wind and temperature structure evolution inside the crater*

The following sections chronologically step through the different phases of the evolving atmospheric structure inside the crater, focusing on hourly changes in vertical temperature structure on selected towers, tethersonde soundings and NE and SSW HOBO lines and hourly changes in wind on a two-dimensional vertical cross-section through the crater basin (**Fig. 7).** Wind data are composited from dual-Doppler wind retrievals (Cherukuru et al. 2015, Lehner et al. 2016a, W17,...) within and above the crater, tall tower profiles, short tower observations at SSW4 and SSW2 and tethersonde ascents.

 **Figure 8** provides sample temperature soundings to assist with the terminology and interpretation of **Fig. 7**. The *dividing streamline* (**Fig. 8a**) is the central streamline upwind of the crater and outside of its influence that, when lifted up the outer sidewall of the crater, just touches the rim (Leo et al. 2015). Flow approaching the crater below the dividing streamline elevation will split horizontally around the crater, while flow above the dividing streamline will be lifted over the rim. The dividing streamline elevation was estimated as the elevation on the NEAR tower where the temperature matched that observed at the base of the RIM tower.

 A continuous *cold-air intrusion* (*CAI*, **Fig. 8b**) occurs on the SSW inner sidewall when negatively buoyant air lifted over the SSW rim flows down the sidewall (Whiteman et al. 2010; Haiden et al. 2011). A diagnostic indicator of the CAI is a super-adiabatic temperature profile on the SSW HOBO line (W17). The HOBO lines extend over the rim and out onto the adjacent plain, so temperature profiles include information on the outer sidewalls.

 A *bifurcation* (**Fig. 8c**) occurs when a stable layer coming over the rim splits into a negatively buoyant current that runs down the inner sidewall and a non-negatively buoyant current that flows quasi-horizontally over the crater. In this situation the TS-SW sounding inside the basin will be warmer than the TS-C sounding, will have a capping stable layer at rim level that is missing the lower negatively buoyant portion that has run down the sidewall, and a layer of warm neutrally stratified air below the rim indicating the presence of a "cavity" that forms between the two currents.

 A *warm air intrusion* (**Fig. 8d**) is produced by a lee wave that brings RL air directly down into the basin above the southwest sidewall (Adler et al. 2012). No capping stable layer is present at or just above the rim in the TS-SW sounding in these instances.

1) OVERFLOW INITIATION PHASE, LATE AFTERNOON TO 1930 MST

Following the reversal of the local surface radiation and energy budgets (**Fig. 4**), separate stable boundary layers form inside the crater and on the adjacent plain (**Fig. 7a**). The cold air within the SBL flows down the topographic gradient as a katabatic flow, replacing the afternoon upslope flows on the plain (Savage et al. 2008) and basin sidewalls (Martinez-Villagrassa et al. 2013). The progressive cooling and deepening of the SBLs are accomplished by radiative (Hoch and Whiteman 2010; Mayer et al. 2010; Hoch et al. 2011) and sensible heat flux divergences (Zardi and Whiteman 2012). The surface radiation and energy budgets reverse earlier in the crater than on the plain (**Fig. 4**) because of the shadow that is cast into the basin from the south and west rims by the setting sun (Hoch and Whiteman 2010; Martinez-Villagrassa et al. 2013). Inside the basin, downslope flows converge on and over the valley floor, with buoyancy forces tending to level the isentropes (Zardi and Whiteman 2012). During this phase the SBL inside the crater develops a vertical temperature structure in which a shallow, strongly stable, relatively quiescent *cold pool* forms on the crater floor, surmounted by a near-isothermal layer that extends to the base of the RL at the crater rim (Whiteman et al. 2008).

 The SBL on the extensive tilted plain deepens slowly (**Fig. 5a**) in the late afternoon and evening. The cold air within the SBL begins to flow down the tilted plain producing an initial katabatic flow within the SBL that then extends some distance above it (**Fig. 5b**) into the nearly quiescent residual layer. The top of the SBL reaches the south-southwest rim at about 1930 MST, about a half-hour before the SBL on the plain farther upstream grows to this height (**Fig. 5a**), illustrating the role of the katabatic flow in lifting the SBL over the rim. The time of overflow initiation thus depends not only on the rim height and the rate of upward growth of the SBL on the plain, but also on the katabatic flow strength.

2) COLD AIR INTRUSION PHASE, 1930 TO 2130 MST

In the cold air intrusion (CAI) phase, a continuous stream of cold air is lifted over the rim. The flow turns down the inner sidewall, accelerating to speeds that depend on its negative buoyancy relative to the ambient environment inside the crater, the slope of the underlying sidewall, the drag of the underlying surface and detrainment at its upper boundary. The CAI accelerates until reaching and overshooting its level of buoyancy equilibrium where its momentum is dissipated through mixing and turbulence.

 The CAI phase begins at 1930 MST when the katabatic flow on the plain first lifts the plain SBL over the crater rim (**Fig. 6c**). Using potential temperature as an air motion tracer, the upstream origin of the inflow is above the surface of the plain at the dividing streamline elevation. This elevation on the NEAR tower is 1710 m MSL (15 m AGL) at 1930 MST but is typically 1715-1720 m MSL (20-25 m AGL) during most of the night (**Fig. 5a**). For comparison, the elevation difference between NEAR and RIM is 38 m AGL. As the SBL over the plain deepens and the katabatic flow strengthens, the SBL carried over the rim deepens (**Fig. 6c**). Sudden decreases in temperature indicate when the SBL reaches individual tower levels.The vertical temperature structure of the overflow closely resembles that above the dividing streamline on the plain when displaced upwards by the elevation difference between the dividing streamline and the rim (**Fig. 7a and b**), indicating that the lifting is adiabatic (W17).

 Several key features of the near-surface wind and temperature field during the CAI phase (e.g., at 2040 MST, **Fig. 9a**) can be seen in plan maps of the crater and its surroundings using data from HOBOs, automatic weather stations and the lowest tower levels. First, the cold, stably stratified air approaching the crater below the dividing streamline splits to the left and right around the crater, with a warmer well-mixed wake forming in the lee of the crater. Outflows are seen over the downwind rim, with temperatures in the wake similar to those inside the crater just below the downwind rim. The outflow and wake characteristics persist throughout the night, as exemplified by **Fig. 9b**.

 The adiabatic lifting of the plain SBL over the rim brings a continuous cold air inflow into the crater basin (**Fig. 6c**). Because the inflow is colder than air in the upper part of the basin, its negative buoyancy causes it to descend the inner basin sidewall, producing colder temperatures on the upwind inner sidewall than on the downwind inner sidewall and producing a super-adiabatic surface temperature profile along the upwind inner sidewall, as exemplified in **Fig. 7b**. This super-adiabatic temperature profile is thought to be produced by vertical wind shear that causes mixing within the flowing layer and progressive warming of surface temperatures with downslope distance. The adiabatic lifting of upstream air over the crater rim, the stably stratified flow and katabatic jet-like wind profiles above the rim, the colder temperatures on the upwind inner sidewall than on the downwind inner sidewall, the super-adiabatic along-slope temperature structure continue during the entire night (**Fig. 7**). The temperature deficit of the air coming over the upwind rim relative to the downwind rim (**Fig. 10**) was typically 3 ± 1°C during the CAI phase and afterwards.

 Temperature oscillations are present in the air coming over the rim. They are initially strongest near the surface but later occur at mid to upper elevations in the stable layer. The strongest temperature drops are usually accompanied by wind speed increases, for example around 2100 and shortly before 2200 MST (**Fig. 6c**). Some temperature oscillations, for example at 2100 and shortly after 2200 MST, can be identified at both NEAR and RIM (**Figs. 6a and c**). Oscillations at NEAR precede those at RIM by about 8 min, corresponding roughly to the advection time over the 1500 m distance between the two sites at wind speeds of 2-3 m s-1. The origin of these oscillations is unknown and they are not seen at FAR where only near-surface temperature observations are available and the advective time would be 35-55 min (**Fig. 6a**).

The cold air coming over the rim continuously descends the sidewall as seen in sonic anemometer data at SSW4 and SSW2 (**Fig. 6f**), with negative vertical velocities at the top of the RIM tower beginning at 2000 MST when the SBL reaches that level (**Fig. 6e**). These negative velocities continue throughout the night and indicate that the origin of the sinking motions is somewhat upwind of the rim. During the CAI phase, the downslope winds on the underlying sidewall are too close to the surface to be seen in the dual-lidar wind retrievals (**Figs. 7a and b**) but are visible in RHI scans from the FLR lidar (**Fig. 11a**).

 As the cold air flows over the rim it turns down the sidewall under the influence of gravity and accelerates from RIM to SSW4. The depth to which the CAI will penetrate into the basin depends on the ambient vertical temperature structure inside the basin, as the CAI will descend until reaching its layer-based level of buoyancy equilibrium (W17). The temperature structure inside the basin, as seen in both the TS-C and NE temperature profiles, includes a 40-m-deep cold air pool with a temperature deficit of 9°C surmounted by an isothermal layer extending to the rim. The near-equivalence of the TS-C and NE profiles (**Fig. 7a**) indicates that the isentropes in the undisturbed bulk of the basin atmosphere are quasi-horizontal. The CAI can more easily penetrate the near-isothermal layer than the stronger stability cold pool below. Model simulations (Haiden et al. 2011; Kiefer and Zhong 2011) suggest that the near-isothermal layer is an equilibrium structure produced by the continuous nocturnal cooling of the air coming over the rim, rising motions that compensate for the downslope CAI mass flux and detrainment at the top of the CAI as suggested by laboratory and numerical experiments (Baines 2005). If there is sufficient negative buoyancy, the CAI may reach SSW2, as evidenced by wind speed spikes at 1945, 2000, 2045 and 2115 MST that coincide with temperature minima at the RIM (**Fig. 6f and 6c**). SSW2, however, is often submerged below the top of the cold pool where the winds remain calm (**Fig. 6f**). Because increases in speed at SSW2 are associated with temperature decreases we can reject the hypothesis that the wind speed excursions are caused by seiches that submerse and re-immerse SSW2 in a sloshing cold pool, as speed increases would then be associated with temperature increases.

3) BIFURCATION PHASE, 2130- 2245 MST

The bifurcation phase was previously described for the 2300-0400 MST steady-state period of IOP7 (W17). In contrast, the bifurcation phase during IOP4 is a relatively short, temporary phase as the atmosphere does not reach a steady state and is driven toward an ultimate warm air intrusion (WAI) phase. Because the phenomenon has been previously reported, we only briefly summarize this phase, pointing out key structural features. Unfortunately, the RIM tower was not high enough to observe the full depth of the SBL after 2000 MST. But, since the lifting of the SBL over the rim was an adiabatic process, the structure above the rim can be determined by projecting the portion of the BASE temperature profile in **Fig. 7** that is above the dividing streamline upward by the vertical distance between the dividing streamline at BASE and the RIM tower.

 The bifurcation phase is initiated by the steady increase in depth of the SBL on the plain (**Fig. 5**). The bifurcation phase begins at 2130 MST when the overflowing SBL deepens to the point that it is composed of both a negatively buoyant lower portion that runs down the inner sidewall and a non-negatively buoyant upper portion that is advected quasi-horizontally over the basin. The separation of the two currents produces an isentrope bifurcation that begins just upwind of the rim. The upper isentrope is at the base of the stable layer carried quasi-horizontally over the basin, while the lower isentrope is at the top of the stably stratified CAI that descends the sidewall. As these two currents pull apart a low wind speed cavity or "stagnant isolating layer" (Winters and Armi 2014) develops between the two portions of the split isentrope. Because the cavity has the same isentrope at its base and top, it has neutral stability.

 During the bifurcation phase the elevation of the katabatic speed maximum on the plain rises above the rim elevation at 2150 MST (**Fig. 5b**) with a resulting increase in wind speed from ~2 m s-1 at all levels on the RIM tower to 3.8 m s-1 at the 3 m level and 5 m s-1 at the tower top. Negative vertical velocity at the top of the tower correspondingly increases from 0.2 to 0.5 m s-1 (**Fig. 6e**). The SBL is 30 m deep at 2200 MST (**Fig. 7c**) with a temperature deficit of 7.5°C (15°C at its top and 7.5°C at its base), but only the lowest 15 m of this overflow is negatively buoyant (i.e., colder than NE'). The negatively buoyant portion descends the upwind inner sidewall, leaving a neutral stability cavity in the atmosphere above TS-SW and a much weaker capping stable layer above TS-SW and TS-C than at RIM. Winds in the stably stratified atmosphere above the capping stable layer are carried across the crater (**Fig. 7c**). While dual-lidar retrievals cannot observe winds close to the sidewalls, the RHI scan from the FLR lidar detected the shallow CAI (**Fig. 11b**). The depth and strength of the CAI during the bifurcation phase is marginally greater than during the CAI phase, although the temperature difference between the upwind and downwind rim is similar (**Fig. 10**). This suggests that the negative buoyancy of the overflow is determined not by this temperature difference, but rather by the local temperature difference between the cold air intrusion and the warmer adjacent air in the cavity. The CAI flows down the sidewall until reaching and overshooting its neutral buoyancy elevation where a weak hydraulic jump forms, creating a zone of rising air above the lower sidewall (**Fig. 7c**). Weak speed oscillations occur on the upper and, especially, lower sidewall during the bifurcation phase (**Fig. 6f**), as they did during the CAI phase.

4) WARM AIR INTRUSION PHASE, 2245-0220 MST

The warm air intrusion phase (**Figs. 7d-g**) begins when a two-wave pattern forms over the basin with warm residual layer air brought directly down into the crater basin in the descending portion of the first wave. The first wave trough (crest) is located over the lower sidewall (valley center). The resulting wave pattern has a wavelength roughly equal to the radius of the circular basin and an amplitude that is typically twice the basin depth. The wave pattern develops in response to increasing downslope wind speeds in the residual layer above the plain SBL as seen in tethersonde soundings at BASE (compare **Figs. 7c and 7d**). The formation of the wave produced an abrupt transition within the basin, as observed in the dual-lidar wind field retrievals (**Fig. 12**). At 2242 MST (**Fig. 12a**) the bifurcation is still present, with the cavity separating the upper and lower currents and with a well-developed hydraulic jump on the lower slope. An unsteady lee wave, however,is present in the upper current, where wind speeds over the rim have accelerated to 2-3 m s-1. By 2247 MST this wave deepens and moves downstream over the sidewall, removing the cavity between the two currents, merging the CAI and WAI currents, and bringing warm RL air down into the basin (**Fig. 12b**). The developing wave brings the statically stable layer flowing over the crater in the upper bifurcation current down into the basin, adding to the CAI stable layer already present on the sidewall and causing an increase in the stable layer depth and strength on the sidewall (compare TS-SW temperature profiles in **Figs. 7c and d**). The downslope wind speed within and the depth of the underlying CAI increases during this phase (**Figs. 11c and d**). The hydraulic jump strengthens (**Figs. 7d-e**) as the CAI became more negatively buoyant relative to the adjacent WAI air and as momentum is transported downward from the WAI into the CAI.

 We hypothesize that the transition from the bifurcation phase to the WAI phase was caused by a growing wave instability in the upper bifurcation current produced by increasing vertical shear at the capping inversion, with weak winds in the cavity below and increasing winds in the stably stratified overflow. Following the abrupt 2245 MST transition, the WAI continued until 0220 MST when the maximum downslope winds at BASE and the mean downslope winds on the RIM and NEAR towers remained above 5 m s-1 (**Fig. 5c**). During this period, downward vertical velocity at the top of the RIM tower generally exceeded 0.6 m s-1 (**Fig. 6e**). Below the descending branch of the wave the 3 m AGL winds on the upper south-southwest sidewall at SSW4 maintained a speed of about 3 m s-1 during most of the WAI phase (**Fig. 6f**). The winds accelerated down the sidewall, producing gusty winds of up to 6 m s-1 at the SSW2 site. Interestingly, a sudden decrease in speed occurred at SSW4 when the flow over the rim exceeded 4.5 m s-1, possibly due to a bluff body flow separation in which the overflow is carried some distance into the crater before the cold air descends to the underlying sidewall and accelerates to SSW2. The strength and the position of the hydraulic jump on the sidewall were somewhat variable during the WAI phase, but the rising motions above the jump converge with the rising branch of the lee wave (**Fig. 7d-g**). Winds on the crater floor (3 m AGL at FLR) were nearly calm before the WAI, but intermittent speed increases occurred during the WAI, disturbing the cold pool on the basin floor (**Fig. 6f**).

 A first estimate of the depth of penetration of RL air into the basin can be obtained from TS-SW and TS-C temperature soundings, although the sounding locations are not always at the base or crest of the wave. The dual-Doppler wind field provides a second estimate, as there is a wind boundary between the nearly quiescent cold air that makes up the bulk of the crater atmosphere and the overflowing RL air (**Fig. 7**). Warmer temperatures below the rim at TS-SW than at TS-C (**Figs. 7d and e**) indicate a WAI in the descending branch of the wave. In the rising branch of the wave, the base of the RL separates the warm, high-speed RL air from the colder air in the bulk of the crater atmosphere and from cold air originating in the crater basin that is lifted well above the crater rim in response to the plunging air above the southwest sidewall (**Fig. 7e-g**). Both the descending and rising branches of the wave are confined above the southwest sidewall.

 Strong downslope winds during the WAI extend well down the sidewall, as seen in **Figs. 11c and d**. The presence of a capping stable layer at rim level above TS-C (**Figs. 7d-g**) indicate that the WAI is confined over the upwind inner sidewall and has not extended downwind to the crater center. The lee wave trough is often upwind of TS-SW (**Fig. 7e-g**). Ragged temperature profiles at TS-SW (**Figs. 7f and g**) indicate that the turbulent tethered balloon ascent path is on the border between the warm RL air and the colder air over the crater center. From 150 s-average dual-Doppler wind retrievals (not shown) the lee wave, while continually present, is quite unsteady, with short-term changes in amplitude and position and with intermittent flow separations occurring at different elevations in the descending branch of the lee wave. Vorticity maxima in the rising branch of the wave travel up the cold air boundary, indicating the presence of rotors or Kelvin-Helmholtz instabilities on this interface.

 Using a technique developed by Vogt (2010), a time-lapse sequence of geo-referenced thermal infrared images was used to visualize changes in the effective surface radiating temperature field inside the crater basin as the atmospheric structure sequenced through the different phases (cite Iris', Martina's AMS talks here). Two photos from the time-lapse sequence illustrate the differences between the CAI and WAI phases (**Fig. 13**). Both images include the shallow cold pool on the crater floor, the nearly isothermal temperature structure on the sidewalls above the cold pool and the colder temperatures on the upwind inner sidewall compared to the downwind inner sidewall. The cold pool is roughly circular during the CAI phase (**Fig. 13a**) but is progressively pushed to the northern side of the basin as winds strengthen in the WAI phase (**Fig. 13b**). The unsteady motions in the cold pool during the WAI phase split the cold pool apart and move it around on the crater floor. Turbulent mixing and sinking motions at the top of the cold pool produce warm temperatures at the floor (compare **Fig. 7c with 7d-g**), intermittent NNE-SSW-oriented warm streaks, and warm and cold patches on the crater floor. The flow strength during the WAI phase is, however, insufficient to sweep the cold air from the basin.

5) SECOND BIFURCATION PHASE, 0220-0410 MST

Downslope wind speeds in the RL upwind of the crater and on the NEAR and RIM towers begin to decrease midway through the WAI phase, presaging the development of a northwesterly mesoscale flow at elevations above 1760 m MST around 0400 MST (**Fig. 5b**). A rather abrupt decrease in the elevation of the downslope wind maximum occurs at 0145 MST. At 0220 MST the RIM tower wind speed drops below about 5 m s-1 and the WAI flow transitions back to a bifurcation, exhibiting the neutral cavity, upper and lower currents, the capping stable layer and the hydraulic jump present in the 2130-2245 MST bifurcation. In contrast to the evening bifurcation this bifurcation is intermittent, with short periods when the WAI phase reasserts itself. The intermittency of the WAI/bifurcation structures is exemplified by comparing **Figs. 7h and 7i**. Both times are within the bifurcation phase, but warm residual layer air is present over the sidewall at 0400 MST but not at 0304 MST. The cold pool on the crater floor begins to re-strengthen during this phase (compare **Figs. 7h and i** with **Fig. 7g**).

6) SECOND COLD AIR INTRUSION PHASE, 0410-0600 MST

The katabatic wind on the plain weakens further and becomes shallower as a mesoscale northwesterly wind (i.e., with an upslope component) develops above the Meteor Crater. By 0500 MST the bifurcation and hydraulic jump disappear and a final CAI phase ensues in which only a weak, shallow CAI comes over the southwest rim (**Fig. 7j**). During this time, the temperature deficit of the air coming over the rim is about 3°C compared to the downwind rim (**Fig. 10**). Wind speeds decrease on the lower sidewall and become calm on the crater floor (**Fig. 6f**). The cold pool re-strengthens. Temperature profiles at this time (**Fig. 7j**) become similar to those in the first CAI phase eight hours earlier (**Fig. 7b**).

**5. Discussion**

During IOP4 the atmospheric structure inside the crater basin underwent a complete cycle progressing sequentially through the overflow initiation, cold air intrusion, bifurcation, warm air intrusion, second bifurcation and second cold air intrusion phases. A two-dimensional conceptual model of the mean structure of the cold air intrusion, bifurcation and warm air intrusion phases on a vertical cross section through the crater basin is given in **Fig. 14** by combining analyses from **Section 4**. One must keep in mind when reviewing the figure that the flow is often non-stationary and that three-dimensional aspects of the atmospheric structure enhanced by the clockwise shifting of wind directions with elevation in the approaching flow are not addressed.

 While the present study benefitted from a comprehensive data set, there are data weaknesses that limit confidence in some aspects of the analyses. It is hoped that these weaknesses can be overcome with additional numerical simulations and field studies. Some scientific questions need further investigation. First, is detrainment occurring at the top of the cold air intrusion? This concept receives support from numerical and laboratory simulations, but observations are insufficient to verify this. Second, is the super-adiabatic, along-slope temperature profile produced by turbulent mixing within the flowing CAI? Third, what are the relative roles in enhancement of CAI winds during the WAI phase of downward momentum transport and enhanced negative buoyancy caused by local warming (Doswell and Markowski 2003; Peters 2016) adjacent to the slope? The speed increase during the WAI phase is not accompanied by a change in the upwind-downwind rim temperature difference (**Fig. 10**), so that this cannot be the appropriate negative buoyancy reference. Fourth, is the predominant two-wave pattern inside the crater forced by the terrain wavelength/amplitude? Fifth, to what extent is the WAI produced by the interaction of the approaching flow above the crater with the *effective topography* (Armi and Mayr 2015) of the crater?

 Following the overflow initiation phase the succession of atmospheric structures inside the crater basin appear to be an atmospheric analog to hydraulic flow features occurring in flumes and weirs as illustrated in many fluid mechanics textbooks (e.g., Massey 2006) or over seamounts in oceans or lakes (Armi and Farmer 2002). They bear similarities to other atmospheric "hydraulic" flows over mountains in which bifurcations, cavities and strong surface winds have been noted (Winters and Armi 2014). The 170 m topographic relief of the crater basin naturally produces weaker surface winds than are reported in downslope windstorms in higher relief locations. They may, nonetheless, provide an analogue to these larger scale flows. It is important to note that the strong winds at the foot of the inner sidewall in all structural phases of IOP4 occur when cold, negatively buoyant air descends the inner sidewall and that the strongest flows produce hydraulic jumps. The stronger winds in IOP4 relative to IOP7 (W17) occurred when a mesoscale flow developed above the plain that had a wind component in the same direction as the thermally driven downslope flow on the underlying plain. The continuously growing SBL depth on the plain was of major importance in initiating the cold air intrusion and bifurcation phases, but the increasing strength of the wind above the plain SBL then assumed increasing importance in initiating the WAI phase.

**6. Summary and conclusions**

A case study has focused on the interactions between an approaching southwesterly katabatic flow that formed on a 1° tilted mesoscale plain and a basin with a surrounding 30-50 m high rim that was formed on the plain by a meteorite impact. The interaction between the nighttime katabatic flow and the crater basin is the atmospheric equivalent of a "hydraulic flow" that evolves through six well-defined phases, three of which are illustrated in the conceptual model of **Fig. 14**. The *overflow initiation phase* includes the initiation and growth of stable boundary layers and katabatic flows over the plain and inside the basin. Above a dividing streamline height the katabatic flow lifts the plain SBL up the outer sidewall of the crater. In the c*old air intrusion phase* the plain SBL overflows the rim and, because it is negatively buoyant relative to air within the crater, descends the upwind inner sidewall detraining into the crater atmosphere at its top and flowing down the sidewall until reaching its level of neutral buoyancy and flowing out into the main crater atmosphere. As the plain SBL deepens and the katabatic flow increases in strength, the flow *bifurcation phase* begins in which the lower negatively buoyant portion of the stable layer continues to flow down the inner sidewall while an upper non-negatively-buoyant portion is carried over the crater in a quasi-horizontal overflow. A cavity forms between the two branches of the bifurcation containing near-neutral stability and low wind speeds. A shallow unsteady wave forms in the lee of the rim in the upper layer, and a hydraulic jump forms above the lower sidewall in the lower descending layer. As the SBL on the plain continues to deepen and the katabatic flow to strengthen, a *warm air intrusion phase* begins. The unsteady wave in the lee of the rim amplifies and brings the upper non-negatively buoyant stable layer down into the crater above the cold air intrusion, carrying warm air directly into the upper elevations of the basin from the residual layer above the plain. Rising air in the hydraulic jump over the lower sidewall merges with the rising branch of the lee wave. The disturbances produced by these phases are largely confined above the upwind inner sidewall, but the strong and turbulent flow near the base of the southwest sidewall during the flow bifurcation and warm air intrusion phases disturbs the pre-existing shallow cold pool on the crater floor. Decreases in wind speed above the rim later in the night cause the warm air intrusion phase to transition back to the flow bifurcation and then cold air intrusion phases before sunrise.

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**Figure 1**. Topographic map of the Meteor Crater and its surroundings. North is at the top of the figure. The crater is on the southwest side of the Little Colorado River drainage, which runs to the northwest. [**Fig01\_MC\_topo\_big.png**] SEBASTIAN SAYS THAT THERE MIGHT BE A BETTER FIGURE

**Figure 2**. Map of a) the crater’s location in the southwestern United States, b) the topography and measurement sites in the crater’s surroundings, and c) the topography and measurement sites in and near the crater, with north at the top of the figures. Sub-figures b) and c) are Universal Transverse Mercator zone 12S projections, with contour intervals at 20 m and 10 m, respectively. The dashed green line shows the location of the vertical dual-lidar cross-section. Dashed black lines in c) are roads.

[THIS FIG COMBINES TWO SEPARATE SUB-FIGURES. OMIT RWP AND NE in b) AND TEXT LISTING INSRUMENTS IN b). OMIT MICROBAROGRAPHS FROM c). c) COMES FROM CAI PAPER. EXTEND DASHED GREEN LINE PAST BASE. COULD USE COLOR TO LABEL THE SITES IN B TO MAKE THEM MORE VISIBLE ][**map\_all\_color\_v1.jpg, Fig01\_map\_all\_color\_v3.jpg**]

**Figure 3**. Three-hourly rawinsonde vertical profiles of a) temperature and b) horizontal wind vectors from BASE (1695 m MSL) in the period from 1600 MST 19 October to 0700 MST 20 October 2013. [Bianca, **20131020\_0000\_1400\_thetarsbaseffrsbaseddrsbase\_3000\_v3.png] [MOVE LEGEND. DAVE: ADD SBL/WIND SPEED BOUNDARY BY HAND] [**change colors to more intuitive evolution, thin out arrows or show direction profile instead. Omit figure even?]

**Figure 4**. Time series of 30-minute-mean net radiation *R* and turbulent sensible heat *H* fluxes at the FAR and NEAR sites on the plain and the FLR site on the crater floor. Local sunset times are 1742, 1742 and 1615 MST, respectively. [**energybalance.eps**]

**Figure 5**. Time-height cross-sections of a) tethered balloon temperatures and b) horizontal winds (arrows) above the upwind plain at BASE. In a), the elevations of the south-southwest rim, the dividing streamline at NEAR and the SBL top at NEAR are shown as dash-dot, dashed and solid lines, respectively. Color contours in b) are downslope wind components along a 35° azimuth. Winds below 1930 m MSL are from tethered balloon ascents; winds above 1930 m are from lidar VAD scans at 15-min intervals. Lines in b) indicate elevations of the south-southwest rim (dash-dot), the downslope speed maximum (solid black) and the 0 and 1 m s-1 downslope (35° azimuth) wind speed isotachs (red solid and red dashed lines, respectively). c) Maximum downslope wind component at BASE; mean downslope wind component in the flowing layer above the dividing streamline at NEAR and for a layer of equal depth at RIM; and mean downslope wind component in the 80-120 m AGL layer at BASE. Named evolutionary phases are bounded by vertical dashed lines (see text). [**upstream.eps**]

["theta" missing in color bar annotation

**Figure 6**. Temperature and downslope wind component time series at NEAR (a and b) and RIM (c and d). Measurement heights (3, 10, 15, 20…50 m at NEAR and 5, 10, 15… 40 m at RIM) increase from dark blue to red. The 2-m AGL temperature and the 3-m AGL downslope wind component at FAR (purple) have been added to a) and b), respectively. e) 40-m AGL vertical velocity at RIM. f) Wind speeds (NOT DS COMPONENT?) at SSW4, SSW2,FLR, and at 5 m AGL at RIM. Hourly g) temperature and h) downslope wind component profiles at RIM, as averaged over a 20-min period centered at the time indicated. Named evolutionary phases are bounded by vertical dashed lines. [3 HR TICKS, 1 HR SUB-TICKS. 0 M/S POINT ON EACH WIND PROFILE? OMIT BLK LINE IN f)] Manuela][**rim-near\_profiles/rim\_T-series&profiles.png**]

Either use "CA intrusion" or "cold air intrusion" but not both

**Figure 7.** a-j) Selected vertical cross-sections of dual-lidar, RIM tower, SSW4 and SSW2 two-dimensional (u, w) vector winds on a 35° azimuth vertical plane (location in **Fig. 2**) with color contours indicating wind speed. Sub-figures are labeled with the wind averaging interval and the phase. Tethersonde horizontal winds (no w-component) at BASE are for closest sounding within ± 10 minutes of the averaging interval. Off-plane wind data are projected perpendicularly onto the plane. The x-coordinate origin is the north rim lidar. A horizontal dot-dash line indicates the south-southwest rim elevation and vertical dashed lines show the locations of the TS-C and TS-SW sites. Solid lines indicate terrain and the base of the RL, with dashed lines delineating the top of the slope SBL. Zones of strong rising motion over the lower south-southwest sidewall are stippled. k-t) Mean vertical temperature profiles over the averaging interval for selected sites (see legend). The tethersonde temperature sounding closest in time to the averaging interval is plotted if within ± 10 minutes. The sloping dashed line depicts an isothermal temperature gradient. [**IOP4\_dual\_9subs\_p1.png and IOP4\_dual\_9subs\_p2.png and theta\_profs\_20min\_xxxx.png**] [LABEL z (m MSL). OMIT ALL TEMPERATURES FROM WIND FIGS. LEAVE VERT LINES TO INDICATE IN-BASIN TETHERSONDE LOCATIONS. COLO CONTOURS FOR WIND SPEED COMPONENT, ASDJUSTING LEGEND. SWITCH DIR OF WIND ARROW LEGEND. OMIT DIFFERENCE PROFILE. DASH-DOT 1733 m MSL IN ALL FIGS. PLOT ONLY SSW AND NE HOBO LINES. INCLUDE AVERAGING PERIOD (e.g., 2005-2010 MST) IN LABELS a) - j)] BUT NOT IN LABELS k) - t). DAVE: ADD LINES ON ALL SUB-FIGURES TO INDICATE BASE OF RL. DAVE: ADD STIPPLING TO INDICATE RISING MOTIONS ASSOCIATED WITH HYDRAULIC JUMP.]

**Figure 8**. Sample temperature profiles illustrating a) dividing streamline elevation, b) cold air intrusion along the upwind inner sidewall, c) bifurcation, and d) warm air intrusion over the upwind inner sidewall. Data come from 0400, 2120, 0400, and 2300 MST, respectively. [**theta\_profs\_conceptual.eps**] [OMIT PLURAL, LABEL TS-SW and TS-C]

**Figure 9**. Spatial pattern of near-surface horizontal winds (vectors) and temperatures (colored dots) at a) 2040 and b) 0040 MST. 5-min-mean data. [**ADD a) AND b). CONDENSE. OMIT LABELS ABOVE FIGURES**] [**20131020\_2040\_windsurface\_temperature.pdf, 20131020\_0040\_windsurface\_temperature.pdf**]

**Figure 10**. Temperature difference between the upwind (NE') and downwind rim (SSW'). [**Tdiff\_NE'-SSW'.eps**]

**Figure 11**. FLR lidar RHI radial velocity scans at a) 2010, b) 2147, c) 2305 and d) 0055 MST. Scans are over the upwind inner sidewall in the vertical plane indicated in **Fig. 2**. The dashed horizontal line at 1733 m MSL is the RIM tower elevation. Blue and red vertical profiles are from the SSW and NE HOBO lines, respectively. Colored dots on the topographical cross-section indicate equivalent radial velocities at 2 m AGL at SSW4 and SSW2. [z (m MSL), CONDENSE/CROP SUB-FIGS, MOVE m s-1 LEGEND TXT RT, LABEL a), …, OMIT SSW. d) SATURATED? CHANGE HORIZ LINE TO DASH-DOT] [**2010\_RHI.png, 2147\_RHI.png, 2305\_RHI.png, 0055\_RHI.png**]

**Figure 12**. Illustration of the abrupt change in flow structure between the a) bifurcation phase and b) warm air intrusion phase. Vectors indicate the winds in the two-dimensional vertical cross section through the crater. Colors indicate vertical velocity. The times indicated are the ending times of 2.5-min means. Red lines are tethersonde temperature soundings at the vertical dashed line locations [FLIP, AND REMOVE EXTRANEOUS THINGS BUT LEAVE THE TWO SOUNDINGS, ADD w (m s-1) TO THE LEGEND. CHANGE HORIZ LINES TO DASH-DOT]

**Figure 13**. Geo-referenced thermal infrared temperature overlay on the Meteor Crater topography at a) 2100 and b) 0000 MST. The effective surface radiating temperature data comes from compositing data from three cameras, two operated side by side on the north rim looking south-southwest and one on the south rim looking north-northeast. Horizontal wind vectors are shown in black at RIM, SSW4 and SSW2 and in white at FLR. [Iris] [ADD WIND VECTORS FROM SSW2, SSW4, FLR. OMIT U BASEL TEXT] [**TIR\_2100.png, TIR\_0000.png**]

**Figure 14**. Two-dimensional conceptual diagram (not to scale) of the a) cold air intrusion, b) bifurcation, and c) warm air intrusion phases. Profiling sites are at the labeled red dots in b). Vectors are 2-D winds; lines above the terrain are selected isentropes. zSBL and zds are the SBL and dividing streamline heights on the plain far upwind of the crater. *θNE'*and *θSSW'*are NE and SSW rim temperatures. LNBE is the layer-based neutral buoyancy elevation. [**concept\_3\_phases.png**]