THE DYNAMICAL MECHANISMS OF COLD-AIR POOL REMOVAL

by

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1. INTRODUCTION

The primary objective of the research described in this proposal is to contribute to our scientific understanding of the dynamical processes affecting the removal of persistent cold-air pools (CAP) from valleys and basins. This research is motivated by a need for improved understanding of CAP break-up, a topic which strongly impacts air-quality forecasts for urban and agricultural basins.

As described in Lareau et al. (2013), cold-air pools form in valleys and basins (spanning a range of scales) due to cooling of the air near the surface, warming of the air aloft, or both. The resulting stable stratification prevents the air within the basin from mixing vertically while the confining topography prevents lateral displacement and favors air stagnation. CAPs may be either *diurnal*, forming at night and decaying the following day, or *persistent*, lasting through multiple diurnal cycles.

Persistent CAPs, which are the focus of this paper, are most common in the winter, often forming during the onset of ridging and lasting through a variety of weak disturbances (Whiteman et al. 2001). When they occur within densely populated basins, the emissions from vehicles, home heating, and industrial sources accumulate and sometimes lead to hazardous air quality. Suppressed temperatures and high humidity within the CAP may also contribute to the formation of dense fog capable of adversely impacting air and ground transportation.

The longevity and strength of persistent CAPs is modulated by interactions between the CAP and passing weather systems. Weak disturbances may temporarily perturb the CAP leading to partial “mix-out”, whereas more vigorous systems completely flush a valley. The degree of these synoptic-scale/CAP interactions, and in particular the timing of their break-up, is often difficult to predict. For example, numerical models may prematurely remove a CAP during a modest wind event, whereas in reality the CAP persists for many more days leading to continually deteriorating air-quality.

The research outlined in this proposal is intended improve our ability to forecast the break up of persistent CAPs by augmenting our understanding of the processes contributing to their removal. We explore, via observations and modeling, the impact of four mechanisms that are hypothesized to contribute to CAP destruction:

(1) The interaction of CAPs with mountain waves
(2) Turbulent CAP erosion
(3) Pressure and wind-stress displacement of CAPs
(4) Density current dynamics within CAPs

The remainder of this document is structured as follows: (1) A synthesis of previous investigations into the life-cycle and break-up of persistent CAPs, (2) a description of observations and data, (3) progress to date on observational analyses, and (4) an outline for subsequent numerical simulations and an overview of research objectives and (5) a timeline.

2. LITERATURE SYNTHESIS

Persistent cold-air pool lifecycle

The life-cycle of persistent CAPs is addressed by a handful of investigations conducted in the western United States, where climatological mean ridging couples with numerous topographic corrugations to favor frequent wintertime persistent CAPs. These

CAPs are deeper than typical nocturnal inversions (Wolyn and Mckee 1989) and vary in duration from two days to more than two weeks.

Synoptic-scale processes are the primary control on the strength and duration of persistent CAPs (Whiteman et al. 1990; Whiteman et al. 2001; Zhong et al 2001; Reeves and Stensrud 2009; Lareau et al. 2013). In particular, persistent CAPs preferentially form during periods of mid-tropospheric warm air advection and subsidence associated with the onset of ridging (Wolyn and Mckee 1989; Whiteman et al 1990; Reeves and Stensrud 2009; Lareau et al 2013). These same conditions favor clear skies and strong nocturnal surface inversions, which couple with warming aloft to produce deep or multi-layered stable profiles (Whiteman et al. 2001, Lareau et al. 2013). During longer-lived events, minor synoptic-scale variations aloft may temporarily strengthen or weaken the CAP (Wolyn and Mckee 1989; Whiteman et al 1999; Reeves and Stensrud 2009; Lareau et al. 2013).

The growth of the convective boundary layer (CBL) within persistent CAPs is generally insufficient to destroy their ambient stratification (Wolyn and Mckee (1989); Vhrovec and Hrabar 1996; Zhong et al. 2001). The presence of low clouds, fog, or snow cover may further diminish CBL growth and prolong the duration of CAPs (Wolyn and Mckee 1989; Zhong et al. 2001). The occurrence of boundary layer clouds also impacts the thermodynamic structure of CAPs, favoring mixing below clouds base and sharp elevated inversions at cloud top (Wolyn and Mckee 1989; Lareau et al. 2013).

Generally speaking, the break-up of persistent CAPs is associated with an approaching baroclinic trough. For example, the most common mode of CAP destruction is a reduction of stability via cold-air advection aloft following the passage of a trough

(Wolyn and Mckee 1989; Whiteman et al 2000; Reeves and Stensrud 2009, Lareau et al. 2013). However, in some cases warm winds preceding a trough are also hypothesized to displace or remove a CAP (Whiteman et al 2001; Flamant et al. 2006). In other instances, advection combines with dynamical processes, such as turbulent erosion and CAP tilting, as a potential path for CAP destruction. In all cases, the degree of interaction between the approaching trough and the CAP can be difficult to predict, and considerable work remains to forecasts for CAP removal processes.

The following subsections provide a more detailed summary of previous investigations specifically addressing the dynamical mechanisms for CAP removal.

Turbulent Erosion

The topic of turbulent CAP erosion has received considerable, though incomplete, consideration in the literature. Petkovšek (1992) uses a semi-analytic model for top-down turbulent erosion of “cold air lakes” in the Slovenian mountain valleys. He proposes that turbulence in the layer above the CAP penetrates downward into the stable stratification. The penetrative distance is limited by the consumption of kinetic energy in work against buoyancy. Assuming that the air within the penetrative region becomes mixed into the layer aloft, Petkovšek shows that the top of the CAP will lower but also strengthen over time. The strengthened CAP subsequently suppresses the turbulent encroachment, thus requiring an accelerating wind for turbulent erosion to continue.

A similar semi-analytic approach is employed by Zhong et al. (2003) to investigate the time-scale for turbulent erosion. Using steady winds and an idealized CAP profile, their results show an exponential decay of the erosion rate with time, consistent with Petkovšek's hypothesis. They conclude, however, that for CAPs of typical strength and winds of usual magnitude, turbulent erosion is a very slow process and unlikely to cause the break-up of CAPs independent of other processes.

The hypothesis of turbulent CAP erosion has also been tested in full numerical simulations. Vrhovec and Hrabar (1996), for example, find that turbulent erosion occurs during accelerating winds but not with modest (5 m s^{-1}) steady winds. Similar findings, requiring strong accelerating winds, are found by Rakovec et al. (2002) and supported by observations for two CAP erosion events during accelerating winds. Conversely, Zhong et al. (2001), examining the processes affecting CAPs in the Columbia Basin, find that turbulent erosion is not a significant mechanism for dissipation, echoing the previous results of Lee et al. (1989) where turbulent erosion was negligible during mountain wave/CAP interaction.

Some caution is required in interpreting these seemingly conflicting results. None of these investigations employ sufficient model resolution to capture the inertial scale eddies and dynamic instabilities (e.g. Kelvin-Helmoltz instability) responsible for mixing along sheared density interfaces (Strang and Fernando 2000). These simulations instead rely on parameterizations for sub-grid scale mixing, which are known to perform poorly in strong stability (Fernando and Weil 2010). Moreover, laboratory experiments show that the length scales associated with buoyancy and velocity discontinuities, such those across the top of a CAP, are different and may not be co-centered (Strang and Fernando 2000). This consideration is not included in the analytic models of Petkovsek (1991) and Zhong et al. (2003), yet has important implications for dynamic instability and turbulence. We will return to this point in addressing the set up for our numerical modeling investigations (Section 5).

CAP Tilting and Displacement

Whether or not strong winds will ventilate a valley depends not only on turbulent erosion but also on the displacement and tilting of a CAP. Lee et al. (1989) run a series of numerical simulations, for example, to determine to what extent mountain waves displace a lee-side CAP. Their results indicate that the mountain wave is unable to penetrate to the lee slopes when an adverse pressure gradient keeps the cold-air banked against the mountains. In the absence of such a gradient, strong winds and warmer temperatures reach the surface and displace the CAP. A similar displacement process is invoked to describe the development of a warm front in the Columbia Basin (Whiteman et al. 2000) and likewise in the Rhine Valley during MAP IOP-15 (Flamant et al. 2006).

In addition to the dynamic impact of mountain waves on CAPs, Petkovšek and Vrhovec (1994) and Zangl (2003) demonstrate that CAPs develop inclined surfaces due to regional pressure gradients and strong winds. Using Margule's formula, Petkovšek and Vrhovec (1994) show that the angle of inclination for the CAP is given by

$$\tan \alpha = \frac{|\nabla P| - |\nabla P'|}{g(\rho - \rho')} \tag{2.1}$$

where α is the angle of inclination, $\nabla P'$ is the pressure gradient aloft, ∇P is the pressure gradient at the surface (assumed to be near zero), g is gravity, and $\rho - \rho'$ difference in density across the CAP interface. This tilting is one mechanism by which the areal extent of a CAP in a basin may be diminished. Zangl (2003) extends this approach, refining the

Margule's formula for a discontinuity in the temperature *gradient,* and shows that a more accurate CAP structure is given by

$$\Delta z = T\sqrt{\frac{dP}{dx}\frac{2R}{Pg\Delta\gamma}}\sqrt{\Delta x} \qquad (2.2)$$

where the depth of the cold-air (Δz) is proportional to the square root of the distance from the leading edge of the cold-air (Δx). P is the pressure, R the gas constant for dry air, g gravity, and $\Delta\gamma = \gamma_c - \gamma_w$ which is the difference in the lapse rate between the CAP air and the environment. This hypothesis is tested using a series of controlled numerical experiments wherein the CAP break-up occurs when the tilt is sufficient to spill cold air over the confining topography.

A similar tilting phenomena, known as "wind set-up" or storm surge, is observed when winds blow along the fetch of some lakes, such as Lake Erie (Schwab 1978). An approximate force balance is accomplished between the hydrostatic pressure gradient within the lake, caused by increased water depth in the downwind direction, and the surface wind stress integrated along the fetch. While CAPs do not have a free surface, a similar mechanism likely occurs within CAPs during periods of strong winds aloft. Zangl (2003), for example, show that the CAP tilt exceeds that attributable to the pressure gradient when ageostrophic winds impart momentum to the CAP.

Literature Summary

Previous investigations of persistent CAP lifecycle help explain the processes that affect their break-up. Synoptically controlled cold air advection aloft is one of the key mechanisms for CAP removal, whereas the growth of the CBL is an unlikely source for destruction unless coupled with other factors. The turbulent erosion of CAPs remains a contentious topic. Some investigations suggest that it is a major mechanism for breakup while others find it insignificant. The displacement and tilting of CAPs is also hypothesized to impact CAP breakup, though these theories have yet to be tested in field experiments.

Despite the diligent work of previous investigators, CAP removal remains a poorly understood and difficult to forecast phenomena. In this research proposal we hope to advance our understanding of the importance of dynamical mechanisms for CAP removal by testing existing theories against observations and expanding on theory using high resolution numerical simulations.

3. DATA

The observations used in this study were collected as part of the Persistent Cold-Air Pool Study (PCAPS), which was conducted in Utah's Salt Lake valley from 1 December 2010 through 7 February 2011. A detailed review of that project, its instrumentation, and preliminary finds can be found in Lareau et al. (2013). The topographic setting and instrument locations for PCAPS are shown in Fig. 1.

Not addressed in Lareau et al. (2013) is the development of a post-processed 30-minute resolution time-height data set for potential temperature, relative humidity, and vector wind during PCAPS period. This data set blends together radiosonde, surface, and remote sensors observations to establish a continuous "best guess" state of the valley atmosphere. We rely heavily on this time-height data and surface observations for the case studies present in section 4, as well as for model set up and verification in section 5.

4. PROGRESS TO DATE

PCAPS IOP-1: Partial CAP Mix-Out Analysis

On a number of days during PCAPS strong southerly winds interacted with persistent CAPs producing waves, displacements, and frontal structures. The most remarkable of these interactions occurred during PCAPS IOP-1 from 00-18 UTC of 3 December 2010 when a strong CAP was substantially displaced, forming a front that advanced back and forth through the Salt Lake Valley.

a. Event Overview

A persistent CAP and poor air quality impacted the Salt Lake valley from 30 November – 6 December 2010 (Fig. 2). The CAP forms on 30 November during rapid warming of the mid- and lower-troposphere following the departure of a cold trough, and is augmented by surface cooling from the freshly snow covered ground. Continued warm advection and subsidence drive the capping inversion progressively lower into the valley atmosphere through 2 December, forming a 700-m deep CAP with a potential temperature deficit in excess of 15 K.

On 3 December a weak short wave trough provides a period of disturbed CAP conditions and a partial mix-out of the valley, which is the topic of this case study. Displacement of the CAP is first evident around 00 UTC when a low amplitude short wave trough (dashed line in Fig. 3) is draped across the west coast of the United States placing Utah in region of enhanced pressure gradient (Fig 3a). A weak lee trough resides over portions of Wyoming and Colorado. The CAP/trough interaction becomes more pronounced at 06 UTC while Utah is just downstream of progressive northern portion of the trough, which has fractured from its southern extremity (Fig. 3b).

The regional pressure gradient over Utah begins to diminish at 12 UTC as an upstream short wave ridge amplifies and the original trough weakens in favor of a deepening lee cyclone (Fig. 3c). Finally, at 18 UTC, the amplifying short wave ridge becomes the dominant feature in the intermountain west, and the pressure gradient over northern Utah slackens, favoring a return to quiescent CAP conditions (Fig. 3d).

The CAP subsequently persists through about 00 UTC on 7 December at which point convection associated with an approaching trough ventilates the valley.

c. Surface Analysis

During the night (00-12 UTC) of 3 December meteorological observations show a series of sharp jumps and drops in temperature, humidity, and wind speed at surface observing sites throughout the Salt Lake valley (Fig. 4). To explain the temporal and spatial variations of these temperature fluctuations we use an objective analysis, combining data from numerous observing sites to construct hourly temperature grids. Figure 5 shows these analyses along with the position of a front that formed between the CAP and the warm air to the south. This front is subsequently referred to as a warm (cold) front when the warm (cold) air is clearly advancing in areal extent.

At 01 UTC cold air (~0° C) and light northerly winds (2.5 m s^{-1}) occupy most of the Salt Lake valley (Fig 5a). Warmer air (5-10° C) and stronger southerly winds (2.5-7 m s^{-1}) are found southwest of a frontal boundary and along the crest of the Traverse Ridge. Strong southerly gap flow (11 m s^{-1}) and locally colder air (~2-4° C) emanate from the Jordan Narrows, where air is flowing out of the CAP in the neighboring Utah valley.

The warm front first moves to the north at 03 UTC, displacing the CAP along the valley axis and bringing warm, windy conditions to southern portions of the valley (Fig. 5b). Winds are particularly strong along the Traverse Ridge and cold gap flow continues in the Jordan Narrows.

The warm front continues its northward advance through 06 UTC (Fig. 5c). Warm temperatures (~10° C) and strong southerly winds (7-15 m s^{-1}) are reported throughout the southern half of the valley, while light winds and temperatures around 0° C remain within the CAP to the north. The across front gradient in temperature is as much as 10° C over 2 km, which is supported by observations of rapid temperature jumps of comparable magnitude (not shown). The warmest air in the valley (~12° C) is found near the base of the Traverse Ridge, where winds are gusting upwards of 13 m s^{-1}.

The front abruptly reverses course between 07 and 08 UTC, returning south through the valley as a sharp cold front (Fig. 5d). Winds to the north of the front (i.e. within the CAP) are coherent in direction and strength, flowing from the NW to SE then turning south along the valley axis. Strong southerly winds continue on the warm side of the front, indicating strong convergence along the leading edge of the cold air. At many locations the cold frontal temperature drop is nearly identical in magnitude to the previous warm frontal rise (e.g. variations in Fig. 4). By 09 UTC the cold front encroaches on the Traverse Ridge, restoring CAP conditions to most of the valley (Fig. 5e). However, warm temperatures and strong winds continue along the southwestern portion of the valley and the crest of the Traverse Ridge at this time.

The front reverses direction once again, moving north as a warm front between 10 and 12 UTC, re-establishing a position across the valley center (Fig. 5f,g). As before, convergent winds accompany a compact temperature gradient in the frontal zone. Not apparent in this hourly analysis are smaller fluctuations in the frontal position, which appear in station observations during this interval (not shown).

As the regional pressure gradient diminishes after 12 UTC winds aloft rapidly decrease, and the front again mobilizes southward as a cold front at 13 UTC. By 14 UTC, the CAP has returned throughout the valley (Fig. 5h) and within the next hour penetrates south through the Jordan Narrows into the Utah valley (not shown). The gap flow reversal effectively marks the end of the disturbed CAP conditions and a return to the more typical quiescent scenario.

It is worth emphasizing that the partial mix-out occurs at night during a period of near zero net radiation (not shown). As such, it is reasonable to interpret the abrupt temperature changes as principally dynamic, not diabatic, in origin.

d. Vertical Evolution

To gain further insight into the evolution of the CAP on the night of 3 December we use data from radiosondes and vertically profiling remote sensors situated at the NCAR ISS sites in the valley center (see Fig. 1). The front, described in the above analysis, passed over this site once as a warm front at 05:20 UTC and once as a cold front around 07:20 UTC, revealing significant changes in the CAP vertical structure.

Figure 6a presents a time-height diagram of potential temperature and wind speed above the ISS site along with the variations of surface wind and temperature (Fig. 6b,d).

Also shown is the aerosol backscatter from the laser ceilometer, which helps to visualize changes in the CAP associated with the frontal passages (Fig. 6c). The period of warm frontal passage is evident near 05:20 UTC as the tightly spaced and vertically oriented isentropes over the lowest 100 m. Prior to the warm front, the CAP top (approximately represented by the 295 K isentrope, Fig 6a) lowers from ~1650 m to ~1450 m between 00 and 06 UTC. The stratification within the CAP is roughly linear, while that aloft is closer to adiabatic. The lowering of the CAP occurs in concert with strong warm air advection aloft. For example, between 00 and 06 UTC the air temperature at 1700 m MSL increases by about 5 K while south winds in the same layer increase from 10 to 15 m s^{-1}. Winds within the CAP are light (0-2 m s^{-1}), indicating strong wind shear across the stable layer.

At the time of the warm frontal passage, the base of the shear layer penetrates to the surface, providing a burst of strong south winds (~7 m s^{-1}) and a jump in temperature of ~6 K. Interestingly, while the lowest 100 m above the ISS site become well mixed following the frontal passage, a capping inversion remains intact above 1450 m, which prevents a full coupling of the surface conditions to the adiabatic layer aloft.

Warm windy conditions remain at ISS until the leading edge of the CAP passes over the site as a cold front at ~07:20 UTC. The arrival of the cold front disturbs the ambient stratification, displacing isentropic surfaces upwards. Winds above the CAP remain strong until 12 UTC, after which they rapidly diminish during a period of weak cold advection. As the air aloft cools the CAP weakens but also increases in depth.

e. The structure of the front

The fine scale structure of the front (i.e. the leading edge of the CAP) is recorded by the laser ceilometer as it passes over the ISS site, revealing strikingly different morphology of the warm (Fig. 7) and cold fronts (Fig. 8). The warm front exhibits a slow descent of the aerosol layers towards the surface, whereas the cold front arrives with a distinct "head" wherein the aerosol concentration and depth both increase rapidly in time. These differences are best understood in the context of density (or gravity) current dynamics.

The passage of the warm front (Fig. 7) is similar to the structure of an "arrested saline wedge," which is a form of density current found at the interface between ocean salt water and riverine fresh water (Simpson 1997). These wedges are generally close to being at rest, lack a distinct "head," and have lesser turbulence along the density interface (Simpson 1996). Their shape is determined, in part, by the shear profile of the opposing flow and the weak surface drag acting within the slowly moving dense fluid (Simpson 1996). It is notable then that the warm frontal passage at ISS does not exhibit an elevated "head" but instead a gradually sloping interface in time. Furthermore, there is a slight suppression of larger amplitude (~100 m) waves along the warm frontal surface starting at 04:10 UTC, concurrent with the onset of the layer's descent. Smaller waves (~10-20 m) are, however, still present. We address these waves in more detail in the following subsection.

In contrast, the cold front (Fig. 8) closely resembles idealized simulations of a rapidly advancing (i.e. super critical) density current moving into a two-layered stratified system (Simpson 1997; White and Helfrich 2012). In this instance, the two-layered environment is composed of a surface based mixed layer separated from an elevated mixed layer by a sharp jump in potential temperature (see Fig. 6a). The density current itself is composed of the much colder air within the CAP and possesses a distinct "head" that undercuts and lifts the upstream environment. There is no upstream bore in this case because the flow relative to the density current is likely strong enough to prevent the upstream propagation of gravity waves (White and Helfrich 2012). Behind the density current head, Kelvin-Helmholtz waves (KHW) form a region of mixing between the pre- and post-frontal aerosol concentrations. This fits with density current theory, wherein breaking KHW lead to a mixing region behind the leading "head" (Simpson 1997).

Further analysis is required to better understand the interaction between the advancing and retreating density currents and the environmental shear conditions, which may also be related to the mountain wave dynamics described below.

f. Kelvin-Helmholtz Instability

Many waves, including KHW, were observed within the CAP during the partial mix-out. For example, Fig. 9 shows a one-hour interval of laser ceilometer backscatter data (starting at 02 UTC) in which a sequence of KHW are apparent, particularly at ~02:50 UTC. These KHW reside in a layer of weak stability within the deeper CAP structure (not shown). Within the crests of the largest billows (~100 m amplitude) high aerosol concentrations (shown as laser backscatter intensity) are extruded into less polluted layer above, while cleaner air is folded towards the surface. A mixed region, with intermediary values of backscatter intensity, is found within the overturning region. The frequency of the largest waves is about once every 2.5 minutes and numerous smaller waves occur at intervals of ~ 30 s. Using an approximate wind speed of 10 m s^{-1} for the flow aloft, and invoking Taylor's hypothesis for frozen turbulence, we estimate the horizontal scale of the KH waves to be ~1500 m. The smaller waves, by contrast, are estimated to have a scale of ~300m. These rough estimates are useful in determining the model resolution required to resolve the inertial scale eddies within this disturbed CAP scenario.

KHW and the breakdown to turbulence may also contribute to the sudden jump in temperature and wind associated with the passage of the surface warm front at ISS (as shown in Fig. 6). Figure 10 shows data from a sounding launched from ISS just 5 minutes prior to frontal passage. Despite the presence of a sharp surface inversion (Fig 10a), there is a ~100 m layer above the surface through which the gradient Richardson number is less than critical (Fig. 10d). Assuming that turbulence in this layer forces the column towards a well-mixed profile (red line, Fig. 10a), the surface temperature should rise from 287 to 293 K. This simple prediction is verified moments later as the surface temperature jumps ~6 K, strong winds mix to the surface (as in Fig. 6b), and aerosol becomes well mixed up to 1450 m (as in Fig. 6c).

It is worth nothing that the KHW and turbulence seen during IOP-1 only partially fit with the theory of top down turbulent erosion. One major complication is the presence of multiple stable layers with varying wind shear profiles. In other words, the wave breaking and mixing is occurring *within*, not at the top of, the CAP. Considerable work remains to understand the role of KHW in the partial destruction of the 3 December CAP. For example, a more complete treatment of the amplitude and time scales of these waves is possible using wavelet or harmonic analysis, and large-eddy simulations will help quantify the contribution of the breaking waves to CAP destruction.

g. Mountain Waves

During the displacement of the CAP on 3 December there is evidence of a mountain wave over the Traverse Ridge. Figure 11 shows a sequence of pseudo-soundings developed from potential temperature observations taken from 6 sites ascending South Mountain (SM), which is part of the Traverse Ridge (see Fig. 1). These pseudo-soundings are compared with the potential temperature profile at the valley center (ISS) for three sequential times. At 02:20 UTC the profiles are similar, both showing a strong inversion extending from the surface to ~1600 m (Fig. 11a). Just 10 minutes later, however, the SM transect experiences rapid warming, developing a nearly adiabatic profile from the ridge crest down to ~1410 m (Fig. 11b). Below 1410 m a sharp (~10 K) inversion is found. The ISS sounding is nearly unchanged at this time. During the subsequent 1.5 hours the adiabatic layer at SM descends to the surface, bringing warm temperatures (~12 C) to the mountain base (Fig. 11c). The difference between the two soundings is striking, with a 15 K inversion remaining in place in the valley center.

When the SM temperature profile is close to adiabatic the winds at the base of the ridge are generally equal to or greater than then those at the ridge crest, supporting the hypothesis of hydraulic flow. These periods correspond closely with the timing of the warm frontal advances, suggesting that the mountain wave impacts the development and motion of the front (Fig. 12). Likewise, lulls in the winds at mountain base correspond to cold frontal advance, indicating that the decay of the mountain wave may allow the CAP to slosh to the south as an advancing density current. Exactly what role the mountain wave plays in the formation of the surface front and the timing of the frontal advances will be the continued focus of our analysis and modeling.

h. CAP Tilting

The enhanced regional pressure gradients during the partial mix-out contribute to CAP tilting and displacement. Figure 13 shows three profiles of potential temperature along a north-to-south transect of the Salt Lake valley. The data reveal a strongly sloping CAP, which grows in depth towards the north, the down gradient direction. The slope of the CAP, determined by the distance between profiles, is approximately 400 m over 25 km, and varies in a manner roughly akin to the square root of the distance from the leading edge of the cold air (not shown). These observations are consistent with the simulations of Zangl (2003) but may also include a tilting component due to the Reynolds shear stress imparting momentum into the CAP. The KHW, discussed above, are a manifestation of this momentum transfer, and further work may help to quantify their role in developing the CAP tilt.

i. Event Summary

Mountain waves, dynamic instability, turbulence, density currents, and regional scale pressure affects all contribute to the evolution of the CAP displacement on 3 December. These phenomena are not necessarily independent. For example, KHW drive mixing proximal to the cold front but also affect CAP tilting over the length of the valley. Likewise, the motion of the density current may relate to the strength of the mountain wave forming over the Traverse Ridge. We have also shown that the classical theory of top down turbulent erosion may not be applicable to the present case where most of the wave breaking occurs within, not at the top of, the CAP. Data collected during two other disturbed CAP episodes, PCAPS IOPs 4 and 9, are now being analyzed and may help us better understand the relative importance of these phenomena from one case to another.

5. RESEARCH TRAJECTORY

Observations alone are insufficient to substantively advance our knowledge of persistent CAP destruction by dynamical processes. To this end, the remainder of this research proposal focuses on numerical experiments devised to elucidate these processes.

Idealized Modeling

In order to better understand the effects of mountain waves, turbulence, and dynamic displacements on the destruction of persistent CAPs we propose a series of idealized large-eddy simulations (LES) informed by PCAPS observations. LES modeling is used, rather than Reynolds averaged Navier-Stokes simulations (RANS), to avoid reliance on boundary-layer parameterizations for the mixing of heat and momentum, which are a known short coming of previous numerical investigations of CAP breakup (e.g. Vrhovec and Hrabar (1996); Rakovec et al (2002); Zangl (2003)).

To avoid the use of these parameterizations our simulations must resolve the inertial scale eddies within the CAP that are responsible for mixing. Rough estimates from the 3 December partial mix-out indicate that the largest eddies and waves are of O(100 m) in the vertical and O(1000 m) horizontal. Lesser eddies occurred at scales O(10

m) vertically, and *O*(100 m) horizontal. To explicitly resolve these motions the model must be run at a resolution of at least ½ of the smallest eddies of interest, suggesting a resolution of 50 m in the horizontal and 5-10 m vertically is the minimum appropriate scale. Modeled spectra of turbulence kinetic energy (TKE) will provide an objective measure of the adequacy of the chosen resolution, as will comparison with eddy-covariance measurements during PCAPS. Adjustments to the model setup will be made as necessary.

<u>a. Experiment Design</u>

Figure 14 shows an idealized CAP profile used in a prototype simulation. Both the temperature and wind profiles are constructed from hyperbolic tangent functions. The wind profile goes to zero at the surface. The cold pool is initialized only in the valley between the two topographic barriers while the atmosphere outside of the terrain is neutral (Fig. 15a). In this case the domain is 15 km long, the mountains 400 m tall and separated by 5 km at ridge crest, and boundary conditions periodic. The top boundary is at 2 km, and the grid spacing is 100 m horizontal and 50 m vertical.

The modeled environment is characterized by the upstream profile, the CAP profile, and the dimensions of the confining topography. From these quantities it is possible to form a number of dimensionless parameters, including bulk and gradient Richardson numbers (e.g. Fig 14d), the non-dimensional mountain height, and the non-dimensional valley depth. Additional parameters of interest include the length scales associated with shear and density interfaces across the CAP, which in this case are given by the hyperbolic tangent functions. To work within the practical constraints of computing time, we will explore model results over a limited range of these parameters, focusing on scales representative of the CAPs found within the Salt Lake valley.

b. Initial Results

Figure 15b-d shows the state of the prototype case at 50-minute increments. Evolving from an initial state of homogenous stratification, after 50 minutes the CAP has thinned everywhere, but much more so on the upstream end of the valley. Cold air is being advected over the downstream ridge from the unblocked upper portions of the CAP, and replaced by warmer upstream air. Notably the thinning does not initially weaken the stratification, but rather sharpens capping inversions without mixing high potential air into the CAP.

A frontal interface forms at the extreme upwind end of the basin after 100 minutes. The front exhibits a density current head with a region of mixing downwind. Larger amplitude waves form with increasing fetch through the basin, and surface temperatures have warmed everywhere expect in the leading edge of the front. After 150 minutes, the warm front is located in the downwind half of the basin, but has lost its characteristic density current head, now more closely resembling the warm front seen in observations. Cold pool air continues to be advected out of the basin and waves have partially mixed the CAP, which is now deeper and less stable. The CAP is finally entirely removed from the basin after 180 minutes (not shown).

These preliminary results are encouraging and provide evidence that LES simulations can capture the salient features of dynamical CAP destruction. For example, the model produces thinning and sharpening of the CAP via differential advection, as well as divergence of the CAP air, which "drains" over the downstream mountain crest. The model also produces a warm front, density current structures, and myriad waves within the CAP.

A number of challenges remain with these simulations. For example, in order to produce an accelerating wind aloft, but to still use periodic boundary conditions, we will need to prescribe a constant momentum flux at the model top boundary. This will be accomplished by modifying the WRF advection module to add a constant tendency term for the vertical flux of horizontal momentum to the advection computation. Tests for this process are ongoing. In addition, during the 3 December mix-out case, we observed that the unsteadiness of the wind may contribute to the "sloshing" of the CAP. We will implement a time varying forcing to approximate this process. Moreover, defining and refining the parameter space that we intend to test will take time, careful consideration, and numerous adjustments.

Real Data Simulations

A portion of this research uses WRF-ARW real-data simulations of PCAPS IOPs to investigate the model ability to reproduce the CAP lifecycle. In particular, we are interested in PCAPS IOP-9 as it offers an example of the premature destruction of CAPs in models that rely on parameterized mixing.

Figure 16 shows a time-height comparison of the WRF simulation, run at 1.35 km, with the potential temperature observations from IOP-9. The WRF systematically under predicts the CAP depth, an error related to the misrepresentation of stratocumulus clouds. This model error subsequently causes the premature removal of the CAP during a period of strong southerly winds aloft. These strong winds are also present in observations, but do not significantly alter the CAP.

Interestingly, the WRF simulation produces a warm front type mix-out during IOP-9 similar to that documented in Section 2. The ability of the WRF to capture this morphology is encouraging, though in this case completely erroneous. These simulations underscore the difficulty in forecasting the timing and extent of CAP breakup, and further motivates the use of observations and idealized LES modeling to help improve our understanding of these events.

Further simulations using borderline LES scales (~450 m) for real data cases are also planned. The goal of this experiment is to determine whether or not improved resolution can affect the model fidelity in handling the structure of CAPs. The results of these simulations are not central to goals of this thesis, but rather an opportunity to further explore the lifecycle and forecasts of persistent CAPs specifically occurring the Salt Lake valley.

6 RESEARCH TIMELINE

December 2012 – January 2013: Refine idealized LES simulation setup, define parameter space, execute model runs.

February-April 2013: Process model results and execute additional runs as necessary. Synthesize model results and observations.

May -July2013. Distill results into a conceptual model for CAP breakup. Prepare thesis manuscript and figures.

August 2013: Thesis defense. Submit journal articles for review.

September 2013: Transition to a post-doc or other employment opportunities.

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