2002

Marine stratus analysis over San Francisco Airport using LAPS

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DOI: <https://doi.org/10.31979/etd.8g6j-czhy>
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Marine Stratus Analysis over San Francisco Airport using LAPS

**A Thesis Presented to
The faculty of the Department of Meteorology
San Jose State University**

**In Partial Fulfillment
Of the Requirements for the Degree
Master of Science**

**by
Giovanni Leoncini**

December 2002

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ABSTRACT

Marine Stratus Analysis over San Francisco Airport using LAPS

by Giovanni Leoncini

The current study uses Local Analysis and Prediction System (LAPS) cloud analyses at a 3.3 km resolution to study marine stratus (MS) dynamics over San Francisco International Airport (SFO). A thick MS layer caused a late clearing over SFO due to trough induced lifting and adiabatic cooling. LAPS incorporated surface and upper air data, used ETA forecasts as first guess fields, and evaluated its cloud fields against visible satellite images. Four LAPS output parameters served as cloud proxies: LWC, FCC, and unbalanced and balanced RH. Grid points with proxy values above the specified thresholds were defined as "cloudy."

ETA analyses in conjunction with the LAPS default threshold values were too dry to produce cloud fields. Lower thresholds produced good agreement with satellite observations for all parameters except LWC, which depends strongly on initial RH values. Further studies should address effects on LAPS analysis fields from first guess fields, terrain resolution, and observational data density.

To my grandfather Alessandro

Acknowledgements

First and foremost, I would like to thank my advisor Dr. Robert Bornstein for his invaluable guidance throughout this research, and most of all for giving me the opportunity to do this work and move on to further graduate studies. Both would not have been possible without his help. To him goes my gratitude. Drs. Doug Sinton and Jindra Goodman, my other committee members, are to also be thanked for their insights and quick turnarounds. I would like to thank the entire LAPS Branch staff at FSL for carefully guiding me through the use of LAPS.

I also wish to thank the Faculty and Staff at the Meteorological Department at San Jose State University (SJSU) for their lectures and for patiently answering countless questions. In particular, I thank Mr. Mike Voss who I bothered most, and Prof. Scot Rafkin for his guidance and assistance during the Bay Area Mesoscale Initiative project.

I do not forget my colleagues and friends at SJSU, who made the Met Dept a friendly and diverse environment. Among them Luca, with whom I shared this 2 ½ year long effort, deserves special thanks for his invaluable support and friendship.

This research was funded by Lockheed Martin Corp. and the NOAA/UCAR Bay Area Mesoscale Initiative project.

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1. INTRODUCTION

San Francisco International Airport (SFO) was the fifth busiest U.S. airport during 2000, but second in flight delays (ACI 2001). For concurrent SFO (all acronyms defined in Appendix) landings on parallel runways, good visibility must exist over the approach zone (Farley 2000), which for SFO starts 900 m over San Francisco Bay and extends about 8 km to the northwest. With excellent visibility in the absence of other phenomena, e.g., strong winds, the landing rate is about 60 aircraft per hour, but with poor visibility, it drops to 30. During the peak hours of 0800 to 1200 and 1800 to 1900 LST (8 h behind UTC), over 30 aircraft are scheduled to arrive each hour, but from June to October marine stratus (MS) penetrations reduce visibility and hence acceptance rate.

The FAA, MIT Lincoln Laboratory, and San Jose State University (SJSU) have developed statistical forecast methods for SFO MS burn-off time (Clark and Wilson 1996; Hilliker and Fritsch 1999). Improved forecasts can result from an increased understanding of physical, dynamic, radiative, and turbulent influences on MS cloud formation, movement, evolution, and dissipation, and on local topographic and meso-meteorological influences on these processes.

a. Marine Stratus

The static stability of the eastern Pacific atmosphere arises from the semi-permanent Pacific High. Its subsiding air warms, dries, and forms an inversion atop the surface-based marine atmospheric boundary layer (MABL), a moist cool layer due to cool sea surface temperatures (SSTs). The summer High is well developed, centered off-coast adjacent to the strong inland California thermal low, which forms due to radiative heating

over the Sonora and Mojave deserts and Central Valley (CV). Resulting increased land-sea temperature gradients produce a sloping inversion base, lower and steeper near the coast (Neiburger et al. 1961), while resulting pressure gradients accelerate air inland, inducing strong offshore low-level divergence. A winter southwestward shift of the High weakens (if not dissipates) the inversion. This allows synoptic disturbances from the northern Pacific to move southeastward, over northern and central California.

Neiburger et al. found that Pacific High air originates over the Gulf of Alaska at about 3 km MSL and then sinks to the MABL top, which deepens by convection. The MABL is thus deep over areas with high SSTs and low subsidence rates. Over Hawaii (west side of high), where SSTs are high and subsidence weak, MABL depth is thus high (2000 m), while on its eastern side (central California), maximum subsidence exists. In addition, minimum SSTs exist there due to coastal upwelling, and thus the MABL is shallow (<400 m). Simulations by Burk and Thompson (1996) showed that coastal SSTs influence MABL heights through interactions with both mesoscale flows and topography. Dorman et al. (2000) also observed maxima and minima MABL depths up- and downwind of major coastal capes, respectively.

Ocean minus air temperature values are negative/positive north/south of Point Reyes. The Washington-Oregon border has a relative maximum positive difference, Point Arena a maximum negative value, and the Southern California Bight a positive maximum. Surface heat flux can thus be either up- or downward directed, but is always too small to drive cloud formation or development.

West Coast summertime surface winds parallel the coast, except off the northern Oregon coast where light and variable flows exist (Dorman and Winant 1995). Speeds increase southward of Point Reyes to a second maximum off Point Conception, where the flow splits: most to the south-southeast, the rest into the Santa Barbara Channel (Dorman et al. 2000). Off shore speeds are faster than those inland, the strongest cross isobar component exists over the Southern California Bight, with speed maximum offshore at 1500 LST. High sea surface wind stresses were observed within 100 km of the coast between Cape Mendocino and San Francisco, while stresses were lower immediately north of Point Conception, as large stresses increase upwelling (Burk and Thompson 1996). Shallow MABLs (offshore and inland) are thus associated with strong winds.

The 10 m winter wind-energy spectrum off Point Reyes shows a strong annual peak, while synoptic frequencies were characterized by a flat continuum (no well defined peak) at a lower energy level. These two time scales equally account for most of the energy, with sharper peaks (same magnitude as synoptic plateau) at diurnal and semi-diurnal frequencies (Dorman and Winant 1995).

Mesoscale MABL features in the San Francisco Bay Area (SFBA) are generally consistent with those in Dorman et al. (Bridger et al. 1993). Its inversion was stronger and lower in the east, and gravity waves were present, consistently with Goodman and Miller (1977) who found such waves responsible for mass fluxes through the inversion. A strong low level jet (LLJ) paralleled the coast (either on it or offshore) from Oregon to the Southern California Bight.

Simulated LLJ core speeds from Burk and Thompson (1996) agreed with observations, e.g., they were right above the inversion base and peaked around 2100 LST, when the jet was highest above the ground and furthest from the coast. During daytime, it weakened, lowered, and shifted back towards the land. Results showed that:

- LLJs arise from baroclinity above the MABL.

- Sea breeze flows create offshore divergence, lowering both the inversion and LLJ.

Onshore flow weakens thermal contrasts and moves the strongest gradient inland, and thus the offshore LLJ moves coastward.

- LLJ time lag between maximum land-sea temperature gradients and LLJ core is explained by the above mentioned baroclinicity and a sea breeze inertial oscillation around an average daily velocity.

- East-west SST gradient does not explain westward MABL deepening.

- LLJs flow along the coast (northwesterly), whereas geostrophic winds are northerly, accelerates the LLJ on its eastern flank where convergence forms. Divergence on its western side produces a sloped inversion and the minimum cloudiness line observed by Simon (1977).

- Horizontal inversion-base height and wind speed variations are related to topography. Upstream of capes with tops above the MABL, winds slow due to trapping, convergence occurs, and inversion bases elevate. Downstream winds accelerate (no trapping), divergence occurs, and inversion bases lower, explaining the zero-convergence line seen by Simon (1977). Hydraulic jumps and expansion fans are thus observed upstream and downstream major capes, respectively.

Idealized simulations of Central California MABLs by Cui et al. (1998) showed gravity waves downwind of the Sierras lower MABL depths and weaken LLJs. Waves within the MABL inversion over California, however, do not generate clouds over the Bay (Bridger et al. 1993; Burk and Haack 2000). Mesoscale and synoptic features also trigger coastally trapped disturbances and eddies, e.g., Santa Cruz Eddy, a summer afternoon meso- γ circulation over the Monterey Bay (Lozej 2001). The former are often followed by MS surges and temperature drops that propagate northward from the Southern California Bight (Nuss et al. 2000) and can affect SFO (Bond et al. 1996).

In the low Stratus-cloud classification scheme of Klein and Hartmann (1993), Mid-Latitude Stratus forms during wintertime over warm western-boundary ocean currents, which produce convection driven by surface heat fluxes. When moist subpolar air enters arctic regions, however, Arctic Stratus forms in multiple layers. Subtropical Stratus forms on the eastern and southern sides of ocean subtropical highs, where northerly flows pass over cold water and a subsidence inversion caps a MABL. The cold water reduces surface induced convection, and hence radiative cloud-top cooling drives convection; California Stratus belongs to this class.

Klein and Hartman (1993) further related over-ocean clouds to SST and static stability (difference between 700 hPa potential temperature and SST). As SSTs in the California Stratus region were only slightly warmer than surface air temperatures, resulting weak upward-directed surface heat and moisture fluxes could not produce observed clouds. Other mechanisms that decrease MABL stability, such as cloud-top radiative cooling, are thus important.

Annual variations of air-surface temperatures and SSTs were small, with an August to October peak and March to May minimum. While surface divergence also did not show large seasonal variations, static stability did, because of the wide range of 700 hPa potential temperatures. Cloud amount also showed a large annual variation, with a July maximum and a December to February minimum, both earlier than that of SST, and thus cloud amount was best correlated to static stability.

The minimum cloudiness line (due to maximum divergence) along 125 W found by Simon (1977) strengthens as summer proceeds. West of this line, stratocumulus is dominant and its cover increases westward. Over the eastern area, however, either clouds are absent or mostly Stratus, which increase eastward. Cloud-cover variance was thus higher near the coast. Although the coverage in both regions was maximum at 0745 LST and minimum at 1345 LST, western coverage changed over the diurnal period by 13% (due to radiative heating), while along the Central coast the change was only 5%. It was 30%, however, over the warm SST Southern California Bight, where sea breeze flows pushed marine air inland, where its moisture diminished.

While radiative flux divergence from water vapor cools air below its dew point, and thus is important in fog evolution, it is not necessary for its formation. Fog formation mechanisms depend solely on surface momentum, heat, and moisture fluxes and on atmospheric temperature, moisture, and wind gradients. Oliver et al. (1978) showed that fog forms with positive or negative surface heat fluxes, radiation effects, and near saturated cold air overrunning warm water; Pilié et al. (1979) have thus observed California MABLs filled with fog. Upwind of this layer, air was warmer than the SST, whereas in

the fog it was colder. Low-level upstream winds were northwesterly, while within the fog speeds were lower and even reversed in direction. Resulting convergence persisted for > 20 h, and induced upward velocities of about 1 cm s^{-1} that initiated condensation. Since corresponding downdrafts would evaporate the same amount of liquid, cloud top radiative cooling must condense the additional liquid transported to the surface. They also reported fog streets with surface upwind edges and lifted downwind (MS line) edges, which also form from ship plumes (Cotton and Anthes 1989; Durkee et al. 2000).

MS microphysical measurements over subtropical areas, reviewed by Miles et al. (2000) showed the following:

- droplet number (N) constant with height above cloud base, while continental cloud N increases with height
- marine clouds have larger effective droplet diameters, same liquid water content (LWC), and lower N than continental clouds (Pruppacher and Klett 1997)
- droplet size deviation (σ) increases with increasing mean diameter (D), and therefore with height
- marine and continental spectral-dispersion (σ/D) profiles show similar values, trends, and two subsets (from two phenomena).

Drizzle is only a minor component of LWC, but dynamically important for cloud evolution and usually poorly documented. Goodman (1977) found SFBA fog parameters within the range of observed MS values of Miles et al. (2000), except that drizzle size was above the upper limit of their instruments. Fog moisture was advected from the ocean (but not through MS bases) and orography induced vertical motions compressed

the MS. Analogous drizzle size distributions, profiles, and spectral broadening with height and time were also found by the sea fog study of Mack et al. (1974), but drop size increased with altitude, contrary to land fog. Fog and MS similarity was demonstrated by a MS deck that lowered to the sea surface, producing fog. Horizontal drizzle variability studied by Paluch and Lenschow (1991) and Vali et al. (1998) could be due to gravity waves or internal turbulent circulations (Miles et al. 2000).

MABL MS thus results from complex interactions among the physical processes discussed above (Stevens 1996). An upper level trough passage over California produced upward motions above the MABL that lifted and weakened the inversion, allowing MS formation. Koracin et al. (2001) simulated large scale subsidence influences on MS to fog transformations off Central California.

A 1-D version of the second-order closure model of Yamada (1978) in a Lagrangian framework accounted for advective effects. Results matched observations, which showed the MS compressed and lowered to the sea surface (to form fog), even as surface heat flux magnitude increased. Below a critical inversion strength, fog never dissipated, but stronger subsidence produced fog earlier and dissipated it quicker. Strong inversions produced sharp moisture and temperature gradients, and hence intense radiative cooling, which increased condensation. Strong subsidence, however, ultimately warmed and dried entraining air, thus evaporating the fog. When upper level mixing ratios were reduced, stronger radiative cooling resulted from the increased MABL-free atmosphere moisture gradient. The dry MABL initially had a slightly larger LWC than the moist one,

but moist air entrainment produced stronger condensation later, independent of radiative cooling rate.

MS topped Northern California MABLS are characterized by weak surface heat fluxes, strong surface moisture fluxes, strong evaporative cooling, and intense top cloud radiative cooling. The latter mechanism, the strongest source of MABL buoyancy, keeps it well mixed, so that MS extends to the inversion base. With intense cooling, droplets precipitate from the MS base, evaporate, cool the sub-cloud layer, increase dew points, cause new condensation, and thus lower its base (Pilié et al. 1979).

If cloud air mixes into a warm downward entrained parcel from the inversion layer, it lowers its virtual temperature below the cloud value, the parcel becomes negatively buoyant, triggering turbulence and further downward entrainment (Lilly 1968; Deardoff 1980; Randall 1980). Later refinements considered liquid water (MacVean and Mason 1990) and the energy used for mixing (Siems et al. 1990). While this describes cloud-top entrainment-instability onset, it does not explain long-lived MS in such warming regimes (e.g. Albrecht 1991).

Moeng et al. (1995), however, proved that entrainment effects on MS dissipation depends on two competing mechanisms: (1) decreased radiative cooling (due to reduced LWC) that implies weaker turbulence (and diminished entrainment) and (2) increased evaporative cooling that enhances entrainment. When the former is stronger, clouds can entrain air without dissipation since radiative cooling restores the LWC lost by evaporation, while if the latter dominates MS dissipates. With strong buoyancy and large surface moisture fluxes, however, the MS grows upward even in the latter case (Randall 1984).

Moeng et al. (1996) noted the agreement of turbulent statistics and MABL structure among LES models, while Stevens et al. (2000) found modeled entrainment rate variations smaller than observational uncertainty. As model vertical resolution increased to 4 m, existing structures were refined, i.e., inversion depth thinned to a finite value and peak specific humidity increased. While the number of down-welling events increased and their area decreased, their thickness was insensitive to resolution.

Despite these encouraging LES results, additional research is needed to further explain entrainment turbulence, e.g., Lewellen and Lewellen (1998) proposed that better resolved turbulence would better reproduce cloud evolution and entrainment. Lock and McVean (1999) suggested entrainment as a function of net turbulent kinetic energy (TKE), and Bretherton et al. (1999) and Stevens and Bretherton (1999) suggested entrainment is controlled by the distance required for inversions to slow rising parcels.

Entrainment is also important in the transition of MS to stratocumulus and then to trade wind cumulus (Wyant et al. 1997). This occurs when stratocumulus is advected from low to high SST areas (Albrecht et al. 1995) or offshore of Northern California where decks turn into cumulus as SSTs warm to the west. This does not include cloud-top entrainment-instability, since this is not necessary for break up (Kuo and Schubert 1988; Krueger et al. 1995). Wyant et al. suggested that warmer SSTs cause increased latent heat fluxes that deepens MABLs, lift clouds, and enhance warm-air entrainment efficiency against radiative cooling. This allows entrained air to reach the sub-cloud layer, and thus creates a weakly stable layer below the cloud; thus only strong surface updrafts can penetrate stratocumulus. The MABL is thus composed of two well mixed

layers separated by a stable transition layer. As entrainment proceeds the decoupling becomes stronger, the cloud deck thins, and updrafts generate small cumuli that transport moisture above the stable transition layer (building conditional instability). With time (and/or increasing SSTs) more updrafts reach the thinning stratocumulus, increase turbulence above the transition layer, penetrate into the subsidence inversion, and entrain dry warm air that dissipates the stratocumulus.

While solar radiation modulates the process, it does not alter its time scale (~ 2 days). Shortwave absorption depends mainly on liquid water path and warming is maximum in cloud interiors (Cotton and Anthes 1989). After sunrise shortwave convergence warms and dries clouds (whose bases elevate), and thus LWC and radiative cooling diminish. The cloud is still turbulent (due to cloud-top radiative cooling), but can become warmer (due to radiative warming) than the sub-cloud layer, again forming a stable transitional layer. Upward moisture flux into the cloud is reduced, eventually resulting in its break up. Late in the afternoons Hignett (1991) found longwave radiative cooling dominant, LWC increased, and horizontally and vertically growing MS. Radiative and evaporative cooling thus keep MABLs well mixed, but shortwave absorption by clouds is under-predicted in models due to an incomplete understanding of these radiative processes (Stephens 1996; Fu et al. 2000).

Observations by Brost et al. (1992a, b) showed shear as the dominant TKE source in stable or weakly unstable stratocumulus topped MABLs. Simulations by Chen and Cotton (1987) also showed a 15 to 20 min periodicity in night-time cloud-top entrainment due to large eddies (from radiative top-cloud cooling). These eddies produce: well mixed

MABLs with undulated tops; upward convective elements that penetrate inversion bases and increase shear; and entrainment from breaking Kelvin-Helmoltz billows. LWC is thus lowered, and the resulting diminished radiative cooling stabilizes the cloud, allowing cooling to increase, thus restarting the cycle. Cycle timing was correctly simulated only when drizzle was included, as it accelerated LWC reduction.

Drizzle forms in the upper part of MS, and thus deep clouds have large drizzle rates (Feingold et al. 1996). Its formation produces cloud-top latent heat release, while precipitation cools and moistens sub-cloud layers. The resulting increased stability of the MS and sub-cloud layers reduces both downdraft motions from the cloud and resulting cloud-top entrainment (Stevens et al. 1998). Drizzle-evaporation forms another conditionally unstable transition layer below (either the MS or sub-cloud layer) that may (depending on SST values) protect the transition and sub-cloud layers from surface turbulence. Turbulence is generated by the "bottom layer" (surface to transition layer base) instability, which results from sub-cloud evaporative cooling and/or high SSTs. Fluxes supply moisture and heat to the sub-cloud layer (coupling it to the bottom layer) and producing cumulus cloud formation during strong drizzle. The original MABL structure is restored a few hours after the precipitation stops, i.e., the transition layer dissipates, preventing formation of trade wind cumulus.

In summary, high SSTs and deep MS drizzle drive all MABL sub-layers into a coupled equilibrium state. Non-equilibrium decoupling is achieved by low SSTs (that diminish surface fluxes) (Paluch and Lenchow 1991), and secondarily by cloud-top shear-induced TKE (that mitigates drizzle-induced sub-cloud layer buoyancy).

Microphysical effects on MABL radiative properties were modeled by the 1-D model of Ackerman et al. (1995), and results underestimated peak saturation values and overestimated in-cloud residence-time and lower-layer evaporation, as the model averaged downdraft and updraft effects. With a narrow initial cloud condensation nuclei (CCN) spectrum, N increases, fewer CCN activate, drizzle rate decrease, and albedo and optical depth increase (Albrecht 1989; Baker and Charlson 1990).

Explicit interaction of long wave radiation with CCN spectra (as opposed to bulk formulations) in the LES model of Jiang and Cotton (2000) correctly increased radiative cooling. Correct cloud albedo and optical thickness resulted from interactions between CCN spectra and concentration, drizzle droplet concentration (N), and total liquid water mixing ratio. LES models are thus necessary to reproduce observed MS TKE features.

b. LAPS

The Local Analysis and Prediction System (LAPS), developed at the NOAA Forecast Research Laboratory (FSL), produces 3-D objective meteorological analyses on a relocatable model domain (changed without recompilation). It ingests 3-D meteorological model fields and data, including NWS surface, maritime, local mesonet, radar wind and rainfall, pilots voice report (PIREPs), Aircraft Communications Addressing and Reporting System (ACARS), wind profiler, Radio Acoustic Sounding System (RASS) and SODAR profiles, RAOB soundings, satellite, and GPS.

LAPS is part of the Advanced Weather Interactive Processing System models, run hourly at U. S. NWS Weather Forecast Offices (WFOs). It utilizes Rapid Update Cycle (RUC) model output as first-guess background fields, with ETA model data as a backup

choice. Other LAPS users include the Taiwan Central Weather Bureau, Thai Meteorological Department, Meteorological Research Institute of Korea, and U. S. Range Standardization and Augmentation Space Center.

LAPS also determines cloud parameters, e.g., cover, top, base, ceiling, albedo, vertical velocity, LWC, droplet mean-diameter, and precipitation rate and type. The Regional Meteorological Service of Italy ingests METEOSAT data into LAPS (Alberoni et al. 1998), while the CREST facility at Lockheed Martin Space and Missiles Co. ingests Polar-Orbiting Environmental Satellites data (Bennett et al. 2000).

LAPS output also has been used to initialize mesoscale models. The University of Oklahoma version (OLAPS) initializes the ARPS thunderstorm model (Brewster et al. 1994), while the China Meteorological Administration runs it operationally with the Regional Atmospheric Modeling System (RAMS). The Office of Naval Research uses it with RAMS on a nested grid domain for the Chesapeake Bay Operational Forecast System. The Air Force Weather Agency uses it in a Four Dimensional Data Assimilation (FDDA) scheme for the Fifth Generation Mesoscale Model (MM5). They run LAPS on 12 domains around the globe several times per hour using the latest observations to produce analyses for a dynamic nudging of MM5 (Smart et al. 2000).

FSL developed a "hot-start" cloud-balanced mesoscale model initialization scheme for LAPS (McGinley and Smart 2001), which allows cloud formation in equilibrium initial-condition fields. Input initial supersaturation vapor pressures in mesoscale models are normally set to subsaturation values to prevent inconsistent initial cloud and vertical velocity fields. Such models thus normally require a >9 h spin up time (during which

clouds can form) to produce the accurate precipitation forecasts now possible with the new scheme (Shaw et al. 2001). The new scheme have been used to analyze thunderstorm (Birkenheuer et al. 2001) and snowfall (Szoke 2000) events and to support the 1996 Summer Olympics (Stamus and McGinley 1997).

c. BAMI

Bay Area Mesoscale Initiative (BAMI) is a consortium between SJSU, Naval Postgraduate School (NPS), and NWS Weather Forecast Office (WFO) in Monterey. It provides real-time web-based distribution of meteorological data from local mesoscale observational networks. NPS retrieves data from various institutions (e.g., California Department of Forestry, Mathematics and Science Teaching Education Project at SJSU, NWS, and San Francisco Bay Area Air Quality Management District), while SJSU developed and maintains the BAMI web site (<http://meso.met.sjsu.edu/bami>) to eliminate non-meteorological information and convert data into uniform archival formats.

Reformatted data are sent to the MesoWest network at the University of Utah, a BAMI-type project for the western U.S. Data then go into the Local Data Acquisition and Dissemination (LDAD) system of the Western Region NWS, which allows WFOs to collect, check, and share local data. LDAD also makes data available via the Forecast Research Laboratory display at <http://www-frd.fsl.noaa.gov/mesonet/>. Local emergency management agencies use such data to obtain information on flooding, fire danger, road icing, and severe weather.

Whereas previous research on summertime SFO MS events has focused on either statistical or simple-modeling approaches, the current effort seeks to use LAPS analyses

to increase understanding of its dynamics and thermodynamics. LAPS skills are evaluated comparing its cloud analyses against visible satellite images, not used in its analyses.

2. METHODOLOGY

While comparisons of LAPS performance against other analysis schemes (e.g., ADAS from the University of Oklahoma) are not yet available, it is still a valuable tool to understand the SFO MS dynamics discussed above. It sequentially ingests data, uses analysis schemes, derives secondary fields, and then uses variational approaches to develop reciprocally balanced variables (e.g., horizontal and vertical wind, geopotential height, and specific humidity). It performs quality checks (QCs) during all ingestion and analysis procedures to detect inconsistent input data and to correct analyses errors. Data ingestion first eliminates values outside the LAPS domain (or for some variables, beyond predetermined buffer zones) and then interpolates input model values to its grid.

LAPS first interpolates and extrapolates ingested data by use of flexible (based on data density) temporal and spatial weights specific to each data type. Variational schemes in LAPS include Birkenheuer (2001) minimization to adjust specific humidity fields and then dynamic constraints in McGinley and Smart (2001) to assure quasi-geostrophic balance.

First guess fields for each analysis routine in the WFO version of LAPS (Fig. 2) are provided by large-scale model outputs interpolated to the LAPS grid. This produces 3 D and surface fields of: temperature, u- and v-wind components, geopotential height, and dew point. At the start of each FDDA analysis cycle, background fields are provided by either the large scale model (i.e., RUC, ETA, or AVN) LAPS is initializing, another large-scale model, or (as a last resort) a previous LAPS analysis. Incomplete surface observations are filled-in by a time-space Kalman filter (McGinley 2001).

Polar and NOWRAD 3-D surface-based radar data (i.e., reflectivity, radial velocity, and Nyquist velocity) are then mapped onto the LAPS grid, negative reflectivity values discarded, and vertical interpolation (to up to four levels) is made between data from different scan angles. LAPS also assimilates data from profilers (i.e., RASS, RAOB, NOAA wind, and 915 MHz wind profiler), ACARS and PIREPs reports, satellite images, satellite soundings, radar, and cloud-drift winds.

A time-window for each data type is defined, based on its data collection frequency and the LAPS analysis cycle-time (between successive analyses). Temporal weights are not applied on data within a window, and outside data are discarded. When the "age" of ACARS (5 s averaged) and profiler wind (1 h averaged) data approaches the LAPS cycle time, however, they are updated using a time-tendency computed from the first-guess field and previous LAPS analysis.

Horizontal wind components are each independently analyzed with a two-pass algorithm (Fig. 3), while vertical speed is derived kinematically (Albers 1995); radar observations are used only in the second pass. The first-guess wind field is interpolated to observation locations, a residual field computed, and QC performed on residuals (threshold depends on data source). Values are spread vertically by use of empirical exponential (distance dependent) weights, except for radar and profiler data which have a higher vertical density. Residuals are analyzed at each level with Barnes (1964) exponential weighting, whose radius of influence is a function of local data density. Residuals and background values are then summed, with weights proportional to the inverse square of their respective errors.

New winds are compared with Doppler radar radial velocities, and if differences are $<12 \text{ m s}^{-1}$, the tangential component is obtained from the preliminary analysis and joined to the radial component. If grids are illuminated by multiple radars, least square solutions of their velocities with background winds are used. The first guess field is then again subtracted, and the loop repeated until analyzed winds fit observations within instrument error magnitude. This process produces a smaller influence radius at each pass, seamless integration of multiple radar measurements, smooth blending with regular observations, and enhanced small-scale structures. Finally the 3-D wind analysis is linearly extrapolated to 10 m to produce first-guess values, while first-guess surface values for all other variables equal background values.

Each surface variable is first independently analyzed, and results inside and outside the LAPS grid are to set boundary conditions by use of a Barnes analysis and cubic spline, which matches interior grid point values to observations. Surface analyses then use a successive correction technique, similar to that used with the 3-D winds. While visibility observations are used directly, observation residuals are used for the u- and v-wind components, temperature, and dew point, with corrections for station elevation.

Three fields are computed for the pressure analyses (i.e., reduced to a chosen reference level, mean sea-level, and surface), which relies on corresponding background values and observational residuals. A third type of variational technique then constrains surface winds and pressures to the full planetary boundary layer (PBL) equations of motion (Stamus and McGinley 1991). Surface visibility is first processed with the Barnes scheme, and then automatically decreased in high relative humidity (RH) areas and be-

low clouds bases (from the previous analysis cycle). Several forecast indices are then calculated, and then all analyzed and derived values are QCed by comparison with (after backwards interpolation) observational values.

A 3-D temperature analysis that uses ACARS, RASS, ROAB, and satellite soundings is performed in the same sequence as with the speed components, but vertical sounding-tops are smoothed. Resulting surface temperatures at grid points with pressures > 750 hPa replaced by values from the 2-D analysis, as follows. A PBL is first defined (50 hPa above terrain) with bottom temperatures equal to 2-D surface values and top values equal to 3-D values (background plus soundings). Corrections to the 3-D analysis are assumed linear throughout the layer.

Superadiabatic layers have upper grid values set equal to lower grid values. This starts at the lowest level, and thus allows surface temperatures (most timely) to define mixing layer values. Height fields are computed via hydrostatic integration of the 3-D temperature field, with 500 hPa values set to first-guess background values (only boundary condition). This strong constraint limits the vertical extent of surface effects, and forces strong convective effects to enter the analysis above the 500 hPa level through other data sources (e.g., satellite cloud-top temperatures).

Cloud analysis processes (Fig. 4) use data from METAR and pilot reports over a region slightly larger than the LAPS domain. Horizontal analysis generates a preliminary 3-D analysis, which provides information on cloud-layer vertical location and horizontal distribution. Satellite derived cloud-top temperatures (GOES 11.24 μm IR imagery) are converted to cloud-top heights by use of the 3-D temperature field, with apparent bright-

ness temperature as a weighted (determined by cloud fraction) average of cloud-top and ground values. Clouds are then inserted, with missing parameters (e.g., tops or bases) assumed. Conflicts between satellite and METAR cloud data are resolved for low or warm clouds, as METAR data overrule satellite data. For higher clouds, however, they only provide potential locations for vertical layers, while coverage is controlled by satellite data to ensure consistency with measured brightness temperatures. Radar and visible satellite data provide further details.

Once fractional cloud cover (FCC) is defined at each vertical level as area coverage multiplied by cloud layer opacity to visible or infrared radiation, cloud and hydrometeor fields are determined (Fig. 5). At each "cloudy" grid point ($FCC > 0.65$), ambient and equivalent potential temperature vertical derivatives (from the temperature analysis) determine cloud type and mean volumetric drop size through a look-up table (Table 3) from Haines et al. (1989). In-cloud vertical velocity is then computed as a function of cloud depth and type. A modified Smith-Feddes model is applied to grid columns in cloudy regions to produce vertical profiles of liquid water and ice content. Cloud ceiling is also computed. If $FCC > 0.10$, cloud base and top height are estimated.

If $FCC < 0.65$, cloud particle diameter is assumed $< 20 \mu\text{m}$, the relative amount of each water phase is varied linearly for temperatures between 263 and 243 K, while a simple radiation model provides ice and water mixing ratios. Ice and LWC are adjusted according to radar observations, which diagnose hydrometeor type following a scheme similar to Ramer (1993). Finally snow cover, accumulation, and liquid equivalent are computed by use of a soil moisture model that uses surface, radar, and satellite data.

Specific humidity analysis involves: first guess fields interpolated (spatially and temporarily) to the LAPS grid, surface moisture mixed through the lowest 50 hPa, and moisture values adjusted by variational technique. Satellite data and the radiance model of Hayden (1988) limit this application to clear sky grid points. A cost function minimizes the maximum allowable distance from various observations, i.e., RAOB, GPS derived total-column water vapor, GOES derived layer precipitable water, and radiance. A Barnes analysis fills in cloudy or skipped grid points, and a final QC ensures a saturated atmosphere at cloudy grid points.

Final balancing is again achieved by use of a variational technique, whose cost function consists of deviations from observations for 3-D winds, geopotential height, 3-D divergence, and horizontal velocity time-tendency. The scheme uses a quasi-linearized version of the equation of motion and ensures consistency between the cloud and wind fields by use of cloud-analysis vertical velocities.

3. OBSERVATIONAL ANALYSIS

a. Episode Selection

Criteria to select the study event include existence of extensive nighttime MS coverage over the SFBA and a "difficult to forecast" late MS burn off. Late is defined as after 1300 LST, when the SFO landing rate typically reaches 45 per hour. The event day also had to be between days with more "typical" burn off times, so that comparisons between the three days would increase understanding of MS phenomena. The 24 h period from 0000 UTC on July 12 (1600 LST on July 11) was thus selected as the event day.

The study incorporates the following data into LAPS: RAOB, NOAA 915 Hz profiler, ACARS, METAR, mesonets (BAMI included), buoy, and ship. Six and 12 h ETA forecasts (issued every 12 h) are used as first guesses and then linearly interpolated to the hour (LAPS analysis time increment). LAPS MS results are compared against 1 km resolution GOES-10 visible images.

The LAPS default configuration is for a Colorado analyses, and has thus been modified to account for the strong ageostrophic flows during the current MS event. Two variational surface analysis parameters were thus increased (within suggested ranges) to allow for larger pressure field modifications and momentum equation residuals. The lower SFBA mean-topographic elevation was also accounted for, as the check for super-adiabatic layers was done between the surface and 700 hPa (versus default 500 hPa), and the pressure reduction level was lowered from 1500 to 500 m.

The study is comprised of five simulations, each with a domain centered over SFO and with a horizontal grid resolution of 3.3 km (Fig. 6). The domain has the 125

grid points in the east-west and 105 in the north-south direction. A vertical resolution of 50 hPa was used, and the domain extended from 1100 up to 100 hPa.

b. Synoptic Overview

ETA 500 hPa heights (Fig. 7) for 0000 UTC on 11 July (1600 PST on 10 July) show a long-wave trough over western British Columbia, with an axis that generally extends southward to a position offshore of Central California. Height gradients along the axis vary, and consequently wind speeds diminish from around 50 m s^{-1} over British Columbia to $10\text{--}15 \text{ m s}^{-1}$ offshore of Central California. A low between Quebec and Ontario produces winds $> 60 \text{ m s}^{-1}$ over the Great Lakes, while a ridge over the central U.S. and Canada (between the western trough and eastern low) results in a high over the Texas panhandle.

Twelve hours later (Fig. 8) as the trough moved eastward, the northern part of its axis developed a stronger negative tilt (axis oriented from northwest to southeast) in its southern part and a stronger positive one (axis oriented from southwest to northeast) in its northern part. While height gradients did not change much, core height values off California deepened by around 20 m. The low over Canada also weakened, did not move, and its wind speeds remained constant. As its area of maximum speed shifted southeastward, the Texas ridge weakened and shifted southward.

By the start of the episode day on 12 July at 0000 UTC (Fig. 9), the northern part of the British Columbia trough had moved northeastward, while its southern part still lagged behind, weakened, and developed a larger negative tilt. After 12 h (Fig. 10), the flow over the Western Canadian coast turned zonal, and the southern end of the trough switch-

ed to a positive tilt because of a westward movement of its southern tip. The low kept moving southeastward without weakening, which allowed height gradients in the high over the central U.S. to weaken. During the following 24 h (ending on July 13 at 0000 UTC), the movement of the low switched to a northeastward direction (Fig. 11), while the trough again increased its height values (by about 40 m). This allowed for a westward expansion of the 500 hPa high to a position over Arizona.

The 850 hPa ETA winds and heights for 11 July at 0000 UTC (1600 PST) show effects from the upper level trough, as a weak low center is offshore of central and northern California (Fig. 12), where it produced weak onshore flow. This flow is squeezed between the eastern edge of the summertime Pacific high (not present in the 500 hPa analyses) and the even lower heights over California associated with surface heating.

The offshore high produced a northerly geostrophic wind west of the offshore low, while the thermal low over the southwestern U.S. also produced weak inland flow inland over the Southern California coast. Onshore flow thus exists over the entire West Coast, as expected during afternoon sea breeze periods (discussed below). A cold pool of air is also seen at this level (Fig. 13) on 11 July at 0000 UTC in the area of the northern part of the trough off shore at the U.S.-Canada border due to advective cooling. Land-ocean temperature contrasts are strong all along the West Coast, but are weakened over the immediate Central California and SFBA by the onshore advection of cool marine air.

Twelve hours later (11 July at 1200 UTC or 0400 PST), surface heating weakened (Fig. 14), and thus the synoptic trough and its associated cyclonic circulation could intensify (core deepened by 20 m) (Fig. 15), while mean sea level pressure (MSLP) rose. This

caused Northern California winds to become southeasterly, thus advecting warm air from the Central Valley offshore of the SFBA. Temperatures over the California-Nevada deserts were still high (21°C), and on-shore winds were still persistent (although weakened) over Southern California.

During the following 12 h, to the episode start on 12 July at 0000 UTC, as the 500 hPa trough weakened slightly, the 850 hPa trough core also rose 10 m (not shown). The cold pool at this level was thus advected south to a position off Washington and is 1°C colder than 24 h before (Fig. 16). Zonal temperature gradients were weaker than 24 h earlier for the first 100 km off of the SFBA. Temperatures were also 2-3°C colder since they increased only slightly from their overnight values due to weaker than normal solar heating (discussed below).

At 1200 UTC on July 12 (Fig. 17), the 850 hPa trough is again 10 m weaker, and while the cyclonic circulation is weakened and extends over a smaller area, it still advects warm air over the Central Valley (CV) westward beyond the coast (not shown). Cold-pool temperatures off of Washington are thus a couple of degrees warmer than 24 h earlier; the still north-northeasterly flow has weakened temperature gradients. By 13 July at 0000 UTC (Fig. 18), the only evidence of the trough was a southward lobe of the 1540 m height contour, along the Central California coast, while concurrent temperatures (Fig. 19) were dominated by a sharp coastal gradient. The 1200 UTC (Fig. 20) height analysis showed the trough completely dissipated, the Pacific high moved eastward, and a weakened thermal trough over the Central Valley, the latter more evident in the temperature

field (not shown). Winds offshore of California are thus more northerly and stronger (25 versus $15\text{--}20\text{ m s}^{-1}$) than on the previous days.

ETA MSLP afternoon pressure and wind analyses at 0000 UTC on 11 July (Fig. 21) show no offshore trough. The Pacific High, however, is on the west side of the domain, as are strong coastal pressure gradients from British Columbia to Baja California (due to intense inland heating) and the thermal low over the southwestern U.S. An anomalous feature is the isobar bulge towards the high offshore of the SFBA that results from the weak upper level trough; they usually orient north-south (decreasing pressure eastward). Concurrent winds over the Northern California coast are now southwesterly, generally strong ($\text{max} > 30\text{ m s}^{-1}$), and follow the coastline in offshore areas. From Northern California to Baja, they are onshore (along the pressure gradient) and reach a maximum speed of around 12 m s^{-1} over the SFBA and Southern California Bight, where the thermal contrast is greatest ($> 20^\circ\text{C}$) (Fig. 22). Temperatures also show the cooling effect of the California coastal waters, which produce a strong west-east temperature gradient.

Twelve hours later (1200 UTC or 0400 PST), when surface heating no longer prevents its formation, a low pressure center (of 1014 hPa) with weak cyclonic winds is located over Northern California (Fig. 23), eastward from its upper level position. Concurrent land-ocean temperature differences are greatly reduced (Fig. 24), but not reversed, with respect to daytime values, and thus onshore wind components are weaker over the SFBA and Southern California, where inland temperatures are still high.

On the start of the episode day (12 July at 0000 UTC), the MSLP field (Fig. 25) over the ocean from British Columbia to Baja California differs from the one 24 h before

because the higher values have strengthened and now show a north-south isobar orientation. Though the upper-level trough is decaying, its cyclonic circulation is still reflected at the surface over Northern California. Concurrent air temperatures over the SFBA are cooler by a few degrees (Fig. 26) relative to those of the previous afternoon (as MS lasted longer on this day). As the upper level trough weakened further until 1200 UTC on July 12 (not shown), its influence at the surface was less evident than 24 h earlier, e.g., Northern California winds are weaker (if not calm), a broad horizontal pressure gradient has replaced the low, and the Pacific High has moved eastward.

During the next 12 h (by 13 July at 0000 UTC) the trough disappears, resulting in stronger horizontal pressure gradients across California than on the previous days (Fig. 27). The following night (13 July at 1200 UTC or 0400 PST), only the Pacific high and the Gulf of California thermal low are present, and onshore winds are weaker (relative to previous nights). The 13th is also characterized by general temperature increases during both the nighttime and daytime periods, due to eastward movement of the Pacific high, whose subsidence compresses and thus warms the atmosphere. The upper level trough, thus develops during the first days (10-11 July), weakens and disappears during the next day (the episode day), finally leading to more typical summer conditions on 13 July.

The above pressure and wind patterns are also evident in concurrent Oakland rawinsonde soundings. The sounding on 11 July at 0000 UTC (Fig. 28) shows the tropopause at about 250 hPa, with temperatures below this level increasing towards the surface to a maximum value of 29°C at the top (900 hPa) of the near-surface elevated subsidence inversion. Within the inversion, temperatures increased by about 15°C from its base at

950 hPa (800 m) to its top near 900 hPa (1200 m). As the rawinsonde site is adjacent to SFB and as near-surface winds are westerly, the inversion caps an adiabatic MABL above a shallow superadiabatic surface boundary layer (SBL). Dew points show a relatively dry upper troposphere, three shallow moist layers between 700 and 900 hPa, and a shallow MS layer in the upper part of the MABL. Speeds were generally low, with a weak jet (60 kt or 30 m s^{-1}) at 200 hPa.

Twelve hours later, on 11 July at 1200 UTC (0400 PST), the upper level trough was deeper (Fig. 29) and the tropopause lower (at 275 hPa), below which level temperatures again increased with decreasing height. The maximum was again at the subsidence inversion top (same level as 12 h earlier), but the troposphere had cooled, e.g., the inversion top was at 27°C . The deepening trough, radiative cooling induced turbulence, and weaker nighttime near-surface divergence, however, raised and cooled the inversion base, resulting in a more intense and slightly elevated layer. MABL temperatures were about 3°C cooler than 12 h before, although the only stability change was elimination of the super-adiabatic SBL. Dew points indicated a drier atmosphere above the still MS capped MABL. Above 500 hPa, winds were faster (as though had deepened) and still southerly, but below that level winds were weaker and directions more variable with height. They turned from southeasterly at 850 hPa to southwesterly at the surface, now influenced less by the weaker sea breeze and more by the trough.

By the start of the episode day on the following afternoon (12 July at 0000 UTC), the trough weakened, tropopause lifted, and temperatures showed only minor variations from the previous sounding (Fig. 30). While the inversion top is still at same height, its

temperature is about 3°C colder, and while the surface value was about the same, MABL-top temperatures were about 3°C colder than 24 h earlier due to the tropopause lifting and increased cloud cover. The atmosphere also continued to dry, producing lower dew points above 900 hPa, below which a saturated layer still existed. Winds were slower than in the two previous soundings, but were more south-southwesterly above 600 hPa and showed veering with height, indicative of warm air advection. Between this level and the MABL top, the southeasterly flow of the previous sounding was replaced by a more southerly flow, while the MABL continued to show southwesterly sea breeze flow.

During the following 24 h, the trough continued to weaken (12 July at 1200 UTC morning rawinsonde not available), so that at the episode end (13 July at 0000 UTC), the tropopause was higher and warmer than 24 h before and the troposphere warmer (Fig. 31). Increased subsidence (due to weakening trough) lowered the inversion base by almost 100 hPa to, bringing it near the surface. Dew points showed a moister atmosphere above 900 hPa and a nearly saturated MABL, except within the superadiabatic SBL. Winds were stronger than on the previous day, more southeasterly, and showed warm air advection below 500 hPa. Surface speeds were also stronger and westerly, again indicative of sea breeze flow.

During the following night (13 July at 1200 UTC or 0400 PST), the tropopause rose, temperatures above the MABL remained constant, but the MABL cooled by 7°C, sharpening the inversion (Fig. 32). Dew points above 450 hPa did not change from the previous 12 h, but values below that level were drier to the MABL-top, lower (at 950 hPa or 600 m) than on the previous days. The top part of the MABL was saturated with MS,

as was the near-surface layer with fog. While the jet was stronger than 24 h previous, speeds everywhere else were lower (especially below 700 hPa) and more barotropic.

c. Marine Stratus Event

On 11 July at 0000 UTC (10 July at 1600 PST), MS covered the coastal plain from Cape Mendocino to Big Sur (Fig. 33), except for the wind shadow zone (given the northwesterly flow) near Santa Cruz. While usually close to the shoreline, it penetrated inland at a few locations, with a maximum north of Point Reyes of 20 km. It was not of uniform thickness, however, as shown by spatial variations in its cloud-top brightness, which also indicate coherent embedded circulations. West of 125 W, the higher SSTs turned the MS turned into stratocumulus.

MS was advected inland by combined sea breeze and upslope topographic flows, helped by onshore flow along the southern edge of the weak offshore cyclone. By 11 July at 0130 UTC (10 July at 1730 LST), a MS tongue entered the Salinas Valley (Fig. 34), and thunderstorms developed on the western Sierra slopes. At 0200 UTC (not shown), uniform MS existed north of Point Arena, with patchiness west of 125 W and south of Monterey Bay.

By 0230 UTC (1 h before sunset), visible satellite images were no longer available, and IR images do not clearly show MS. However MS continued to advect inland throughout the night of the 10th-11th and deepen by radiative cooling, but was not replaced by new offshore cloud elements. It thus likely reached its maximum inland extent somewhat after sunrise at 1257 UTC (0457 LST on 11th), when solar heating started to dissipate it directly through absorption and indirectly through turbulent mixing.

The first new visible image (Fig. 35) at 1500 UTC at 0700 LST (2 h after sunrise on the 11th) showed MS at the Coastal and Diablo mountain ranges, thus a slight reduction of its maximum nighttime inland extent. MS covered the entire coast, except for the southern SFBA, while clear inland areas southeast of Big Sur were due to the Santa Lucia range, whose peaks penetrate above the MABL. Note that the southern MS boundary (dividing it from stratocumulus to the south) does not reach to the coast.

Morning turbulence from surface warming on the 11th and downslope southwest winds dissipated MS from its inland edge westward (Figs. 35 and 36), and thus SFO cleared at 1742 UTC or 0942 LST (not shown). The MS-stratocumulus dividing line again started to advect in-land (southeastward) by the late afternoon sea breeze, passing the coast north of San Francisco by 2100 UTC (not shown), where it dissipated over the land surface (Fig. 37). As the northern edge of the dividing line remained fixed, new MS formed, while the more southerly stratocumulus advected towards land. MS kept withdrawing until 2300 UTC (Fig. 37) and thus off-shore and coastal areas north of Cape Mendocino were clear and/or covered by thin MS during the afternoon.

During the late afternoon (Fig. 38) and night of the episode (again not shown), the now colder atmosphere (due to the offshore upper level trough) allowed for more condensation. Therefore by 12 July at 1500 UTC or 0700 LST (Fig. 39), a thicker MS cloud had penetrated all way through the Sacramento River Delta into the CV to the Sierras foothills (triangular shaped cloud in Fig. 39). The coast south of Big Sur and the Salinas Valley were, however, clearer than on the previous day. The thicker MS thus dissipated back towards the coast more slowly on this day than on the day before (compare Figs. 36 and

40). MS still was continuous over the entire SFBA at his time, and SFO clearing time was delayed until 2053 UTC (1253 LST). By 2100 UTC or 1300 LST (Fig. 41), the MS stopped its withdrawal, was on the coast from Cape Mendocino to Half Moon Bay, and was only slightly offshore to the south.

During the afternoon and nighttime hours of the 12th, the MS again moved inland, while northerly winds advected clear and patchy areas southward. Its maximum inland extent at 1500 UTC on the 13th (Fig. 42) was, however, the smallest of the three days, since a shallow MABL allowed topography to confine it to a small horizontal area. It thus barely advanced inland between Cape Mendocino and Point Reyes, while the Santa Cruz and Santa Lucia Mountains generated wide clear spots. It did, however, fill the Salinas and Santa Clara Valleys, as well as the SFBA, where a narrow spike pointed towards the CV through the Delta. It was continuous offshore, but with a long band of clear skies between the coast and 125 W.

The MS steadily withdrew westward during the morning of the 13th, with SFO clearing at 1804 UTC (1004 LST). Mixing due to strong winds offshore enlarged the clear area southward and the MS organized into longitudinal lines. Clearing continued, and by the afternoon (2300 UTC), it was present only west of 125 W and along the coast south of San Francisco (Fig. 43).

A SFO sodar sounding provided sporadic estimates (to within 15 to 65 m) of inversion base height (or MS top) while a nearby ceilometers measured over-head cloud-base height. On 11 July (Fig. 44), bases remained around 760 m from 0600 to 2200 UTC and then slowly rose as the upper level trough deepened. The MS was thin (65 m) and

dissipated around 1800 UTC, 20 min later than the official clearing time. The disagreement results from instrument uncertainty and because ceilometers can only detect overhead clouds, while official clearing times concern conditions in the SFO approach zone.

The 12 July episode-day soundings (Fig. 45) showed decreasing cloud ceilings (450 down to 400 m) in the 6 h from 0600 UTC, as radiative cooling thickened the MS by an 300 m relative to previous day values. Post 1200 UTC (cloud-base) ceilings remained constant until 1730 UTC, when inversion base heights (or cloud tops) oscillated between 600 and 750 m as MABL advection produced turbulent eddies. Ceilings rose after 1730 UTC (as MS thinned due to solar radiation) and dissipated around 2030 UTC (again earlier than the official SFO clearing time of 2054 UTC). Inversion base heights remained constant until the end of the data record at 13 July 0000 UTC.

Renewed data on 13 July at 0600 UTC (Fig. 46) showed ceilings decreasing from 270 to 100 m in 3 h, thus becoming lower than on the two previous days. The MS base remained constant until 1400 UTC, when it began to again rise. It then remained constant at 450 m from 0600 to 1800 UTC, after which it decreased slightly, as surface pressure rose (mentioned above) and divergence decreased slightly. After 2300 UTC (1500 UTC), it rose again when solar radiation strengthened and the MS cleared at 1800 UTC.

In summary, the upper level trough with its associated vertical motion, first generated lifting necessary to raise the 11 July MABL above many topographic features, which resulted in a wide inland MS coverage all night, during which the sea breeze flow did not reverse. Adiabatic cooling from the trough on the 12th increased condensation, and thus produced thicker (than the day before) MS that could pass over the same topographic fea-

tures as on the previous day. It was, however, thick enough to penetrate into the CV and to more slowly dissipate, producing a late SFO MS burn off. When the trough dissipated on 13 July, the MABL thinned and MS was limited to a small inland area, and thus SFO again cleared early.

4. LAPS ANALYSES

The 850 hPa LAPS afternoon temperature analysis on 11 July at 0000 UTC (10 July at 1600 LST) (Fig. 47) shows a 4°C warm-land cool-ocean coastal thermal contrast (offshore isotherms roughly parallel the coast). It also shows an inland-directed sea breeze cold air push (isotherms bend over SFBA) into the warm CV. The nighttime warm pool over the northern CV at 0300 UTC (not shown) moved to its southern end by 0600 UTC (Fig. 48). The pre-sunrise coastal-inland contrast was only 2°C at 1500 UTC (Fig. 49), as the CV warm pool had been replaced by uniform temperatures.

The expected daytime CV warm pool was not seen at 11 July on 2100 UTC or 1300 LST (Fig. 50) due to cooling from the approaching upper-level trough. By the start of the episode period on 12 July at 0100 UTC (Fig. 51), it had still not formed, and the coastal gradient region, narrower than on the previous day, is confined to land areas. Five hours later (Fig. 52), the trough is gone and the offshore gradient is thus about the same strength as during the previous night. Effects from the trough still exist, however, as temperatures over most of the domain are still lower than on the previous night. By 1500 UTC or 0700 LST (when GOES10 visible images show maximum inland MS penetration), the trough has mostly dissipated and thus temperatures are the same as 24 h earlier (not shown). During the next 12 h, the last remnants of the trough faded, warming occurred at 850 hPa over land and ocean (not shown) and the MS burned back.

The 950m hPa LAPS analysis on 11 July at 0000 UTC (10 July at 1600 LST) was within the MABL (Fig. 53), where it remained for 48 h (Figs. 28-30). Sea breeze advection of offshore cold air inland into the Salinas, Santa Clara, Central, and (to a les-

ser extent) Santa Rosa Valleys thus resulted. Note that high topography peaks intersected (blank areas) the 950 hPa surface and that temperature maxima existed on southwest facing slopes. The flow was strongest into the northern half of the CV, while penetration into its southern half was blocked by southerly flow. Post sunset inland temperatures decreased, while offshore values remained relatively constant, so that by 11 July at 1200 UTC (0400 LST) (Fig. 54) the coastal contrast was greatly reduced.

During the following 12 h, while over-land air warmed after sunrise, the approaching offshore upper-level trough produced a cooling of 4°C. By the start of the episode on 12 July at 0000 UTC (1600 LST on July 11), an afternoon offshore cool pool thus formed (Fig. 55). During the next 12 h, nighttime radiative cooling was reduced by warming from the now weakening upper level trough. Thus by 1200 UTC (Fig. 56), early morning temperature values were generally similar to those of 24 h earlier, except in the CV, where values were slightly warmer and where a more normal west to east gradient had reformed. As the upper-level trough continued to fill after sunrise on the event day, solar heating was not countered by its induced cooling, and thus temperatures were higher than on the previous day. Inland and offshore temperatures on the 13th (not shown) were higher than on the previous two daytime and nighttime periods, due directly to continued filling of the trough and indirectly (in offshore regions) to MABL lowering, which brought the 950 hPa surface into the inversion.

Afternoon surface (i.e., shelter-level) temperatures on 11 July at 0000 UTC (10 July at 1600 LST) were (as expected) warmer than corresponding 950 hPa values, showed a stronger coastal contrast, and thus produced a stronger inland sea breeze push (Fig.

57). At 0300 UTC (Fig. 58), the early evening push is even more evident over the Sacramento River Delta, Salinas Valley, and Santa Clara Valley, while southwest facing slopes of the Santa Cruz, Diablo, and Coastal ranges were again warmer. As the evening proceeded, over land values decreased, while offshore values generally slowly increased, partly due to latent heat release associated with MS formation. A more important factor is the upper-level trough, which with its tilt to the west with height, would appear first at the surface. Its associated cyclonic circulation advected warm air eastward, so that by 1200 UTC (0400 LST) (Fig 59), isotherms were perpendicular to the coast (with elimination of coastal gradients). This evolution was such that the MS south of Point Reyes at 0130 UTC (Fig. 34) turned into stratocumulus by 1500 UTC (Fig. 35).

After sunrise on the 11th, offshore temperatures remained constant, but the pattern changed due to solar heating of the land; and thus by 1800 UTC (1000 LST) cold air had again penetrated into the CV through the Delta and Pacheco Pass (Fig. 60). Daytime inland temperatures increased during the next 6 h (not shown) to values slightly warmer offshore (but still cooler inland) as compared to those 24 h previous, but nighttime values (Fig. 61) again decreased inland and increased off shore (again due to warm air advection). Two hours after sunrise on the 12th at 1500 UTC (Fig. 62), isotherm orientation in the northwest part of the domain parallels visible cloud cover, with colder values over clear areas. Over the next 9 h daytime hours (Fig. 63), inland areas generally warmed and offshore areas cooled slightly, so that the afternoon coastal contrast became larger than that 24 h earlier. The 13 July surface temperature evolution was analogous to the previous day, but with colder offshore and warmer inland values (not shown).

The 950 hPa LAPS winds for 11 July at 0000 UTC (Fig. 53) show the approaching upper level trough in the offshore southeasterly winds, normally northerly under Pacific High influence. Afternoon surface thermal heating effects are also visible in the mesoscale cold air sea breeze advection into the northern CV and channeling along the Santa Clara Valley. During the later afternoon, evening, and night of the 11th (Fig. 54), mesoscale forcings weakened. The trough also strengthened slightly, so that it extended its influence southward and its cyclonic circulation could advect air from over the warm water west of the domain eastward towards the SFBA.

The area with 950 hPa southerly winds thus shrunk, and 12 h later at the start of the episode on 12 July at 0000 UTC, typical summertime large scale northwesterly winds appeared over the ocean (Figs. 54 and 55). After sunrise on the 12th, the increase of surface heating further weakened the trough, and northerly 950 hPa winds replaced the slower southerly flow over most of the domain (not shown). During this daytime period, winds over the SFB were west-southwesterly and advected MS inland, while over the southern CV they oscillated between southerly and southeasterly. The oscillation resulted from the trough (which tended to switch the geostrophic northerly flow to southerly), as this area is less sensitive to coastal contrast because of blocking by its eastern-boundary Santa Lucia Mountains.

During the remaining part of the night (12th to 13th) thermal contrasts weakened further, and 2 h after sun rise (1500 UTC) cyclonic flow was present again, probably due to surface thermal heating. As the event period ended, cyclonic flow slowly vanished and was replaced by a normal northwesterly flow with an imposed sea breeze.

Surface winds on 11 July at 0000 UTC or 10 July at 1600 LST (Fig. 57) show an offshore cyclone with northerly winds off Point Reyes, southwesterly flow well offshore of the SFBA, weak upslope flow over the Coastal Range north of Point Reyes, and a developed sea breeze into the CV (through the SFBA), Santa Clara, and Salinas Valleys. The next 6 h (Fig. 58) were characterized by the oscillating vortex movement and by its weakening (as solar heating weakened). By 11 July at 1200 UTC (Fig. 59), the weaker night-time mesoscale land ocean contrast allowed the upper level trough to produce southwesterly winds in the northwest domain corner. A convergence zone over the Santa Clara Valley resulted from the breeze from the SFB and the downslope winds from the Santa Cruz Mountains. The sea breeze still reached into the CV and split against the western side of the Sierras.

Southwesterly winds extended southward, reaching Point Reyes during the morning of the 11th (Fig. 60), while a strong afternoon sea breeze penetrated into the CV through both the SFB and Pacheco Pass. During the afternoon and the night, the trough weakened, and by 12 July at 0900 UTC (Fig. 61) no longer affected surface winds. The nighttime thermal contrast was weaker than the daytime value, and thus the sea breeze did not extend as far off shore (Figs. 57 and 61). As the night progressed, the sea breeze continued to weaken, and after sun rise on the episode day (12th) it reached a minimum intensity (Fig. 62).

After 1500 UTC on the 12th, winds were driven by solar heating and a new sea breeze (not shown) formed over the CV, Santa Clara, and Salinas Valleys. By 13 July at 0000 UTC (Fig. 63), a cyclonic circulation formed north of Point Reyes (as on the 11th),

and the remaining study hours were characterized by a strengthened geostrophic circulation on which the sea breeze was superimposed.

LAPS mean sea level pressures (MSLPs) on 11 July at 0000 UTC (10 July at 1600 LST) generally showed isobars parallel to the coast (Fig. 64) due to daytime heating, while the approaching trough produced isobar bending in the southwestern part of the domain. During the night, MSLP values increased over land, but decrease offshore due to the strengthening trough, and by 1200 UTC (Fig. 65) the nighttime coastal gradient had relaxed and offshore isobars tended towards an east-west orientation.

After sunrise (1258 UTC) on the 12th (not shown) trough influence decreased and isobars again parallel the coast, as pressure values increased over the ocean with respect to the previous day. At the same time, trough induced winds started to appear. Afternoon MSLPs on 12 July at 0000 UTC (Fig. 66) were generally thus 1-2 hPa larger than 24 h earlier, but the coastal gradient was generally unchanged, consistent with the ETA analyses. As on the previous day, nighttime pressures were higher than daytime values and the coastal gradient reduced (Fig. 67). The weakening trough produced nighttime values again higher than 24 h before, and for the next 12 h values increased and no trough was visible on 13 July at 0000 UTC (Fig. 68). The last day of the period showed only a typical diurnal cycle, with generally higher pressure values.

Balanced LAPS fields were sometimes unrealistic, e.g., its 1000 hPa winds were too little influenced by SFBA topography (Fig. 69). In addition, its 950 hPa temperatures were overly smooth (Fig. 70), as they showed only coastal contrasts, while unbalanced analyses discussed above provided additional details. Results depend on how strongly

the variational algorithm is set to force geostrophic balance. In the current case, radar data were not utilized, and nighttime observations were frequently concentrated around the SFB, rare further inland, and almost absent over the ocean. The shortest wave length resolvable by the data thus cannot resolve most mesoscale features. On the other hand, unbalanced analyses performed well as they give more weight to observations.

Disagreement exists as to which atmospheric moisture parameter(s) best determines cloud cover extent. Evaluation of LAPS cloud fields against satellite images requires threshold values to determine cloud location, and thus four output parameters were chosen (by trial and error) as MS proxies: LWC, FCC, unbalanced RH, and balanced RH.

A reasonable LWC for California coastal MS is 0.2 g kg^{-1} (Goodman, 2001, personal communication), a value larger than in LAPS. A value of 0.1 g kg^{-1} produced better coverage (not shown), but was often discontinuous in time and only present over data rich areas (i.e., San Francisco and Monterey Bay). A threshold of 0.02 g kg^{-1} showed more MS details (Fig. 71) and produced realistic representations over land areas only during the 12 h before the episode peak coverage (at 12 July at 1500 UTC). The LAPS underestimated MS coverage before and after this period is due to ETA analyses, which provided too-dry first-guess RH fields (e.g., Figs. 72 and 33). As the ETA 95% RH contour is a few grid points (80 km resolution) from the coast, LAPS "sees" the MS eastern edge several hundred kilometers off the coast.

The LAPS FCC threshold that best represented the observed widespread MS coverage during the first two days of the period is 0.01 at 950 hPa (Figs. 28-30). LAPS 1000 hPa (first grid below 950 hPa) clouds were always within the higher level coverage.

Afternoon FCC values on 11 July at 0000 UTC (10 July at 1600 LST) (Fig. 73) thus showed good agreement with the corresponding satellite image (Fig. 33), except that the clear coastal areas north of Point Reyes and south of Monterey Bay should be cloudy. Three hours later (not shown), the MS had grown slightly, consistent with the last available visible satellite image at 0230 UTC.

The nighttime LAPS MS progressed inland until 11 July at 0900 UTC (not shown), when it reached the CV north and south of the SFBA. It then withdrew, cleared over SFO, and again advanced inland. At 1500 UTC (Fig. 74), its extension was generally in good agreement with satellite observation (Fig. 35), except for a slight over extension in the western CV. The morning withdrawal on the 11th was reproduced, but MS was overestimated north of SFBA and underestimated to its south. Even with the minimum FCC MS inland extension at 2100 UTC (not shown) SFO never cleared, but satellite images continuously showed afternoon MS close by (e.g., Fig. 37). The concurrent afternoon spatial non-homogeneity in the stratocumulus in the data sparse area southwest of San Francisco was not captured.

LAPS MS started to progress inland at 2200 UTC (1400 LST) on the 11th, and the MS field one hour later (Fig. 75) was again in general agreement with observations. While SFBA cloudiness ($FCC > 0.01$) was overestimated, the area had a steep FCC gradient and scattered cloudiness (Fig. 37). As the afternoon progressed, LAPS MS continued to advance inland and intensify, and thus at the start of the episode on 12 July at 0100 (11 July at 1700 LST) it well represented the observed MS (not shown).

During the rest of the afternoon and night of the 11th, LAPS MS advanced inland, reaching the CV at 12 July at 0400 UTC (just past start of episode day), after which it slowly decrease in area. From 0900 UTC onwards, LAPS MS again grew, and by 1500 UTC had both correctly covered the entire coast and penetrated into the CV (Figs. 76 and 39). It overestimated, however, both the extent of the western CV coverage and the observed triangular cloud during the morning and early afternoon (until 2100 UTC) on the 12th, when MS was overestimated consistently over the CV, SFB, and (sometimes) Salinas Valley. During this period, SFO should have cleared but it did not, and thus a higher FCC threshold (0.15-0.20) performed better (not shown) in this area. From 12 July at 2200 UTC to 13 July at 0200 UTC (time of last satellite image), LAPS MS was over the SFBA, but observations only showed it along the coast or a few kilometers inland.

Between 0300 and 0600 UTC on the 13th, LAPS MS remained offshore and its 950 and 1000 hPa coverage remained similar. After 0700 UTC (consistent with Fig. 46), nighttime 950 hPa FCC values dropped and 1000 hPa values better matched the observed MS. LAPS MS advanced discontinuously inland during the rest of the night (as during the previous night) and reached its maximum inland extent at 1400 UTC (not shown). At 1500 UTC or 0700 LST (Fig. 77), MS was slightly underestimated along the coast and overestimated over the Sacramento River Delta, but by only a few grid points.

Its withdrawal was well reproduced (not shown) and at 1800 UTC it was still over the SFB as observed (not shown). Although observations showed that SFO cleared at 1804 UTC, the 1900 UTC LAPS analysis (not shown) showed SFO one grid point inside the MS edge. LAPS MS then withdrew offshore but returned inland at 2200 UTC, while

the observed MS stayed onshore and cleared from the north. The 2300 UTC analysis (Fig. 78) and the observations (Fig. 43) thus show dissimilar patterns over the ocean, probably due to data scarcity.

MS evolution was thus well described by FCC, but a fairly low threshold value was required. LAPS computed cloud ceilings and tops for all grid cells with both a FCC > 0.1 and a cloud vertical velocity $> 0.65 \text{ m s}^{-1}$, again because ETA analyses were too dry.

An unadjusted RH of 85% at 950 hPa (Fig. 79) on 11 July at 0000 UTC (10 July at 1600 LST) well represented the observed MS, then located slightly inland north of Monterey Bay and close to coast to its south, but the relatively poor LAPS resolution produced small errors over the San Francisco and Monterey Bays (Fig. 33). During the next 2 h, LAPS MS advanced inland more slowly and discontinuously than observed, and thus only reached the CV at 0700 UTC (not shown). By 1500 UTC (Fig. 80), except for its shape in the CV, LAPS MS closely resembled the observations (Fig. 35), as it is along the coast and within the valleys. MS withdrawal due to morning turbulence was well captured, and its initial overestimation of SFB coverage improved with time (Figs. 81 and 36). Grid resolution again caused the subsequent small underestimation in daytime coverage until the start of the episode day on 12 July at 0200 UTC or 11 July at 1800 LST (not shown).

During the late evening and night, LAPS MS advanced inland and reached the CV at 12 July at 0400 UTC, after which it advanced only irregularly. Its maximum extent over the CV was at 1300 UTC, the analysis closest to sunrise. At 1500 UTC or 0700

LST (Fig. 82), the MS was well represented, but it extended too far along the CV western edge. MS withdrawal over the next 6 h was again well described (Figs. 83 and 39), with a slight overestimation over the SFB and small coastal over- and underestimations. The MS edge was again over SFO (only a few grid points away), while the satellite again showed SFO just outside its edge.

After 2200 UTC on 12 July, afternoon 950 hPa RH values dropped over the entire domain and then 1000 hPa values best represented MS extent (not shown). The drop occurred a few hours before that shown by the FCC values discussed above, but sodar and ceilometer observations cannot confirm either time, as SFO was then clear (Fig. 45). From 12 July at 2200 UTC to 13 July at 0200 UTC (12 July at 1800 LST), LAPS well approximated MS coverage over the entire domain (not shown), with a slight tendency to underestimation especially over Monterey Bay. After that, 950 hPa values best represented the MS, which started to climb the steep topography east of Point Reyes.

During the late evening and night of the 12th, LAPS MS slowly (but continuously) advanced inland and reached its maximum CV extent at 1300 UTC on the 13th (not shown), again the analysis closest to sun rise (1259 UTC). From 1500 UTC (0700 LST) to the end period (not shown), LAPS again well represented the observed MS withdrawal with a tendency for underestimation, and thus a lower RH of 80% performed slightly better (not shown). It did not, however, again (as with FCC) capture the clearing offshore.

Again ETA analyses caused the general MS underestimation by unbalanced RH values. Proof of this is in data-rich areas (e.g., SFBA), where the same threshold resulted in MS overestimation.

Balanced RH produced a more complicated vertical structure, and thus both the 950 and 1000 hPa levels were necessary to obtain good MS simulation at the chosen threshold of 75%. On 11 July at 0000 UTC (10 July at 1600 LST), LAPS approximated the continuous afternoon MS along the coast fairly well (Fig. 84). Monterey Bay was correctly partially clear (Fig. 33), but LAPS produced too much MS over the SFB and a too clear offshore area between Cape Mendocino and Point Reyes.

During the next 2 h as clear areas filled (not shown), LAPS better matched the observations. During the evening and night, LAPS MS advanced inland, reached the CV at 1200 UTC on the 11 July, and then withdrew. Lack of nighttime data during the next 3 h (not shown) caused 950 hPa RH values to change abruptly from a wide MS coverage to a clear domain and then back again. The 1500 UTC field only slightly overestimated MS over the western CV and did not fill the entire Salinas Valley (Figs. 85 and 35). The following 3 h (Fig. 86) were characterized by a too fast morning burn-off over Monterey Bay and by uniform MS well offshore of the Bay, where observations (Fig. 36) showed stratocumulus. From 1900 UTC, MS remains along the coast as observed, but was underestimated offshore until the start of the episode day on 12 July at 0000 UTC, when its horizontal extent was close to observed (but less uniform) over the entire domain.

During the next couple of hours, the real MS advanced inland, but no nighttime satellite observations exist. By 0400 UTC on 12 July or 11 July at 2000 LST (not shown) a 950 hPa MS tongue penetrated into the CV, while 1000 hPa MS filled most valleys. MS then advanced less continuously and at 1500 UTC again well represented the real MS

everywhere (Figs. 87 and 39), except for a consistent overestimation in the CV until 1900 UTC (not shown) that produced a too slow burn off.

Throughout the same 5 h period, the LAPS MS edge showed a steep gradient of balanced RH values. All cloudy grid points also had $RH > 99\%$, as after the wind was balanced the RH was balanced, as temperature changes are introduced via new geopotential height values. All grid points with upward vertical velocities were also assumed saturated. The 2000 UTC analysis better matched the observation (not shown). During the next 3 h, 950 hPa balanced RHs dropped (not shown), clearing the domain (consistent with the unbalanced RHs). MS was underestimated over the ocean north of SFB, but well estimated along the coast. Balanced 950 hPa LAPS RH values rose from 13 July at 0200 UTC, producing MS. This difference from the unbalanced RH results is due to the post-balance saturation mentioned above.

The remaining afternoon and nighttime hours (until 2 h after sunrise at 1500 UTC on the 13th) were characterized by a slow continuous inland movement of LAPS MS at 1000 hPa (not shown). The concurrent 950 hPa MS was more discontinuous, first slowly disappearing from east to west, reforming at 1000 UTC, and then again disappearing. At 1500 UTC on 13 July, LAPS well approximated the MS (Fig. 88). The observed continuous southeastward MS movement at 2300 UTC was, however, more discontinuous than the LAPS results, which as usual overestimated coverage (Figs. 89 and 43).

Balanced RH thus also underestimated MS, as the realistic threshold of 95% did not provide good estimates, again due to poor ETA analyses. It did, however, perform better than unbalanced RH, as it produced MS fields more consistent with the wind fields

(e.g., offshore clearing during the afternoon of the 13th). Post-balanced saturation allowed divergence and temperature fields to alter RH values. This produced better MS values, although unbalanced winds captured mesoscale details not in balanced analyses.

5. CONCLUSION

Previous efforts to forecast MS over SFO have used statistical or semi-empirical approaches, while the current study has sought to understand its summertime dynamics by a series of mesoscale analyses. The Local Analysis and Prediction System (LAPS) has previously successfully analyzed cloud events and initialized mesoscale meteorological models. It ingests virtually any type of meteorological data, and because of its variational algorithms, its relatively simple cloud and soil models can produce physically consistent analyses for a wide variety of meteorological parameters.

The current study thus compared LAPS results (with a 3.3 km horizontal grid) from an analysis of a summertime MS event over SFO against 1 km resolution visible satellite images, which were withheld from the LAPS analyses. The LAPS default configuration for Colorado was modified to account for a strong ageostrophic flow during the current MS event. Two variational surface analysis parameters were thus increased (within suggested ranges) to allow for larger pressure field modifications and momentum equation residuals. The lower SFBA topographic elevations were also accounted for, as the check for superadiabatic layers was carried out between the surface and 700 hPa (versus the default 500 hPa), and the pressure reduction level was lowered from 1500 to 500 m. The study incorporated the following data into LAPS: RAOB, NOAA 915 Hz profiler, ACARS, METAR, mesonets (BAMI included), buoy, and ship. Six and 12 h ETA forecasts (issued every 12 h) were used as first guesses and linearly interpolated to the hour (LAPS analysis time increment).

Criteria to select the study event included extensive nighttime MS coverage over the SFBA and a "difficult to forecast" late MS burn off. The event also had to fall between days with more "typical" burn off times, so that comparisons between the three days would increase understanding of MS phenomena. The 24 h period from 0000 UTC on July 12 (1600 LST on July 11) was thus selected as the event day. A detailed synoptic analysis showed that an upper level trough, with its associated vertical motion, first generated the lifting necessary to raise the 11 July MABL above many SFBA topographic features. This resulted in wide inland MS coverage all night, during which the sea breeze flow did not reverse. Adiabatic cooling from the trough on the 12th increased condensation, and thus produced a thicker (than the day before) MS layer that again passed over the topographic features, penetrated into the CV, more slowly dissipated, thus producing a late SFO MS burn off. When the trough dissipated on 13 July, the MABL thinned, MS was limited to a small inland area, and thus SFO again cleared early.

As no universal definition of LAPS clouds exist, four of its output parameters were used as cloud proxies to evaluate its cloud analysis: LWC, FCC, unbalanced RH, and balanced RH. For each parameter, a threshold was defined, and grid points with values above the threshold were considered "cloudy."

ETA analyses first-guess fields for LAPS were too dry, however, to produce reasonable cloud analyses. With physically meaningful LAPS thresholds (e.g., $RH > 95\%$ or $LWC > 0.2 \text{ g kg}^{-1}$), MS fields were systematically underestimated, especially in data poor regions. Lower thresholds produced more reasonable cloud analyses for all parameters, except LWC which depends strongly on initial RH values and which thus led to overest-

imated MS in data rich areas. A $FCC > 0.01$ reproduced MS extents well, and within the limits of the 3.3 km LAPS resolution, correctly analyzed its edges. Unbalanced RH at a reasonable threshold of 85% also performed fairly well, as it properly located MS edges and consistently captured observed MS lowerings. Balanced RH produced further improvements, as it produced more complex vertical structures, was consistent with wind fields, and showed better agreement with observations in data poor regions.

The three parameters located MS layers at observed vertical levels, including areas above many SFBA topographic features. The three also correctly lowered the MABL between the late afternoon and night of the 12th, even though their timing was not always consistent.

Recommendations for future work concern LAPS grid resolution and first guess meteorological fields. Although its cloud analyses compared well against observations, grid resolution is critical, as topographic details are fundamental for accurate analyses and for improved simulation of MS movements over SFO. At a 3.3 km horizontal grid resolution, the San Francisco peninsula is only a few grid points wide, and its local topography is characterized by even smaller features that can block MS movements to SFO. Vertical grid resolution must also be fine enough to resolve the small hills typical of the SFBA. LAPS analyses at 1 and 10 km resolutions of the same MS event (not reported herein) showed poorer performance than the 3.3 km resolution results, e.g., 10 km resolution analyses provided fields with too few mesoscale details. The 1 km resolution analyses were, however, strongly dependent on data availability, with missing data in key

regions greatly affecting predicted MS fields. The current SFBA data density thus produced best results with a 3.3 km resolution.

First guess RH fields also need to show better agreement with observation to provide LAPS with more accurate background fields for it to perform better in the absence of other data sources. A second series of LAPS analyses carried out as part of the current study (again not reported herein) withheld surface BAMI mesonet data from LAPS. These results should be further analyzed to evaluate the role of such data in LAPS SFO cloud analyses.

Visible satellite images (withheld from LAPS in the present study as validation data) can greatly improve LAPS cloud parameter analyses. Such data should thus be used in future LAPS studies of SFO MS to improve its analyses of cloud parameters (such as LWC or cloud type) that cannot be objectively determined from satellite images alone. Balanced LAPS cloud analyses that use satellite data can provide useful initial fields for mesoscale models (e.g., RAMS and MM5) that could provide more accurate SFO MS forecasts. They can study the physical processes (discussed above, e.g., radiative and turbulent flux divergences) related to MS formation, advection, and dissipation. Improved MS forecasts would reduce weather related delays and financial losses at this challenging site.

Appendix: List of Acronyms

ACARS	Aircraft Communications Addressing and Reporting System
AVN	Aviation Global Model
BAMI	Bay Area Mesoscale Initiative
CCN	Cloud Condensation Nuclei
CV	Central Valley
ETA	Regional atmospheric model of National Center for Environmental Prediction
FAA	Federal Aviation Administration
FCC	Fractional Cloud Cover
FDDA	Four Dimensional Data Assimilation
FSL	Forecast System Laboratory
GPS	Global Positioning System
LAPS	Local Analysis and Prediction System
LDAD	Local Data Acquisition and Dissemination
LES	Large Eddy Simulation
LLJ	Low Level Jet
LST	Local Standard Time
LWC	Liquid Water Content
MABL	Marine Atmospheric Boundary Layer
METAR	Aviation Routine Weather Report
MM5	Fifth generation mesoscale model of NCAR
MS	Marine Stratus
MSL	Mean Sea Level
MSLP	Mean Sea Level Pressure
NOAA	National Oceanic and Atmospheric Administration
NPS	Naval Postgraduate School
NWS	National Weather Service

PBL	Planetary Boundary Layer
PST	Pacific Standard Time
PIREP	Pilot voice report
QC	Quality Check
RAMS	Regional Atmospheric Modeling System
RAOB	Rawinsonde observation
RASS	Radio Acoustic Sounding System
RH	Relative Humidity
RUC	Rapid Updated Cycle
SBL	Surface Boundary Layer
SFB	San Francisco Bay
SFBA	San Francisco Bay Area
SFO	San Francisco Airport meteorological station
SODAR	SOund Detection And Ranging
SJSU	San Jose State University
SST	Sea Surface Temperature
TKE	Turbulent Kinetic Energy
UTC	Coordinated Universal Time
WFO	Weather Forecast Office

References

- Ackerman, A., O. Toon, and P. Hobbs, 1995: A model for particle microphysics, turbulent mixing, and radiative transfer in the stratocumulus topped marine boundary layer and comparisons with measurements. *J. Atmos. Sci.*, **52**, 1204-1236.
- Airports Council International, 2001: The world busiest airports. [Available on-line at <http://www.airports.org/traffic/passengers.html>]
- Alberoni P.P., S. Costa, A. Selvini, J. McGinley, and V. Levizzani, 1998: Implementation and test of a mesoscale data assimilation system for northern Italy. *MAP meeting*, Chamonix, France.
- Albers, S. C., 1995: The LAPS wind analysis. *Wea. Forecasting*, **10**, 342-352.
- Albrecht, B. A., 1989: Aerosols, cloud microphysics, and fractional cloudiness. *Science*, **245**, 1227-1230.
- Albrecht, B. A., 1991: Fractional cloudiness and cloud-top entrainment instability. *J. Atmos. Sci.*, **48**, 1519-1525.
- Albrecht, B. A., C. S. Bretherton, D. Johnson, W. H. Schubert, and A. S. Frisch, 1995: The Atlantic Stratocumulus Transition Experiment (ASTEX). *Bull. Amer. Meteor. Soc.*, **76**, 889-903.
- Backer, M. B., and R. J. Carlson, 1990: Bistability of CCN concentrations and thermodynamics in the cloud-topped boundary layer. *Nature*, **345**, 142-145.
- Barnes, S. L., 1964: A technique for maximizing details in numerical weather map analysis. *J. Appl. Meteor.*, **3**, 396-409.
- Bennett, D. A., K. Hutchison, S. C. Albers, and R. D. Bornstein, 2000: Preliminary results from polar-orbiting satellite data assimilation into LAPS with application to mesoscale modeling of the San Francisco Bay Area. *Preprints, 10th Conf. on Satellite Meteor. and Oceanography*, Long Beach, CA, Amer. Meteor. Soc., 118-21.
- Birkenheuer, D., 2001: Utilizing variational methods to incorporate a variety of satellite data in the LAPS moisture analysis. *Preprints, 11th Conf. on Satellite Meteorology and Oceanography*, Madison, WI, Amer. Meteor. Soc.
- Birkenheuer, D., B. Shaw, S. Albers, and E. Szoke, 2001: Evaluation of local-scale forecasts for Severe Weather of 20 July 2000. *14th Conf. On Numerical Weather Prediction*, Fort Lauderdale, FL, Amer. Meteor. Soc.

- Bond, N. A., C. F. Mass, and J. E. Overland, 1996: Coastally trapped wind reversals along the United States west coast during the warm season. Part I: Climatology and temporal evolution. *Mon. Wea. Rev.*, **124**, 430-445.
- Bretherton, C. S., M. K. MacVean, P. Bechtold, A. Chlond, W. R. Cotton, J. Cuxart, H. Cuijpers, M. Khairoutdinov, B. Kosovic, D. Lewellen, C.-H. Moeng, P. Siebesma, B. Stevens, D. E. Stevens, I. Sykes, and M. C. Wyant, 1999: An intercomparison of radiatively driven entrainment and turbulence in a smoke cloud, as simulated by different numerical models. *Quart. J. Roy. Meteor. Soc.*, **125**, 1-46.
- Brewster, K., F. Carr, N. Lin, J. Straka, and J. Krause, 1994: A local analysis system for initializing real-time convective-scale models. *Preprints, Tenth Conference on Numerical Weather Prediction*, Portland, OR, Amer. Meteor. Soc.
- Bridger, A. F. C., W. C. Brick, and P. F. Lester, 1993: The structure of the marine inversion layer off the Central California coast: Mesoscale conditions. *Mon. Wea. Rev.*, **121**, 335-351.
- Brost, R. A., D. H. Lenchow, and J. C. Wyngaard, 1982a: Marine stratocumulus layers. Part I: Mean conditions. *J. Atmos. Sci.*, **39**, 800-817.
- Brost, R. A., J. C. Wyngaard, and D. H. Lenchow, 1982b: Marine stratocumulus layers. Part II: Turbulence budgets. *J. Atmos. Sci.*, **39**, 818-836.
- Burk, S., and W. Thompson, 1996: The summertime low-level jet and marine boundary layer structure along the California coast. *Mon. Wea. Rev.*, **124**, 668-686.
- Burk, S. D., and T. Haack, 2000: The dynamics of wave clouds upwind of coastal orography. *Mon. Wea. Rev.*, **128**, 1438-1455.
- Chen, C., and W. R. Cotton, 1987: The physics of the marine stratocumulus-capped mixed layer. *J. Atmos. Sci.*, **44**, 2951-2977.
- Clark, D. A., and F. W. Wilson, 1996: The marine stratus initiative at San Francisco International Airport. Project Report ATC-252, MIT Lincoln Laboratory, Lexington, MA.
- Cotton, W., and R. Anthes, 1989: *Storm and Cloud Dynamics*, Academic Press, 883 pp.
- Cui, Z., M. Tiernstroom, and B. Krishnamurti, 1998: Idealized simulations of atmospheric coastal flow along the central coast of California. *J. Appl. Meteor.*, **37**, 1332-1363.

- Deardoff, J. W., 1980: Cloud top entrainment instability. *J. Atmos. Sci.*, **37**, 131-147.
- Dorman, C. E., and C. D. Winant, 1995: Buoy observations of the atmosphere along the west coast of the United States, 1981-1990. *J. Geophys. Res.*, **100**, 16,029-16044.
- Dorman, C. E., T. Holt, D. P. Rogers, and K. Edwards, 2000: Large-scale structure of the June-July 1996 marine boundary layer along California and Oregon. *Mon. Wea. Rev.*, **128**, 1632-1652.
- Durkee, P., K. Noone, R. J. Ferek, D. W. Johnson, J. P. Taylor, T. J. Garrett, P. Hobbs, J. G. Hudson, C. S. Bretherton, G. Innis, G. M. Frick, W. A. Hoppel, C. D. O'Dowd, L. Russell, R. Gasparovic, K. E. Nielsen, S. A. Tessmer, E. Öström, S. Osborne, R. C. Flagan, J. H. Seinfeld, and H. R., 2000: The impact of ship-produced aerosols on the microstructure and albedo of warm marine stratocumulus clouds: A test of MAST hypotheses Ii and Iii.. *J. Atmos. Sci.*, **57**, 2554-2569.
- Farley J., 2000: Statistical stratus forecasts for San Francisco International Airport. M. S. Thesis, Meteorology Department, San Jose State University, 98 pp.
- Feingold, G., W. Cotton, B. Stevens, and A. Frisch, 1996: The relationship between drop in-cloud residence time and drizzle production in numerically simulated stratocumulus clouds. *J. Atmos. Sci.*, **53**, 1108-1122.
- Fu, Q., M. C. Cribb, H. W. Barker, S. K. Krueger, and A. Grossman, 2000: Cloud geometry effects on atmospheric solar absorption. *J. Atmos. Sci.*, **57**, 1156-1168.
- Goodman, J. K., 1977: The microstructure of California coastal fog and stratus. *J. Appl. Meteor.*, **16**, 1057-1067.
- Goodman, J. K., and A. Miller, 1977: Mass transport across a temperature inversion. *J. Geophys. Res.*, **82**, 3436-3471.
- Haines, P.A., J. K. Luers, and C. A. Cerbus, 1989: The role of the Smith-Feddes model in improving the forecasting of aircraft icing. *Third Conference on Aviation Weather Systems*, Anaheim, California, Amer. Meteor. Soc, 258-263.
- Hayden, C. M., 1988: GOES-VAS simultaneous temperature-moisture retrieval algorithm. *J. Appl. Meteor.*, **27**, 705-733.
- Hilliker, J. L., and M. Fritsch, 1999: An observations-based statistical system for warm-season hourly probabilistic forecasts of low ceiling at the San Francisco international airport. *J. Appl. Meteor.*, **38**, 1692-1705.

- Hignett, P., 1991: Observation of the diurnal variation in a cloud-capped marine boundary layer. *J. Atmos. Sci.*, **48**, 1474-1482.
- Jiang, H., and W. Cotton, 2000: Large eddy simulation of shallow cumulus convection during BOMEX: sensitivity to microphysics and radiation. *J. Atmos. Sci.*, **57**, 582-594.
- Klein, S. A., and D. L. Hartmann, 1993: The Seasonal Cycle of Low Stratiform Clouds. *J. Climate*: **6**, 1587-1606.
- Koraçin, D., J. Lewis, W. T. Thompson, C. E. Dorman, and J. A. Businger, 2001: Transition of stratus into fog along the California coast: observation and modeling. *J. Atmos. Sci.*, **58**, 1714-1731.
- Krueger, S. K., G. T. McLean, and Q. Fu, 1995: Numerical simulation of the stratus-to-stratocumulus transition in the subtropical marine boundary layer. Part II: Boundary layer circulation. *J. Atmos. Sci.*, **52**, 2851-2868.
- Kuo, H.-C., and W. H. Schubert, 1988: Stability of cloud-topped mixed layers. *Quart. J. Roy. Meteor. Soc.*, **114**, 887-916.
- Lewellen, D. C., and W. S. Lewellen, 1998: Large-eddy boundary layer entrainment. *J. Atmos. Sci.*, **55**, 2645-2665.
- Lilly, D. K., 1968: Models of cloud-topped mixed layer under a strong inversion. *Quart. J. Roy. Meteor. Soc.*, **94**, 292-309.
- Lock, A. P., and M. K. MacVean, 1999: The generation of turbulence and entrainment by buoyancy reversal. *Quart. J. Roy. Meteor. Soc.*, **125**, 1017-1038.
- Lozej, C., 2001: The Santa Cruz eddy: observations and numerical simulations, *14th Conf. On Numerical Weather Prediction*, Fort Lauderdale, FL, Amer. Meteor. Soc.
- Mack, E. J., U. Katz, C. Rogers, and R. J. Pilié, 1974: The microstructure of California coastal stratus and fog at sea. Calspan Rep. No. CJ-5404-M-1 67, 67 pp.
- MacVean, M., and P. Mason 1990: Cloud top entrainment instability through small-scale mixing and its parameterization in numerical models. *J. Atmos. Sci.*, **47**, 1012-30.
- McGinley, J. A., 2001: Toward a surface data continuum: Use of the Kalman filter to create a continuous, quality controlled surface data set. *Preprints, 18th Conf. on Weather Analysis and Forecasting*, Ft. Lauderdale, FL, Amer. Meteor. Soc.

- McGinley, J., and J. Smart, 2001: On providing a cloud-balanced initial condition for diabatic initialization. *14th Conf. On Numerical Weather Prediction*, Fort Lauderdale, FL, Amer. Meteor. Soc.
- Miles, N. L., J. Verlinde, and E. E. Clothiaux, 2000: Cloud droplet size distributions in low-level stratiform clouds. *J. Atmos. Sci.*, **57**, 295-311.
- Moeng, C., D., Lenchow, and D. Randall, 1995: Numerical investigations of the roles of radiative and evaporative feedbacks in stratocumulus entrainment and breakup. *J. Atmos. Sci.*, **52**, 2869-2883.
- Moeng, C.-H., W.R. Cotton, B. Stevens, C. Bretherton, H.A. Rand, A. Chlond, M. Khairoutdinov, S. Krueger, W.S. Lewellen, M.K. MacVean, J.R.M. Pasquier, A.P. Siebesma, and R.I. Sykes, 1996: Simulation of a stratocumulus-topped planetary boundary layer: Intercomparison among different numerical codes. *Bull. Amer. Meteor. Soc.*, **77**, 261-278.
- Neiburger, M., D. S. Johnson, and C. Chien, 1961: Studies of the structure of the atmosphere over the eastern Pacific Ocean in summer; Part I: The inversion over the eastern North Pacific Ocean. *Univ. Calif. Publ. Meteor.*, **1**, 1-99.
- Nuss, W. A., J. M. Bane, W. T. Thompson, T. H., C. E. Dorman, F. M. Ralph, R. Rotunno, J. B. Klemp, W. C. Skamarock, R. M. Samelson, A. M. Rogerson, C. Reason, and P. Jackson, 2000: Coastally trapped wind reversals: progress toward understanding. *Bull. Amer. Meteor. Soc.*, **81**, 719-744.
- Oliver, D.A., W.S. Lewellen, and G. G. Williamson, 1978: The interaction between turbulent and radiative transport in the development of fog and low-level stratus. *J. Atmos. Sci.*, **35**, 301-316.
- Paluch, I., and D. Lenschow, 1991: Stratiform cloud formation in the marine boundary layer. *J. Atmos. Sci.*, **48**, 2141-2158.
- Pilié, R., E. Mack, C. Rogers, U. Katz, and W. Kocmond, 1979: The formation of marine fog and the development of fog-stratus systems along the California coast. *J. Appl. Meteor.*, **18**, 1275-1286.
- Pruppacher, H. R., and J. D. Klett, 1997: *Microphysics of clouds and precipitations*. 2d ed., Kluwer Academic, 954 pp.
- Ramer, J., 1993: An empirical technique for diagnosing precipitation type from model output. *Fifth Conference on Aviation Weather Systems*, Vienna, Virginia, Amer. Meteor. Soc.

- Randall, D. A., 1980: Conditional instability of the first kind upside down. *J. Atmos. Sci.*, **37**, 125-130.
- Randall, D., 1984: Stratocumulus deepening through entrainment. *Tellus*, **36A**, 446-457.
- Shaw, B. L., J. A. McGinley, and P. Schultz, 2001: Explicit initialization of clouds and precipitation in mesoscale forecast models. *14th Conf. On Numerical Weather Prediction*, Fort Lauderdale, FL, Amer. Meteor. Soc.
- Siems, S. T., C. S. Bretherton, M. B. Baker, S. Shy, and R. E. Breidenthal, 1990: Buoyancy reversal and cloud-top entrainment instability. *Quart. J. Roy. Meteor. Soc.*, **116**, 705-739.
- Simon, R. L., 1977: The summertime stratus over the offshore waters of California. *Mon. Wea. Rev.*, **105**, 1310-1314.
- Smart, J., S. Albers, D. Birkenheuer, B. Shaw, S. Early, J. Edwards, P. Stamus, and S. Wagoner, 2000: LAPS at the Air Force Weather Agency. *Fsl Forum*, **2**.
- Stamus, P., and J. McGinley, 1997: The Local Analysis and Prediction System (LAPS): Providing weather support to the 1996 Summer Olympic games. *13th International Conference on Interactive Information and Processing Systems for Meteorology, Oceanography, and Hydrology*, Long Beach, CA, Amer. Meteor. Soc.
- Stamus, P. A., and J. A. McGinley, 1991: Friction in the LAPS high-resolution surface analysis. *12th International Conference on Weather Analysis and Forecasting*, Dallas, TX, Amer. Meteor. Soc.
- Stephens, G., 1996: How much solar radiation do clouds absorb? *Science*, **271**, 1131-33.
- Stevens, B., 1996: On the dynamics of precipitating stratocumulus. Ph.D. Dissertation, Atmospheric Science Department, Colorado State University, 92 pp.
- Stevens, D. E., and C. S. Bretherton, 1999: Effects of grid resolution on large-eddy simulation of radiatively driven entrainment through a strong inversion. *J. Comput. Phys.*, **129**, 284-295.
- Stevens, B., W. R. Cotton, G. Feingold, and C. Moeng, 1998: Large-eddy simulations of strongly precipitating, shallow, stratocumulus-topped boundary layers. *J. Atmos. Sci.*, **55**, 3616-3638.
- Stevens, D. E., J. B. Bell, A. S. Almgren, V. E. Beckner, and C. A. Rendleman, 2000: Small-scale processes and entrainment in a stratocumulus marine boundary layer. *J. Atmos. Sci.*, **57**, 567-581.

- Szoke, E., 2000: An assessment of the utility of a local model for operational mountain snowfall predictions. [Available on line from http://laps.fsl.noaa.gov/cgi/LAPB.pubs_00.cgi]
- Vali, G., R. D. Kelly, J. French, S. Haimov, and D. Leon, 1998: Fine scale structure and microphysics of coastal stratus. *J. Atmos. Sci.*, **55**, 3540-3564.
- Wyant, M. C., C. S. Bretherton, H. A. Rand, and D. E. Stevens, 1997: Numerical simulations and a conceptual model of the stratocumulus to trade cumulus transition. *J. Atmos. Sci.*, **54**, 168-192.
- Yamada, T., 1978: A three-dimensional, second-order closure numerical model of meso-scale circulation in the lower atmosphere. Topical Rep. ANL/RER-78-1, Radiological and Environmental Research Division, Argonne National Laboratory, Argonne, IL, 67 pp.

Figure 1

Figure 2

LAPS WIND ANALYSIS

Figure 3

LAPS CLOUD ANALYSIS

Figure 4

LAPS DERIVED CLOUD / HYDROMETEOR FIELDS

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