OPEN WATER / WET SOIL EVAPORATION FROM THE RIO GRANDE RIVER

FINAL REPORT

by

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I.  STATEMENT OF THE PROBLEM

The U.S. Bureau of Reclamation has the responsibility to maintain the water level in the Rio Grande while ensuring that competing water demands are fulfilled and that the amount of water leaving New Mexico is sufficient to satisfy water obligations to Texas and Mexico. Evaporation from the river along with evapotranspiration from the riparian and agricultural areas represents approximately two-thirds of the total losses from the system of surface water between Cochiti Reservoir and Elephant Butte Reservoir (Brower, 2000). Accurate estimates of the evaporative water demand along the river, including the riparian areas, will allow for better management of the annual water supply by providing the water managers with more accurate estimates of the annual water supply. The field campaign described here is part of an effort to quantify the amount of natural evaporation depletion losses from the active channel surfaces flows and exposed soil and develop simple methods for their estimation.

Evaporation from wetted sands is a particular issue. Evaporation of water that is stored in sandbars or in sand fringes of the channels represents a loss of available water. Water may be temporarily stored in sand during periods of decline in the mean flow, but even during stable river flows there will be some wetting of sands due to capillary action and some loss to evaporation. However, it is not known whether this is a serious loss mechanism or not.

The estimation methods developed here will be used in the ET toolbox to estimate water losses and use in near real time. The goal of the ET Toolbox project is to develop a
methodology for automatically acquiring daily riparian and crop water use estimates, as well as open water and channel evaporation estimates. The primary purpose of the ET Toolbox is to accumulate daily rainfall and water use estimates for each area within specified river reaches. The daily ET estimates and summary year-to-date cumulative ET estimates from the toolbox are available to users and water managers via the Internet for use in planning (Brower, 2000).

II. SITE DESCRIPTION

The field campaign was situated on the Rio Grande River at the south end of the Rio Grande Nature Center State Park in Albuquerque, NM (Figure 1). In the vicinity of Albuquerque, the Rio Grande has a sand and gravel riverbed. The river has low, sandy, erodible banks in this area, and the wide sandy river channel has many bars. The presence of numerous wandering bars and middle islands is a strong characteristic of this reach. The river channel averages between 420 and 510 ft wide.

The experiment site is approximately 1¼ miles north of where Interstate 40 crosses the river. Here the river is flanked to the west by a narrow riparian area, primarily consisting of dense salt cedar with several mature cottonwood mixed in, at the base of a steep cliff that is rimmed above with residential homes. To the east is a considerably wider stretch of riparian area. On this bank, the riparian area is primarily composed of mature cottonwoods with a narrow strip of tall grass and a few saltcedar along the river’s edge. Behind the riparian zone on the east bank is a
bike trail which follows along a levee. Beyond the levee is a residential neighborhood and the city of Albuquerque proper. A closer view of the river and riparian areas at our experiment site can be seen in the satellite photo in Figure 2. A panoramic view of the river at our site can be seen in Figure 3.

While the month of August is within the “monsoon season” of New Mexico, August of 2006 brought uncharacteristically heavy rains causing the river stage to be much greater than anticipated. Figure 4 shows the average daily discharge in the Rio Grande. The discharge was unusually high at the start of the campaign although the river level was steadily falling. However on the 13th, a record-setting rainfall dramatically lifted the river level again. The north sandbar was largely washed away so that the instruments needed to be re-sited closer to shore. These rains had two deleterious effects on the measurement campaign. The first is that the ensuing high waters limited the size of the sand bars. One of the requirements for meaningful evaporation measurements is a homogeneous surface in the upwind direction extending a distance that is 50 to 100 times the height at which the atmospheric measurements are made. This requirement was fully met for the north tower only during the 10th through 13th of August. The actual measurements for the other days are a combination of sand bar and open water evaporation. The other effect is that fully dry sand bars were never observed. In all cases for all days, the water level was never less than two inches below the top of the sand.

![Figure 3.](image)

**Figure 3.** A panoramic view of the river from the lidar position at 0800 on the 11th of August. This day is the lowest water level observed during the study. Note the extent of the sandbar in the lower center of the photograph.

![Figure 4.](image)

**Figure 4.** The daily mean discharge (cubic feet per second) from USGS station 08330000 on the Rio Grande at Albuquerque, NM. The station is located at 35°05'21"N, 106°40'49"W in Bernalillo County, New Mexico. ([http://waterdata.usgs.gov/nwis/nwisman/?site_no=08330000&agency_cd=USGS](http://waterdata.usgs.gov/nwis/nwisman/?site_no=08330000&agency_cd=USGS))
bars. In most cases, the silty clay layer capping the sand bars was usually moist. That moist sand bars will evaporate is not in question. Whether sandbars with tops of order a foot above the mean river level have significant evaporation is unknown and could not be measured here. A comparison of figures 3 and 5 show the magnitude of the change in the sandbars as a result of the storms. Figure 5 in particular shows the damp nature of the sandbars that were observed.

III. INSTRUMENTATION

In order to measure evaporation from the Rio Grande and in particular, from the sand bars inside it, a set of hydrological and meteorological sensors were fielded on an island on the far side of the river channel as well as along the along the bank. An intensive measurement campaign was conducted from 9 August, 2006 [day of year (DOY) 220] through 18 August, 2006 (DOY 230). This campaign included the Los Alamos National Laboratory (LANL) scanning Raman lidar as well as high frequency, time series measurements of the meteorological instruments (Figure 6).
**A. Meteorological Instrumentation.** Two meteorological stations were installed in the Rio Grande River. Tripods (North Station) were placed on a sand bar just off the east bank of the river to the north of the Raman lidar location (Figure 6). Along with the primary tripod at the North Station, two additional tripods were installed nearby; one for a Licor water-vapor temperature profile system and one for a net radiometer. The second station (South Station) was placed south of the Raman lidar on an island on the west side of the river channel (Figure 7). This island was well-separated by water from the land mass to the west. The South Station was installed later during the field campaign, 13 August (DOY 225), and continued to collect data through the end of the experiment. Instrumentation mounted at each of the stations along with brief descriptions is detailed in Tables 1 and 2. Both stations were operating at a measurement rate of 10 Hz. A description of available data sets for both stations is contained in Appendix 1.

The instruments on the sand bars were sited to provide the maximum amount of fetch (the distance of sand in the direction of the mean wind) and so were located on the northern sides of the bars. The north site had as little as 30 m of fetch and as much as 75 m. The south site had the most, with nearly 100 m when installed. Both of the sandbars were predominantly sand with a 1 to 3 cm “cap” of moist clay silt. Neither of the sandbars had any vegetative canopy. The average height of the sandbars above the river level ranged from 2 to 20 cm for the southern sandbar (see for example figure 8) to 10 to 30 cm for the north. The prevailing wind direction during this time of year was from the S-SW and was expected to follow the river channel, the flow being constrained by the cliffs on the west side of the river. The north and south EC towers were both oriented to the south to take advantage of this.

Instrumentation fielded in this campaign makes up an energy/water budget and meteorological flux station. The water/energy budget is comprised of measurements of net radiation, storage in the canopy (available energy), turbulent fluxes of sensible and latent heat, and momentum flux. Available energy is estimated with
measurements from a net radiometer. Sensible, latent heat (evaporation) and momentum fluxes are measured by combining a sonic anemometer and a fast response hygrometer. Each of the EC systems was comprised of a 3-D sonic anemometer (CSAT3, Campbell Scientific Inc.) and a KH20 Hygrometer (Campbell Scientific Inc.). Additional meteorological instrumentation included an air temperature and humidity probe (HMP-45, Vaisala) on each tower. An infrared temperature sensor (IRPT5, Apogee Inc.) was positioned on the tower to measure the surface temperature of the sandbars (figure 9). All of the instruments were wired and programmed to a single data logger (CR5000, Campbell Scientific Inc.) so that simultaneous and synchronized data acquisition could be accomplished. The EC systems were programmed to sample at a rate of 20 Hz while the temperature and humidity sensors were sampled at a rate of 1 Hz. The high frequency EC data series were conditioned following common procedures in micrometeorology described in Kaimal and Finnigan, 1994.

A sonic anemometer measures the three dimensional components of the wind flow (u, v, and
w, the three components of the wind speed) at high rates, up to 60 Hz. These wind observations were used to compute the vertical transport term of the eddy covariance fluxes. The krypton hygrometer measures the fluctuations of atmospheric water vapor concentration (q’) at the same rate as the sonic anemometer (20 Hz). Together, the fluctuations in the vertical wind speed (w’) and water vapor concentration (q’) about their mean values were used to calculate the evaporative flux from the lake as \( \overline{w’ q’} \), where the overbar indicates time averaging.

The covariance of these variables is the “standard” reference for flux determination (subject to various corrections for instrument limitations [Webb et al., 1980; Massman, 2000; Massman and Lee, 2002]). This method is accepted as the most physically based technique to measure evaporation and was used in this project as the “truth set”. The temperature/humidity probe is used as part of a standard meteorological station that will be

![Figure 10. Comparison of measured latent heat fluxes from the North and South Stations on August 17, 2006. Fluxes were measured using the eddy correlation technique.](image)

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Height Above the Surface</th>
<th>Measures</th>
</tr>
</thead>
<tbody>
<tr>
<td>CSAT3, Sonic Anemometer</td>
<td>1.8 m</td>
<td>Wind speed, direction, turbulent quantities</td>
</tr>
<tr>
<td>KH20, Krypton Hygrometer</td>
<td>1.8 m</td>
<td>Water vapor concentration fluctuations</td>
</tr>
<tr>
<td>HMP45C, Temp-Humidity Probe</td>
<td>1.8 m</td>
<td>Atmospheric temperature and humidity</td>
</tr>
<tr>
<td>IRTP5 Infrared Thermocouple Sensor</td>
<td>1.8 m</td>
<td>Surface temperature</td>
</tr>
<tr>
<td>Hydraprobe</td>
<td>5 cm below the surface</td>
<td>Volumetric Soil Moisture</td>
</tr>
<tr>
<td>TCAV Soil Thermocouple Probe</td>
<td>2, 6 cm below the surface</td>
<td>Soil Temperature</td>
</tr>
</tbody>
</table>
used as a data set for evaluating the utility of more advanced models for estimating evaporative flux from the river and sandbars.

As previously stated, the latent heat flux (LE) calculated using the sonic anemometer and krypton hygrometer is used as the “truth set”; all other models will be compared to it. Figure 10 illustrates eddy correlation estimates of evapotranspiration as measured from the North Station and South Station on August 17th.

### B. NASA Meteorological Instrumentation.

An eddy correlation towers was set up in Albuquerque, NM in March 2005. The instrument was located approximately 30 m to the north of the lidar location and 5 m from the river. The instruments employed on this station are detailed in Table 3.

Evaporation data from this station were made available by Jan Kleissl, New Mexico Tech.

While the station is on the bank next to the river, it was emplaced in this location that juts into the river to estimate river evaporation. When the winds are predominantly from the north or south (i.e. along the river),

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Height Above the Surface</th>
<th>Measures</th>
</tr>
</thead>
<tbody>
<tr>
<td>CSAT3, Sonic Anemometer</td>
<td>2.5 m</td>
<td>Wind speed, direction, turbulent quantities</td>
</tr>
<tr>
<td>LI7500 Hygrometer</td>
<td>2.5 m</td>
<td>Water vapor concentration fluctuations</td>
</tr>
<tr>
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<td>1.8 m</td>
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<td>2.5 m</td>
<td>Surface temperature</td>
</tr>
<tr>
<td>NR lite</td>
<td>3 m</td>
<td>Net Radiation</td>
</tr>
<tr>
<td>Soil Thermocouple Probe</td>
<td></td>
<td>Soil Temperature</td>
</tr>
</tbody>
</table>

[Pinheiro, 2005]
the height of the sensors should provide an estimate of river evaporation rather than the wooded region that surrounds the location (http://infohost.nmt.edu/~jk/CFNC.htm).

**C. Raman Lidar System.** The University of Iowa / Los Alamos National Laboratory (LANL) Raman lidar was used during the field campaign and located along the east bank of the river in between the two meteorological stations as seen in Figure 12. The Raman lidar system was operational between the hours of 0600 and 2200, weather permitting, with the exception of one 24-hour period on 10 August. The lidar scanned over a path of about 180° beginning downstream and ending upstream from the lidar. A description of available data sets for the Raman lidar system is contained in Appendix 2.

Raman lidars use a technique originally pioneered by Fiocco and Smullins (1963), Cooney *et al.* (1969), Cooney (1970), and Melfi *et al.* (1969). A Raman lidar operates by emitting a pulsed laser beam, usually in the ultraviolet or near ultraviolet, into the atmosphere. Atmospheric gases, such as nitrogen, oxygen, and water vapor interact with this light via the Raman scattering process, causing light of longer wavelengths to be scattered. The amount of the wavelength shift is unique to each molecule. This enables the measurement of different atmospheric gaseous species by this technique.

The University of Iowa / Los Alamos National Laboratory (LANL) Raman lidar (shown schematically in Fig. 12) is a typical Raman lidar. In this lidar, the laser is mounted below the telescope. A series of mirrors and lenses is used to expand the beam to make it eyesafe and collinear with the telescope. A forty-five-degree angled mirror is used to change the optical direction to vertical allowing the system to make vertical

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**Figure 12.** The University of Iowa / Los Alamos National Laboratory Raman lidar system along the bank of the Rio Grande River in Albuquerque, NM.

**Figure 13.** Diagram showing the layout of the University of Iowa / Los Alamos Raman lidar. With the exception of the scanning mirror, the arrangement is typical of Raman lidars.
soundings. With the scanning mirror mounted, the system can perform three dimensional scanning near the earth’s surface. At the back of the telescope, a series of dichroic beam splitters are used to separate the elastically scattered light from the light at the two Raman shifted wavelengths from nitrogen and water vapor. Narrow band interference filters block unwanted wavelengths in each channel. To work during the day, many systems operate in the region of the spectrum below about 300 nm where ozone and oxygen strongly absorb sunlight and are thus “blind” to solar photons. Solar blind operation requires the use of a laser near 250-260 nm so that the Raman shifted lines will be below 300 nm. Because of the limited amount of returning light, Raman lidar systems tend to use large, powerful lasers and large telescopes. Therefore, they are unusually large and require significant amounts of power. The typical maximum horizontal range for the LANL lidar is approximately 700 m when scanning, with a corresponding spatial resolution of 1.5 m over that distance. The upper scanning mirror allows three dimensional scanning in 360 degrees in azimuth and ±22 degrees in elevation.

The Raman technique obtains the water vapor mixing ratio, \( q_w(r) \), from the ratio of the signal magnitude in the water vapor channel, \( P_{H2O}(r) \) to the magnitude of the signal in the nitrogen channel, \( P_{N2}(r) \), using (Melfi, 1972):

\[
q_w(r) = \frac{P_{H2O}(r)}{P_{N2}(r)} \left[ \frac{C_{N2} \sigma_{N2} M_{N2}}{C_{H2O} \sigma_{H2O} M_{H2O} f_{N2}} \right] \exp \left[ \int_0^h \left[ \kappa_L(r', \lambda_{N2,R}) - \kappa_L(r', \lambda_{H2O,R}) \right] dr' \right]
\]

(1)

where \( \lambda_{N2,R} \) and \( \lambda_{H2O,R} \) are the Raman N\(_2\) and H\(_2\)O scattered wavelengths; \( P_{N2}(r) \) and \( P_{H2O}(r) \) are the received signals in the nitrogen and water vapor channels; \( \sigma_{N2} \) and \( \sigma_{H2O} \) are the Raman backscatter cross-sections for the laser wavelength; \( n_{N2}(r) \) and \( n_{H2O}(r) \) are the number density of nitrogen and water molecules at range, \( r \); \( \kappa_L(r, \lambda_{N2,R}) \), and \( \kappa_L(r, \lambda_{H2O,R}) \) are the total attenuation coefficients at the Raman-shifted wavelengths of nitrogen and water vapor molecules; \( f_{N2} \) is the fractional \( N_2 \) content of the atmosphere (0.78084); and \( C_{N2} \) and \( C_{H2O} \) are the system coefficients which take into account the effective area of the telescope, the transmission efficiency of the optical train, and the detector quantum efficiency at the Raman shifted wavelengths. Thus the water vapor mixing ratio at any distance is given by the ratio of the magnitude of the signal in the water vapor channel to the magnitude of the signal in the nitrogen channel, a multiplicative constant (the part in square brackets in Eq. 1), and a small exponential correction due to difference in extinction between the nitrogen-shifted and water vapor shifted wavelengths. Comparison of the lidar signals to conventional hygrometers can be used to determine the
multiplicative constant. Because the signal to noise ratio decreases with distance, the uncertainty in the mixing ratio values is a function of distance from the lidar. For a mid-range distance (~350 m), the estimated uncertainty is approximately 3.6 percent. This is consistent with the comparisons of lidar and calibrated references over land surfaces along horizontal paths. The standard deviation between the lidar and capacitance hygrometer data taken at concurrent times and locations was found by regression to be ±0.34 g/kg [Eichinger et al., 2000; Cooper et al., 1996].

When the Raman lidar aims along a given line of sight, data are obtained every 1.5 m along that line. By aiming the lidar in a series of different directions, a two or three dimensional map of the water vapor concentrations can be assembled. Figure 14 is a typical scan from the LANL/UI Raman lidar showing the water vapor concentration in one vertical plane along the river during the campaign.

**Figure 14.** A vertical slice of the atmosphere showing the water vapor concentration made during the campaign.

**Figure 15.** Measured evaporation rates from the eddy correlation towers located on sand bars in the Rio Grande. Abnormal excursions (such as evening of 13/14th) were due to storms.
experiment. The intense red color at the bottom is a result of the attenuation of the laser beam by the ground or canopy (note that for some angles there is no enhanced return from the water in the river). The change in the lidar signal as it reaches the canopy top enables one to determine the shape and orientation of the surface.

IV. EVAPORATION ESTIMATES

A. River Eddy Correlation

Results from the eddy correlation towers located on the two sand bars in the river are shown in figure 15. Essentially the two sets of measurements are the same to within instrument uncertainty. The most notable exception is the evening of 13/14 August which experienced one of the most significant major storm events in New Mexico history.

It is interesting to note in figure 16, an example of the daily variation of the energy fluxes at the north site, that the soil heat flux is largest in the morning and that the latent energy flux is relatively small in the morning. The sensible heat flux is generally small and varies little throughout the day. This is an intermediate condition between a field situation where the sensible, latent, and soil heat fluxes vary proportionally with the net radiation and a deep water body in which the latent and sensible heat fluxes are relatively small, but relatively unchanging with net radiation. Note that what we are calling the soil heat flux is the amount of solar energy stored at the surface, which could be either the sand surface of the islands or in the river water. This soil heat flux is
larger than the latent energy flux in the morning and larger than the sensible heat flux during most of the daylight hours.

This intermediate condition is an indicator that the eddy correlation instruments were likely placed so that the source region included both the sand islands and the river. For the sand areas alone to act in the manner shown in figure 16 would imply some kind of energy storage mechanism. While energy is stored in the sand, solar energy is deposited at the soil surface and can be stored deeper only after transport by conduction. This is an inefficient mechanism when compared to energy storage in river waters where solar energy is deposited at all depths and turbulent mixing distributes the energy rapidly. It is difficult to imagine a heat storage mechanism for the sand surface that would allow the observed sensible and soil heat fluxes to occur. Again, the most likely interpretation of these results is that the sensors were “seeing” a mixture of sand and river surfaces.

B. NASA Eddy Correlation

The NASA station located on the bank of the river would be expected to measure the average evaporation in the river over a much larger area than the instruments located on the bars close to the river and would be expected to include both open water and sand bar evaporation. This is true when the winds blow from the north or south along the river (which is normally the case). It then represents a condition that should be between the evaporation rates of the sand bars and open water.

A comparison of the measured evaporation rates from the north tower in the river and the NASA station located close by on the bank is shown in figure 18. What is striking is that the evaporation rates are the same to within the uncertainty of the measurements. This indicates that there is little or no difference between the larger scale evaporation and that from near the surface of the sand bars. This in turn implies that there is little difference between the open water evaporation rate and the sand bar evaporation rates. It may also indicate that the instruments in the river were measuring a mixture of the river and the sandbars. This is probably true, especially for those days when the river was high and the sand bar fetch was extremely limited. Since there were days when there was significant fetch for the sand bars, if the issue was limited fetch, then one would expect that the measurements on the bank would agree for days of limited fetch and differ for days of significant fetch. Significant differences between the NASA site at the side of the river and the two sets of instruments were not observed.
Figure 18. A comparison of measured evaporation rates from the eddy correlation tower located on the north sand bar in the Rio Grande and the NASA tower located on the east bank.

C. Raman Lidar

Data from the Raman lidar was processed to produce evaporation estimates over the river. Figure 19 is a comparison of the evaporation estimates from the eddy correlation instrument on the southern sand bar site and from the lidar at the same location for the 16th of August. The trends from the two instruments are similar with a root mean square difference of 23 W/m². This difference is slightly larger than the uncertainty of the eddy correlation.

Figure 19. A comparison of evaporation rates from the lidar and eddy correlation instruments over the south sand bar site.
instruments (10 to 15%).

A collection of plots of the spatially resolved lidar ET estimates is shown in figure 21. What is striking is the lack of differences across the river. There are no systematic trends over the sand bars. There are several sections for which ET estimates are not available because they contain vegetation or are on the far side of vegetation. Table 4 is a compilation of the RMS differences in the ET estimates across the river and the fraction of the average ET estimate.

The differences that are seen over the river are higher than the fractional differences between the lidar and eddy correlation instruments. This would imply the differences may be real, but there appears to be no recurring differences. This may be due to local, temporary differences in water temperature in the river. Because the river is located in a low area (forested area to the east and a bluff to the west, some sections of the river may not illuminated by the sun at any given time. These sections would change as the sun crossed the sky. Wind may also play a role. Note that the fractional differences increase after noon when the wind along the river picks up.

If the evaporation rates over the sand bars is different than that over the river, it would be expected that the water vapor concentrations would be measurably different. As can be seen in figure 22, the water vapor concentration across the river and island is remarkably uniform. Vertical slices of the atmosphere are shown at four different times; all of which show little variability across the river.

### Table 4. Differences in Estimated ET over the River

<table>
<thead>
<tr>
<th>Time</th>
<th>RMS Diff W/m²</th>
<th>Fractional Diff</th>
</tr>
</thead>
<tbody>
<tr>
<td>0900</td>
<td>7.4</td>
<td>10%</td>
</tr>
<tr>
<td>1000</td>
<td>10.9</td>
<td>13%</td>
</tr>
<tr>
<td>1100</td>
<td>16.9</td>
<td>16%</td>
</tr>
<tr>
<td>1200</td>
<td>13.4</td>
<td>10.3%</td>
</tr>
<tr>
<td>1300</td>
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<td>24%</td>
</tr>
<tr>
<td>1400</td>
<td>63.7</td>
<td>23%</td>
</tr>
<tr>
<td>1500</td>
<td>66</td>
<td>32%</td>
</tr>
<tr>
<td>1600</td>
<td>35</td>
<td>20%</td>
</tr>
</tbody>
</table>
Figure 21. Plots of the evaporation rates over the river during the 16th of August. Location 0,0 is the location of the lidar as shown on the map. North is to the left. Each grid square is 25m on a side.
V. CURRENT EVAPORATION ESTIMATION METHODS

A. ET Toolbox

The Bureau of Reclamation operates the Agricultural Water Resources Decision Support (AWARDS) system. Through the Evapotranspiration (ET) Toolbox interface, the AWARDS system provides information to farmers and water managers on when and where to water is needed. The calculation of daily ET, as applied in the AWARDS system, and Daily Consumptive Use (DCU), as applied in the ET Toolbox, are based upon open water coefficients used in the ET toolbox. After Jensen (1998).
derivation on a Reference ET \((ET_0)\) and a coefficient \((K_C)\) that is specific to the type of surface or canopy in the area of interest (Brower, 2004). This coefficient may be dependent on the Growing Degree Days (in the case of vegetation) or the month of the year. The coefficient \((K_C)\) is applied to \(ET_0\) to estimate the daily \(ET\) in inches using:

\[
ET = (K_C) \ ET_0
\]

where \(ET_0\) is the reference \(ET\) as calculated by the Penman equation for a specified surface, which may be open water or a well-watered grass or alfalfa crop. While values for the coefficient, \(K_C\) have been measured by a number of researchers, values for open water are not common and values for sand bars were not found. Jensen (1998) gives open water evaporation coefficients that are used in the ET Toolbox (figure 23).

Based on the \(ET\) daily values, the Daily Consumptive Use \((DCU)\) is calculated for an area as:

\[
DCU_{total} = \left[ \frac{\sum_{k=1}^{N} ET_{0k} \ K_C \ acres_k}{1.98347} \right] - \text{rain} = \left[ \frac{\sum_{k=1}^{N} ET_k \ acres_k}{1.98347} \right] - \text{rain}
\]

with \(acres_k\) is the acreage of the surface type of a given area, and \(rain\) is the NEXRAD estimated daily accumulated rainfall. \(DCU_{total}\) is calculated as a flow in units of ft\(^3\)/s.

The ET Toolbox uses a daily average Penman formulation as its estimate of \(ET_0\). The modified formulation for \(ET_0\) used was developed by researchers at the New Mexico State University. The formulation was is empirically derived from experimental data based on a grass referenced method that combines energy balance and heat and mass transfer functions (ASCE, 1990). In terms of the variables used in this analysis, the ET Toolbox reference ET can be written as:
\[
ET_o = \frac{10}{L_e} \left[ \frac{\Delta}{\gamma + \Delta} R_n + \frac{2 \gamma}{\Delta + \gamma} D F(u) \right] \quad \text{mm / day}
\]

\[
R_n = 0.95 \left[ \frac{(1 - R)}{0.041868} S_{\text{downwelling}} \right] - 64 \quad \text{cal / (cm}^2 \text{ day)}
\]

where:

- \( S = \) downwelling solar radiation (MJ/m\(^2\)-day),
- \( D = \) water vapor deficit, defined in the ET Toolbox as
  \[
  D = \frac{(e_{sat}(T_{\text{max}}) + e_{sat}(T_{\text{min}}) - \rho_{\text{max}} e_{sat}(T_{\text{max}}) - \rho_{\text{min}} e_{sat}(T_{\text{min}}))/2}{\text{mbar}},
  \]
- \( \rho_{\text{min}} = \) minimum relative humidity during the day,
- \( \rho_{\text{max}} = \) maximum relative humidity during the day,
- \( T_{\text{min}} = \) minimum temperature during the day,
- \( T_{\text{max}} = \) maximum temperature during the day,
- \( R = \) albedo of the earth’s surface, taken in the Toolbox as 0.21,
- \( F(u) = \) wind function, defined in the ET Toolbox as
  \[
  F(u) = 15.36 (1.0 + 0.0062 \times 3.6 \times 24 \times u(2m))
  \]
  \((\text{km/day})\),
- \( u(2m) = \) wind speed at an altitude of 2 m (m/s),

The Jensen (1998) open water evaporation coefficient for August that is used in the ET Toolbox is 0.85. Using this coefficient and the Penman formulation described above, the predicted \( ET \) for the study period is shown in Fig 24. It can be seen that the ET Toolbox consistently over-predicts the actual \( ET \) by 31% during the study period. Table 5 is a listing

<table>
<thead>
<tr>
<th>Penman</th>
<th>Measured</th>
<th>Kc</th>
</tr>
</thead>
<tbody>
<tr>
<td>158.7</td>
<td>110.7</td>
<td>0.698</td>
</tr>
<tr>
<td>138.1</td>
<td>81.1</td>
<td>0.587</td>
</tr>
<tr>
<td>168.1</td>
<td>89.2</td>
<td>0.531</td>
</tr>
<tr>
<td>127.7</td>
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<td>0.659</td>
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<tr>
<td>129.7</td>
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<td>139.0</td>
<td>79.3</td>
<td>0.571</td>
</tr>
<tr>
<td>73.0</td>
<td>41.2</td>
<td>0.564</td>
</tr>
<tr>
<td>129.8</td>
<td>94.8</td>
<td>0.731</td>
</tr>
<tr>
<td>154.2</td>
<td>96.0</td>
<td>0.623</td>
</tr>
<tr>
<td>153.2</td>
<td>101.3</td>
<td>0.661</td>
</tr>
</tbody>
</table>

average 0.649
std dev 0.0999

Table 5: A comparison of daily Penman ET estimates and measured ET values. The calculated Kc values for each day are also shown, with the average and standard deviation.
of the measured ET values and the Penman predicted values along with the calculated $Kc$ value for each day. The $Kc$ calculated for the study period is 0.65 ± 0.1. While this figure is significantly lower than the ET Toolbox value, it should be used with caution since the study period was quite short and the weather during the period was usually wet.

The reason for the difference can be seen in figure 25, a comparison of half-hourly Penman predictions and the measured evaporation rates. The pattern is that the Penman approach overestimates evaporation in the mornings and more correctly estimates it in the afternoon. A more direct one to one comparison is shown in figure 26. Note the high Penman estimates at low measured evaporation rates. These correspond to the morning values where the Penman method overestimates the evaporation. There is a cluster of values where the Penman method predicts no evaporation and yet there are significant measured values. These come from night times when there is no net radiation to drive the Penman model. Because the river is a wet area in the middle of a larger desert region, any kind of a wind will cause significant evaporation. These are limitations of the Penman model that is driven by the available energy $(Rn - G)$. If these values change, the model has no method to correct or compensate for these changes.

B. US Army Corps of Engineers (USACE) Method

In the USACE Rio Grande model (USACE 2005), evaporation from the river channel is a function of the amount of water flowing in the river. At the bankfull discharge rate, all sand bars in the river channel are assumed to be covered with water and the wetted sands not subject to evaporation loss. This bankfull discharge generally has a 1- to 2-year return interval (Leopold
and others, 1964). As stream discharge drops below the bankfull discharge, sand bars become exposed and subject to evaporative losses. A factor of 0.25 is used to correlate the evaporation of water by capillary action through sand bars in the river channel to an adjacent evaporation pan measurement (Sorey and Matlock, 1969).

For the reach from Albuquerque to Bernardo (bankfull discharge = 4820 cfs and corresponding surface area = 5175 acres) the evaporative losses are:

For \( Q < 4820 \text{ cfs} \);
\[ L = Pan_e (124 Q^{0.44}) + 0.25 Pan_e (5175 - 124 Q^{0.44}) \]

For \( Q \geq 4820 \text{ cfs} \);
\[ L = Pan_e (124 Q^{0.44}) \]

where:

\( Q \) = Mean daily discharge at the upstream end of the reach, in cfs;

\( L \) = Loss from water surface evaporation and wetted sands in the reach, in acre-ft/day; and

\( Pan_e \) = Pan evaporation data for the site nearest to the reach under consideration, in ft/day.

Similarly for the reach from San Felipe to Albuquerque (bankfull discharge = 4820 cfs and corresponding surface area = 2718 acres) the evaporative losses may be estimated by:

For \( Q < 4820 \text{ cfs} \);
\[ L = Pan_e (84 Q^{0.41}) + 0.25 Pan_e (2718 - 84 Q^{0.41}) \]

For \( Q \geq 4820 \text{ cfs} \);
\[ L = Pan_e (84 Q^{0.41}) \]

An average estimate is shown in figure 27 of the predicted evaporative losses during the study period. Pan evaporation estimates were taken from the Los Lunas pan site south of Albuquerque. This method greatly overestimates the losses (by a factor of more than 2). In as much as the study period was quite wet and overcast, thus reducing the size of the oasis effect, this is not surprising.

**Figure 27.** The evaporative losses predicted using the USACE method compared to the ET Toolbox method and measured evaporation. Method 1 is the method suggested for Albuquerque to Bernardo reach and method 2 is the method suggested for San Felipe to Albuquerque.
VI. CONCLUSIONS

The analysis of the energy balance measurements made at the Rio Grande riverbed in Albuquerque, NM in August 2006 show no differences in evaporation between the evaporation from wet sand and the river. The energy balance systems show no differences in patterns between days when the sand bars are mostly flooded and those when significant sand fetch existed. The lidar shows no systematic trends in evaporation over the sandbars. The lidar also shows the water vapor content over the river to be essentially constant, with the sand bars being neither more moist or drier than the surrounding river. We are forced to conclude that to the ability of the instruments to measure (approximately 10%), the evaporation rates over the wet sand bars is the same as that over the river.

A comparison of the eddy correlation measurements and the method currently used in the ET Toolbox show that the Jensen (1998) open water evaporation coefficient for August (0.85) consistently over-predicts the actual $ET$ by 31% during the study period. This may be due to the unusually wet conditions and rain during the study period.

It should be recognized that the data obtained during the campaign represent an atypical situation for the river bottom, large volume of water flowing through the Rio Grande as a result of intense monsoonal activity. The sand fetch surfaces were minimal at best and quite variable in size as the river rose and fell. The soil water content in the sandbars were completely saturated. The result is that eddy covariance footprints were variable in time and space, making interpretation of the latent heat fluxes tenuous. What was measured was a mixed water and sand in varying mixture. The best one can say about the LE fluxes from this study is that they represent what could be termed “river channel” evaporation.

The river at the time of the campaign is not a typical example of the river bed condition for this particular location and time period. What was measured is an extreme case but certainly not typical. More data that represent the conditions that the BOR is interested is required to improve wet sand model evaporation estimates.

Spectral analysis of the high frequency turbulence data suggest that the eddy covariance measurements are of high quality as are measurements from the lidar. It is apparent that the greatest uncertainty in the surface energy balance is the soil heat flux which is unusually large for soil surfaces and varies in an unusual manner. Under the conditions of the study in 2006, this term was not trivial and could easily account for 40% of the energy balance for the riverbed site.
This effect was not anticipated and the least attention was paid to this term. The measurements made did not adequately quantify the soil heat flux. Future measurements will have to include a more comprehensive measurement plan to measure soil temperatures at multiple depths (not just 2, and not just approximately) and must also include the river. It is necessary to measure the energy in the top 0-2 cm because it may be that the storage component for this depth will be highly significant. Residual calculation of G from the energy balance suggests enormous soil heat/water fluxes which can get as high as 450 W/m². Computing G from the soil temperature and soil water content data results in G fluxes ranging between 50 – 150 W/m². While this seems reasonable, these calculations do not enable energy balance closure calculations. The energy balance closure provides some measure of the reliability of the LE and H fluxes.

We have learned that the measurement of the soil heat fluxes, G, is critical and will require substantially more effort to quantify this term. This implies that any model (the ET-toolbox for example) that estimates G from a wet sand surface as a simple fraction of the net radiation is woefully inadequate. This is due to the complicating effect that water plays in modifying the thermal conductivity of the soil which in turns complicates the G flux estimation. When dealing with a dynamic surface (wet/dry sand, lateral water feeding into the matrix, variable meteorological conditions, etc) it is difficult to reconcile soil heat flux with the other 3 terms of the energy balance. We believe that there are phase lags for soil temperatures, soil heat storage, and soil heat flux. In an environment where the size and shape of the sand bars changes dramatically each day, daily images of the footprint for the EC sites is needed to ensure proper interpretation of the turbulent energy fluxes for heat and water. The soil heat flux component is highly variable and for this location can be very large. An important component of a follow-on effort would be to model G as a function of variable soil water content/soil conductivity/soil temperature/radiation loading/phase lags with LE, H and Rn. This may be too complicated because of the variability in time and space for the ET toolbox. However, it would be useful to determine how much can be accomplished with routine meteorological measurements in order to incorporate it into the ET Toolbox.

This campaign has gathered data for one end of the spectrum, very wet. Also needed is data for a more typical condition, extended exposed sand surfaces that are not continually saturated; that gets wet from a good afternoon/night shower (monsoon or otherwise) but not so much that the river rises and substantial continuity of the eddy covariance footprint is lost.
Lastly, a future campaign should incorporate changes to the lidar scanning protocols to optimize high resolution scanning over the islands and nearby river. Knowing that any differences will be small, the system can be adjusted to better measure the area and likely concentrations.

VII. ACKNOWLEDGMENTS

The authors gratefully acknowledge the assistance and funding of the U.S. Bureau of Reclamation. The assistance of the Middle Rio Grande Conservancy District and in particular that of David Gensler is greatly appreciated. The nature center flux tower was erected by Kristi Arsenault (NASA) and brought to life and maintained by Jan Kleissl (UCSD). We appreciate access to their data.
VIII. BIBLIOGRAPHY


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Appendix: List of Variables/Parameters

\[ A = \text{Total available energy } (Rn - G) \quad (\text{W/m}^2) \]
\[ A_s = \text{Available energy at the soil surface} \quad (\text{W/m}^2) \]
\[ c_p = \text{Specific heat of air} \quad (1.005 \text{ kJ/(kg - °C)} \text{ for dry air}) \quad \text{MJ/(kg - °C)} \]
\[ C_c, C_s = \text{Resistance combination equations} \]
\[ d_o = \text{Displacement height} \quad (\text{m}) \]
\[ D = \text{Vapor pressure deficit in the air at the reference height} \quad (\text{kPa}) \]

\[ D = e_{sat} \left(1 - \frac{RH}{100}\right) \quad (\text{kPa}) \quad (A-9) \]

\[ D_0 = \text{Vapor pressure deficit in the canopy} \quad (\text{kPa}) \]

\[ D_0 = D + \frac{\Delta A - E_c (\Delta + \gamma) r_{aa}}{\rho c_p} \quad (A-11) \]

\[ e_{sat} = \text{Saturated water vapor pressure at the air temperature} \quad (\text{kPa}) \]

\[ e_{sat}(T) = 0.6108 \exp \left[ \frac{17.27 T_{air}}{(T_{air} + 237.3)} \right] \quad (\text{kPa}) \quad (A-3) \]

\[ E_c = \text{Evaporation from the closed canopy} \quad (\text{W/m}^2) \]
\[ E_s = \text{Evaporation from the bare substrate} \quad (\text{W/m}^2) \]
\[ E_t = \text{Total evaporation} \quad (\text{W/m}^2) \]
\[ G = \text{Soil heat flux} \quad (\text{W/m}^2) \]
\[ LAI = \text{Leaf Area Index} \]

\[ LAI = 24 \text{ height of the canopy} \quad \text{for clipped grass} \]
\[ LAI = 5.5 + 1.5 \ln(\text{height of the canopy}) \quad \text{for alfalfa} \quad (A-8) \]

\[ P = \text{Atmospheric pressure (sea level 101.4 kPa)} \]
\[ P = 101.3 \left[ \frac{293 - 0.0065z}{293} \right]^{5.256} \]  

\( r_{sa} = \) Aerodynamic resistance \( (s/m) \)

\[ r_{sa} = \frac{\ln[(z-d_0)/z_{om}] \ln[(z-d_0)/z_{ov}]}{0.40^2 u} \]  

\( r_{ss} = \) Surface resistance of the land cover \( (s/m) \) (69 s/m for clipped grass 0.12 m high)

\[ r_{ss} = \frac{200}{LAI} \]  

\( R = \) Ideal Gas constant, 287 J/(kg - K)

\( RH = \) Relative Humidity \( \) (no units)

\[ RH = relative \ humidity = 100 \frac{e}{e_{sat}} \% \]  

\( Rn = \) Net long and short wave radiation \( \) (W/m\(^2\))

\( T_{air} = \) Air temperature \( (°C) \)

\( T_{air\ absolute} = \) Absolute air temperature \( (K) \)

\( u = \) wind speed \( (m/s) \)

\( z_{om} = \) Roughness height for momentum \( (m) \)

\( z_{ov} = \) Roughness height for water vapor transport \( (m) \)

\( \alpha = \) Priestley-Taylor coefficient, a constant often taken to be 1.26

\( \gamma = \) Psychrometric constant \( \) (mbar/°C)

\[ \gamma = \frac{c_p P}{0.622 \lambda} = 0.0016286 \frac{P}{\lambda} \] (kPa/°C)  

\( \Delta = \) Slope of the saturated water vapor pressure curve versus temperature \( (kPa/°C) \)
\[ \Delta = \text{slope of } e_{sat}(T) = \frac{4098 \ e_{sat}}{(T_{air} + 237.3)^2} \quad (kPa/\degree C) \quad (A-2) \]

\[ \lambda = \text{Latent heat of vaporization (MJ/kg)} \]
\[ \lambda = 2.501 - 0.002361 T_{air} \quad (MJ/kg) \quad (A-5) \]

\[ \rho = \text{Air density (kg/m}^3) \]
\[ \rho = \frac{P}{R \ T_{air\ absolute}} = \frac{3.486 \ P}{273.2 + T_{air}} \quad (kg/m^3) \quad (A-10) \]

\[ \rho_w = \text{water density (1000.0 kg/m}^3) \]

2. Priestley-Taylor Model

\[ E_t = \alpha \ \frac{\Delta A}{\Delta + \gamma} \quad (A-1) \]

3. McNaughton-Black Model

\[ E_t = \frac{c_p \ \rho \ D}{\gamma \ r_{cs}} \quad (A-6) \]

4. Penman Model

\[ E_t = \frac{\Delta A}{\Delta + \gamma} + \frac{73.64 \ \rho_w \ \gamma (1 + 0.54u) \ D}{\Delta + \gamma} \quad (A-11) \]

5. Penman-Monteith Model

\[ E_s = \frac{\Delta \ A_s + \rho \ c_p \ D_0 / r_{sa}}{\Delta + \gamma (1 + r_{ss} / r_{sa})} \quad (A-11) \]
\[ E_c = \frac{\Delta (A - A_d) + \rho \ c_p \ D_0/r_c}{\Delta + \gamma (1 + r_c/r_c)} \]  

(A-11)