DEVELOP A REMOTE SENSING TOOL TO ESTIMATE EVAPORATIVE LOSS FROM RESERVOIRS

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INTRODUCTION

According to the U.S. and Mexico water delivery treaty of 1944 (Treaty series 944. 2008.), the U.S. needs to deliver a percentage of the water stored in Elephant Butte Reservoir to Mexico every year, which in full water years is 60,000 ac-ft. In years of drought this was to be reduced according to the storage in the reservoir. Elephant Butte Reservoir is located in south central New Mexico (Latitude: N33:20:09, Longitude: W107:10:56, Elevation: 1395 m, Figure 1). The reservoir is approximately 4 miles (6.4 km) wide and 40 miles (64 km) long. The capacity of Elephant Butte reservoir is estimated to be over two million acre-feet (2.4 billion m^3). Evaporation from the reservoir is estimated to be as much as 1/3 of the approximate average inflow, which would be approximately 250,000 acre-feet per year (280 million m^3 per year) (Herting , et al. 2004).

Due to drought conditions from 1992 to 2002, Mexico did not receive the amount of diversions at El Paso, TX that they were entitled to and below El Paso, Mexico did not deliver the water to the Rio Grande river specified in the 1944 treaty. The deficit reached 1.5 million acre-feet at its highest point, costing U.S. agricultural producers in the Rio Grande Valley $1 billion. Part of the water delivery problems for both countries was the amount of water being used by reservoir evaporation in the upstream storage reservoirs. Consequently, the need exists to quantify the amount of water lost by evaporation to better explain why the storage in the reservoirs was considerably lower than predicted by inflow gauging stations.

Reservoir evaporation measurements from an inflow–outflow water balance method, pan measurement method, or eddy covariance method are time- and labor-intensive (Herting , et al. 2004; Sammis , et al. 2004; Linacre 1994). Additionally, the accuracy of the water balance method may be affected by bank storage and release, silting, percolation, and precipitation (Heather , et al. 2004). Pan E measurements assume that E is uniform over the entire surface area of the reservoir and the pan / reservoir

evaporation coefficient is correct. Pan E measurements are affected by extra heat conductance through the walls of the pan (Herting , et al. 2004; Linacre 1994) which results in a reservoir pan coefficient around 0.7. The accuracy of the eddy covariance measurements can be affected
by environmental factors such as fetch distance and wind direction, etc. In addition, a point E measurement cannot provide accurate spatial estimates over reservoirs that exhibit underwater currents, upwelling, and significant changes in depth. Furthermore, Mexico (or U.S.) farmers may doubt the E data provided from the U.S. (or Mexico) because the data are not reported by their own government. There is a need to create a method to measure E loss by remote sensing techniques that can be used internationally.

Remote Sensing Methods To Estimate ET

Different methods have been developed to estimate spatial ET on land or water, based on satellite data (Courault , et al. 2003). There are two main methods: direct and indirect. Direct methods mainly use thermal infrared data (TIR) and the energy budget equation. Indirect methods use the assimilation procedure and Soil-Atmosphere Transfer models. These methods use different wavelength data and obtain ground surface characteristics such as albedo, emissivity, and leaf area index (Courault , et al. 2003).

Direct Methods

The Direct Simplified Methods, which are empirical methods, are often used to estimate ET. The methods assume the daily ET is linearly related to the cumulative temperature difference, T_s-T_a (surface temperature minus the air temperature) (Courault , et al. 2003). On a local scale, accuracy could reach 85–90% (Steinmetz , et al. 1989). However, if the method is used at the regional scale, the accuracy will be around 70–80% because the input parameter (air temperature) must be interpolated from local measurement and the satellite remote measured surface temperature may be as must as 1.5- 2 degrees C in error.

SEBAL is a method of computing E as a residual in the energy budget. It was developed by Bastiaanssen , et al. (1998). It combines empirical and physical parameterization. Inputs include local weather data (mainly wind speed) and satellite data (radiance). From the input data, the R_n (net solar radiation), NDVI (normalized difference vegetation index), albedo, roughness length, and G (soil heat flux) are calculated. The sensible heat flux is calibrated by contrasting two points (wet, well-irrigated vegetation and dry ground). Then the ET is calculated as the residual of the energy budget (Bastiaanssen , et al. 1998). The accuracy can be 85% on a daily basis and 95% on a seasonal basis (Bastiaanssen , et al. 2005) at a single site. Based on the contrast between wet and dry areas, similar models such as SEBI, -S-SEBI and SEBS were developed (Menenti and Choudhury 1993; Roerrink , et al. 2000; Su 2002).

In the residual models, a two-source model (Kustas and Norman 2000) divides the energy calculation into two parts. One is the canopy and the other is the soil. The model estimates ET with an accuracy of about 90% (Kustas and Norman 2000). However, the model is more complicated than SEBAL, and accurate surface temperature data are needed, which requires better atmospheric correction algorithms than are currently available.

INDIRECT METHODS

Indirect methods deal with soil and plant energy exchange with the atmosphere with a fine time step of 1 s to 1 hr (Courault , et al. 2003). Indirect methods accurately describe crop functioning, and can allow access to intermediate variables such as soil moisture and LAI (leaf area index), which are related to the physiological and hydraulic processes that can be linked to other meteorological and hydrologic models (Courault , et al. 2003).

RESEARCH OBJECTIVES

The general objective of the project is to develop a remote sensing tool to estimate evaporation (E) loss from reservoirs to aid US-Mexico border water delivery management.

RESEARCH METHODOLOGY/APPROACHES

A Remote Sensing ET (RSET) model for plants was developed and validated (Wang , et al. 2005). This model was developed to use ASTER data and has a 90-m resolution of ET output. Because ASTER data are available only on a 16-day time scale, the model was modified to use the MODIS L1B data to output 1-km land surface ET on a daily time scale.

The following material first briefly describes the RSET model. Then, the modification for MODIS L1B data is described.

RSET MODEL THEORY

The RSET model general flowchart is shown in Figure 2. The model inputs are satellite data (ground surface reflectance and temperature) and local weather data (solar radiation, humidity, and wind speed). ASTER provides temperature and reflectance data. The model then calculates NDVI, the soil heat flux (G) and sensible heat flux (H). Finally, the model outputs the spatial ET (mm d^{-1}) according to the energy budget equation.

The method uses the energy budget equation to calculate each pixel λET_{ins} (instant latent heat loss) at the time of the satellite overflight.

$$\lambda ET_{ins} = Rn - G - H \qquad (1)$$

where λET_{ins} is the instant latent heat loss (W m^{-2}), λ is the latent heat of vaporization of water (J kg^{-1}), ET_{ins} is the instant hourly evapotranspiration (mm hr^{-1}, or kg hr^{-1} m^{-2}), which is calculated as a residual of the energy budget, R_n is net solar radiation (W m^{-2}), G is soil heat flux into the soil (W m^{-2}), and H is sensible heat into the air W m^{-2}); R_n is calculated according to the local solar radiation data (R_s, W m^{-2}) (Walter , et al. 2002).

$$R_n = 0.95(1-\alpha)R_s - R_{nl} \quad (2)$$

Here, R_{nl} is net long-wave radiation (W m^{-2}), α is albedo, and α is calculated by the equation in (Liang 2000) from ASTER surface reflectance data.

$$\alpha = 0.484\alpha_1 + 0.335\alpha_3 - 0.324\alpha_5 + 0.551\alpha_6 + 0.305\alpha_8 - 0.367\alpha_9 - 0.0015 \quad (3)$$

Here, α_i is the reflectance for ASTER data band i.

According to (Walter , et al. 2002),

$$R_{nl} = 277.8\sigma T_s^4(0.34 - 0.14\sqrt{e_a}) \quad (4)$$

where T_s is the mean absolute surface temperature (K), which is obtained from the satellite data, σ is theStefan-Boltzmann constant (2.042x10^{-10} MJ K^{-4} m^{-2} hr^{-1}), and e_a is the actual vapor pressure (kPa),

$$e_a = \frac{RH}{100} e_s(T_a) \quad (5)$$

Here, $e_s(T_a)$ is saturation vapor pressure (kPa), T_a is air temperature (ºC) and RH is the relative humidity (%).

$$e_s(T_a) = 0.6108\exp(\frac{17.27T_a}{T_a + 237.3}) \quad (6)$$

$$T_a = T_s - dT - 273 \quad (7)$$

dT is the difference between surface temperature and air temperature (K, equation 15).

Based on (Bastiaanssen , et al. 1998),

$$G = c \times R_n \quad (8)$$

where c is a coefficient

Based on data from Clothier , et al. (1986), Choudhury (1989), Kustas and Daughtry (1990), Van Oevelen (1993) and Bastiaanssen , et al. (1998), the following equation was obtained (R^2 =0.66, Figure 3).

$$c = -2.70NDVI^4 + 3.98NDVI^3 - 1.64NDVI^2 - 0.11NDVI + 0.41 \quad (9)$$

where NDVI is the normalized difference vegetation index calculated from satellite data.

NDVI is calculated as the following:

$$NDVI = \frac{\alpha_3 - \alpha_2}{\alpha_3 + \alpha_2} \quad (10)$$

where α_3 and α_2 are the reflectance data of bands 3 and 2 respectively.

For the sensible heat flux calculation, two pixels are chosen from the satellite data. One pixel is a wet pixel that is a well-irrigated crop surface with full cover and the surface temperature (T_s) is assumed to be close to air temperature (sensible heat flux H is assumed to be 0). The second pixel is a dry, bare agricultural field where λET_{ins} is assumed to be 0. The two pixels tie the calculations for all other pixels between these two points.

At the dry pixel, assume λET_{ins}=0; then, according to equation 1,

$$H = R_n - G \tag{11}$$

Then we can get dT_{dry}, which is dT at the dry spot (K) according to the following equations (12-14),

$$H = \frac{\rho \times c_p \times dT}{r_{ah}} \tag{12}$$

Here, ρ is the air density (mol m^{-3}), c_p is the specific heat of air (29.3 J mol^{-1} K^{-1}), dT is the near surface temperature difference (K), and r_{ah} is the aerodynamic resistance to heat transport (s m^{-1}), where

$$r_{ah} = \frac{\ln(\frac{z_2}{z_1})}{u^* k} \tag{13}$$

Here, z_1 is a height just above the zero plane displacement height of a plant canopy, set to 0.1 m for each pixel, z_2 is the reference height just above the plant canopy, set to 2 m for each pixel, u* is the friction velocity (m s^{-1}), and k is the von Karman constant (0.4). We calculate u* from the observed windspeed as

$$u^* = \frac{u(z)k}{\ln(\frac{z-d}{z_m})} \tag{14}$$

where u(z) is the wind speed at height of z, d is the zero plane displacement height (m), and z_m is the roughness length (m) (Campbell and Norman 1998). The calculations of these two variables are described in the paragraph after Eq. (15). From Eqs. (11-14) and the input data, dT_{dry} can be calculated. At the wet spot, we assume H=0 and dT_{wet}=0 (dT at this wet spot). Then according to the surface temperature at the dry and wet spots ($T_{s,dry}$ and $T_{s,wet}$, both in K), we can get one linear equation for each pixel,

$$dT = \left(\frac{dT_{dry} - dT_{wet}}{T_{s,dry} - T_{s,wet}}\right) \times T_s - \left(\frac{dT_{dry} - dT_{wet}}{T_{s,dry} - T_{s,wet}}\right) \times T_{s,wet} \tag{15}$$

With these calculated values of dT, the H at each pixel can be calculated according to Eqs. (12-14). The value of u* can be solved for each pixel by assuming that at 200 m above the surface the wind speed is the same. The wind speed at 200 m can be calculated from the weather station data (Equation 14). For example, assuming that the wind speed (u(z)) and measurement height (z), roughness length (z_m) and zero plane displacement height (d) are given at a weather station, Eq. (14) can be solved for u* at the weather station. Then, the wind speed at 200 m can be solved by using Eq. (14) again, given u_*, z=200 m, z_m and d. The parameter d in Eq, (14) for other pixels is set to 0 because it is negligible when z=200 m. The z_m for each pixel is calculated by a regression equation according to the pixel NDVI value. The equation is obtained by fitting z_m to a linear equation in NDVI. For example, if we know that pecan has z_m =1.2 m and NDVI =0.57, alfalfa has z_m =0.07 m and NDVI =0.42, and a bare agricultural field has z_m =0.003 m and NDVI =0.18, then we can obtain a regression equation for z_m (Figure 4).

Because atmospheric stability may have effects on H, an atmospheric correction is conducted (Figure 5). First, u^* and wind speed at 200 m at the local weather station are calculated. Then, z_m, u^*, and dT for each pixel are computed. The values of r_{ah} and H without the atmospheric correction can then be obtained.

For atmospheric correction, the stability parameter, or Obukhov length, L (m), is calculated. Using the stability parameter, u^*, r_{ah}, and H are corrected. Then an iteration is conducted for the L, u^*, r_{ah}, and H calculations until H does not change more than 5%. The correction equations are as follows (Campbell and Norman 1998; Stull 2001).

$$L = -\frac{{u^*}^3 T_s}{kgH} \tag{16}$$

When L<0, H is positive and heat is transferred upwards under unstable conditions; when L>0, H is negative and heat is transferred downwards under stable conditions; when L=0, no heat flux occurs under neutral conditions. Because the daytime satellite overflight occurred at local noon time at our Las Cruces study areas in New Mexico, the atmosphere should have been unstable. Thus, when L>0 (stable) occurred, we forced L=∞ (neutral).

The momentum correction term is

$$\varphi(\frac{z}{L}) = 0 \qquad \text{for L=}\infty \tag{17}$$

$$\varphi(\frac{z}{L}) = -2\ln(\frac{1+\beta}{2}) - \ln(\frac{1+\beta^2}{2}) + 2\tan^{-1}(\beta) - \frac{\pi}{2} \qquad \text{for L<0} \tag{18}$$

$$\beta = [1-15(z-d)/L]^{0.25} \tag{19}$$

$$u* = \frac{ku(z)}{\left[\ln(\frac{z-d}{z_m}) + \varphi(\frac{z}{L})\right]} \quad (20)$$

We use z=200 m and d is taken as negligible (d=0).

The correction term for the heat transfer is

$$\psi(z) = 2\ln(\frac{1+\beta_z^2}{2}) \text{ for L<0} \quad (21)$$

$$\psi(z) = 0 \qquad \text{for L=}\infty \quad (22)$$

$$r_{ah} = \frac{\ln(\frac{z_2}{z_1}) - \psi(z_2) + \psi(z_1)}{u^* k} \quad (23)$$

After H is corrected by the atmospheric effects, λET_{ins} for each pixel is calculated using Eq. (1). The daily ET (ETdaily, mm d^{-1}) is calculated by assuming that the evaporative fraction is constant over the day:

$$ET_{daily} = \frac{\lambda ET_{ins}}{\lambda ET_{r,ins}} ET_{r,daily} \quad (24)$$

where $ET_{r,daily}$ is the daily reference ET for well-irrigated alfalfa. The $ET_{r,daily}$ can be obtained from the FAO Penman-Monteith equation (weather.nmsu.edu). The value of λET_{ins} (W m^{-2}) is the instantaneous λET for a well-irrigated alfalfa field calculated from Eqs. (1-8), using α=0.23, c=0.04, and T_s= $T_{s,wet}$).

MODEL THEORY FOR MODIS DATA

The MODIS model basically used all the same equations and algorithms as for the ASTER except for NDVI, albedo, temperature and reflectance calculations. For ASTER data, NASA provides the surface temperature and reflectance products already. Therefore, they were not calculated in the ASTER model.

Although MODIS data provides reflectance and temperature data, the products have geo-registration problems, in which a pixel in a product may not correspond to the pixel in another product even though the geo-location (longitude and latitude) values of the two pixels are the same (Sung-Ho Hong, personal communication, New Mexico Technology University, http://infohost.nmt.edu/~hong/). Therefore in this study only the raw data from the MODIS sensor were used. These data are stored in the MODIS L1B files. Reflectance and surface temperature were calculated from these input data.

Follow those calculations, the algorithm for the RSET previously described was used to calculate ET, except that the NDVI and albedo algorithms were modified because the bands used in these calculations from the ASTER data set are not available from the MODIS data set.

REFLECTANCES CALCULATED FROM MODIS L1B DATA

When using MODIS data (L1 B 1-km data), land surface reflectances (inputs for albedo and NDVI calculation) can be calculated from bands 1, 2, 3, 4, 5 and 7. Bands 1 and 2 are the 250-m aggregated 1-km bands. Bands 3, 4, 5, and 7 are the 500 m-aggregated 1-km bands. (MODIS Level 1B Product User's Guide 2005):

$$r_i = reflectance\_scale\ (SI - reflectance\_offset\ ) \qquad (25)$$

where r_i (unitless) is the reflectance of band i, SI (scaled integer) is the raw L1B data value for the corresponding pixel, and reflectance_scale and reflectance_offset are the conversion factors to convert SI to reflectances. The L1B data product provides the conversion factors.

GROUND SURFACE TEMPERATURE CALCULATED FROM MODIS L1B DATA

Bands 31 and 32 in L1B data are the 1-km thermal bands. They can be used to calculate temperature after calculating the radiance value. Using the radiance for bands 31 and 32 (MODIS Level 1B Product User's Guide 2005), we have:

$$L = radiance\_scale\ (SI - radiance\_offset\ ) \qquad (26)$$

where L is the radiance (W m^{-2} μm^{-1} sr^{-1}), SI is the raw L1B data value, and L1B data provide the radiance conversion factors (radiance_scale and radiance_offset).

The land temperature (T_s, K) can be calculated based on the Eq. (6.6) (Planck's law) in Morse , et al. (2000) and Eq. (2) in Jiménez-Muñoz and Sobrino (2003). Because there are two thermal bands, 31 and 32, temperatures for both bands 31 and 32 ($T_{s,31}$ and $T_{s,32}$, K) are calculated and the average of the two temperatures ($T_{s,31}$ and $T_{s,32}$) are used as the to calculate ground surface temperature:

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$$T_{s,31} = \frac{1247.849}{ln(\frac{584.4937}{L_{31}} + 1)\varepsilon_0^{0.25}} \qquad (27)$$

$$T_{s,32} = \frac{1196.98}{ln(\frac{474.6834}{L_{32}} + 1)\varepsilon_0^{0.25}} \qquad (28)$$

where L_{31} and L_{32} are the radiances calculated from MODIS bands 31 and 32 and

ε_0 = 1.009 + 0.047 ln (NDVI) (29)

for NDVI >0. Otherwise, ε_0 (emissivity, unitless) is assumed to be 0.95—for example, for water (Van de Griend and Owe 1993).

NDVI Calculation

The NDVI (unitless) can be calculated using Eq. (4.2) on page 54 in Morse , et al. (2000), i.e.,

$NDVI=(r_2-r_1)/(r_2+r_1)$ (30)

where r_1 and r_2 are band 1 reflectance and band 2 reflectance, respectively.

Albedo Calculation

The albedo can be calculated as (Liang , et al. 2002):

$albedo =8\ 0.160r_1 + 0.291r_2 + 0.243r_3 + 0.116r_4+0.112r_5 + 0.081r_7- 0.0015$ (31)

where r_1, r_2, r_3, r_4, r_5, and r_7 are the corresponding reflectances of band 1, 2, 3, 4, 5 and 7.

MODIS Model ET Compared to ASTER's

First, the ASTER RSET model (Wang , et al. 2005) was evaluated for accuracy using plant ET data. Then, the MODIS land ET was compared with ASTER ET to evaluate the MODIS model ET.

The Site for ET Measurements by Eddy Covariance

The Las Cruces pecan crop study area is in southern New Mexico. Figure 6 shows the Las Cruces crop areas. A 5 ha pecan orchard was planted in 1970 at 10.0 m × 10.0 m tree spacing. A 5 ha alfalfa field is located 2.5 km southeast of the pecan orchard. From 2002 to 2004, in the pecan orchard and the alfalfa field, eddy covariance systems (LI-COR, Lincoln, NE) were set up to measure the ET from the canopies (Figure 6). The daily total ET was processed from the measurements. The processing method is the same as in Wang , et al. (2007). The daily ET was compared with the RSET ASTER model outputs.

ASTER Land ET and MODIS ET Comparison

Different locations at Las Cruces, NM, USA were chosen to compare the ASTER and the MODIS model land ET. The locations included bare ground, partially vegetated ground, and fully vegetated ground.

The ET from ASTER data is at 90-m resolution. ASTER calculations aggregated to 11 pixels in each direction to get equivalent 1-km data. The average ET was compared to the corresponding daily 1-km MODIS ET.

ASTER data sets for four days were used. They were taken on June 8, 2005, September 7, 2003, May 18, 2003, and September 4, 2002.

CALIBRATION AND VALIDATION

EVAPORATION DATA

ROOSEVELT RESERVOIR

In order to apply the remote sensing calculation to a reservoir the *c* coefficient in Eq. (8) for calculation of G into a reservoir must be derived. By measuring the reservoir evaporation by a water balance technique, the sensible heat flux from remote sensing data, and net radiation from solar radiation, the energy balance equation can be solved for heat flux G into a reservoir. The water balance data as inflow, outflow and change in storage were acquired for the Roosevelt Reservoir system of reservoirs in the Salt River Project in Arizona (Hydrologist of Salt River Project, personal communication) for the time period 1973-2006. Roosevelt Reservoir (33°40'34.30"N, 111° 7'35.28"W) has similar physical characteristics to Elephant Butte Reservoir (Figure 7) - the same surrounding desert vegetation, reservoir turbidity, and reservoir temperature. Consequently, data from this reservoir were used to derive the empirical heat flux function G as a fraction of net radiation. The function was then used in the remote sensing calculation to calculate E from Elephant Butte Reservoir. The scaling factor *c* in future work will be related to the turbidity and temperature of the reservoir. The values currently reported in the literature are for clear cold reservoirs and are not appropriate for reservoirs at lower elevations with high turbidity in the southwest.

ELEPHANT BUTTE RESERVOIR

The E loss (mm d^{-1}) data at the Elephant Butte Reservoir was used to evaluate the remote sensing model by comparing the daily reservoir evaporation to that calculated from the long term inflow outflow change in storage water balance of the reservoir.

The Bureau of Reclamation at El Paso, TX provided the daily storage and surface area data of Elephant Butte Reservoir. The USGS Website, usgs.gov, provided the daily flow rate of inflow (USGS 8358400 San Marcial, NM) and outflow (USGS 08361000 Rio Grande below Elephant Butte Dam, NM) for the reservoir for the time period 1977-2006.

E WATER BALANCE DATA CALCULATION

E (mm day^{-1}) data is calculated based on inflow (Q_{in}, $m^3\ s^{-1}$), outflow (Q_{out}, $m^3\ s^{-1}$), and storage (V, m^{-3}), and the water body area (S, m^2).

$$E = (3600 \times 24 \times Q_{in} - 3600 \times 24 \times Q_{out} - \Delta V)/S \qquad (32)$$

where ΔV is the daily storage change.

Data that resulted in negative values or values of E greater than 20 mm d^{-1} were assumed in error and discarded. Monthly averages were then calculated for the time period used in the analysis.

DISTINGUISHING RESERVOIR AREA FROM LAND AND CLOUD AREA

NDVI and surface temperature are used to distinguish reservoir from land and cloud. When NDVI is close to 0 or negative, then the pixels are deemed to be water bodies or clouds. However, the temperature of water bodies is much higher than the temperature of clouds and this separation technique was used in the MODIS algorithms calculations to separate the pixel calculations into the equation governing vegetation vs. reservoir evaporation. .

R_N CALIBRATION

Weather stations usually measure R_s, so that R_n for the reservoir surface needs to be calculated. The 2005 measured daily R_s and R_n data for Elephant Butte Reservoir in (Almy 2006) were used to deduce a regression relationship between the two variables (Figure 8). Based on Eq. (2), the intercept in the regression equation is the long wave radiation. Because the regression equation in Fig. 8 is for daily radiation, the intercept was divided by 24 to derive the hourly radiation function.

$$R_n = 0.887R_s - 2.6497/24 = 0.887R_s - 0.11 \quad (33)$$

G CALIBRATION

Calculation of G for MODIS was calibrated using the Roosevelt Reservoir data and Eq. (1). R_n was calculated based on Eq. (33). H was calculated by the ASTER RSET model for the calibration. The value of λET_{ins} was calculated by scaling the corresponding monthly $\overline{E}$,

$$E_{daily} = \overline{E} \times \frac{ET_{rdaily}}{\overline{ET_r}} \quad (34)$$

$$E_{ins} = E_{daily} \times \frac{ET_{rins}}{ET_{rdaily}} \quad (35)$$

Here, E_{daily} is the daily evaporation (mm d^{-1}), $\overline{ET_r}$ (mm d^{-1}) is the monthly average ET_r during 1973-1977 for the Roosevelt Reservoir area (data at Desert Ridge weather station in Arizona was used), and E_{ins} is the hourly instant evaporation (mm h^{-1}, or kg h^{-1} m^{-2}),

$$\lambda ET_{ins} = \lambda E_{ins} \quad (36)$$

Using Eq, (1), G can be solved. In each month of 2005 and 2006, May-September, 5 G data points were calculated. Then the corresponding values of G/R_n were calculated and the average G/R_n was obtained.

Temperature Correction of Cold Spots with Elevation

We also tried to determine the effects of elevation and consequent adiabatic cooling of air on the land surface temperature for cold spots. This information can be used for temperature correction for a cold spot if the spot has a different elevation from the area of E or ET calculations. The Sierra Mountain region of California was chosen which had elevations varying from 165 m – 2310 m (Fig. 9). The region corresponds to a square area whose top left corner is given by latitude and longitude coordinates 39˚27´03.24˝N, 121˚04´17.87˝W respectively and lower right corner is given by 38˚20´10.35˝N, 119˚49´46.83˝W.

Several days in the months of April, May and June were selected because during those months moisture in the soil profile has been recharged from snow melt and moisture does not limit ET. For these locations temperature values were recorded in a table. The lowest elevation point was noted, and temperature at this point was used to calculate elevation difference for all locations by subtracting it from other elevation values. The temperature difference and elevation difference data were represented by a scatter plot and a regression was obtained.

Problems/Issues Encountered

There were some days when there were clouds over the area of interest. Sometimes, a cold spot may not be available around an area of interest. A weather station may not be available around an area of interest.

Currently, downloading satellite and weather data are not automated completely, nor is processing the data. For example, an operator must manually choose a cold and a hot spot; then, E or ET can be processed. The automation need to be completed in the future.

Research Findings

Temperature Correction of Cold Spots with Elevation

Figure 10 shows the effects of elevation on the land temperature of cold spots. It shows every 1000 m difference of elevation, two cold spots at the different elevations can have 4.4 K difference.

RESERVOIR EVAPORATION FROM WATER BALANCE

Figure 11 shows the monthly average evaporation for Elephant Butte and Roosevelt Reservoirs from the water balance methods. As expected, the evaporation rate follows the solar radiation curve, low in the winter and increasing reaching a maximum at the summer solstice, June 21. Roosevelt Reservoir had higher evaporation in the winter time than Elephant Butte Reservoir because the air temperature in the surrounding desert at Roosevelt Reservoir is higher than at Elephant Butte Reservoir; contributing more sensible heat advection transferred to Roosevelt Reservoir compared to Elephant Butte Reservoir. In the summertime (June-August) the evaporation was about 5.2-5.8 mm d^{-1}.

MODELED E AND ET

Figures 12 and 13 show modeled ET maps for the Las Cruces area and the Elephant Butte Reservoir . Figure 12 was from ASTER data and Fig. 13 was from MODIS data.

The ET comparisons (Fig. 14) show that the ET values from MODIS and ASTER were very close. MODIS calculations were 3% lower than the composite ASTER calculated ET. Because the ASTER ET model was validated for plant ET (85% accuracy; Wang , et al. 2005), the model for MODIS plant ET also has about 85% accuracy for plant areas.

Table 1 shows the comparison of measured and modeled E for Elephant Butte Reservoir during summertime (June to August) in 2006 and August in 2005. The difference between the measured and modeled E was within 1.4 mm d^{-1} and the average bias is -0.24 mm d^{-1}. The average measured E was 5.6 mm d^{-1} in the summer time vs. 5.9 mm d^{-1} of the modeled.

CONCLUSIONS

A model for estimating evapotranspiration using MODIS data was developed and evaluated for both plant areas and water bodies. It has a 1-km spatial resolution and a daily temporal resolution. The remote sensing model is capable of estimating water body evaporation in summertime and capable of calculating evapotranspiration over land. For the summertime E estimate, the accuracy is within 1.5 mm day^{-1} and the average bias is only -0.24 mm day^{-1}. The evapotranspiration accuracy is about 85%. The average evaporation of Elephant Butte Reservoir in summer time was 5.6 mm day^{-1}. The model accuracy is acceptable and it is capable of aiding international water delivery management.

Reservoir G calculation was calibrated in this project. It is proportional to net solar radiation and the relation was calibrated using Elephant Butte Reservoir data. Cold spot temperature variation with height was obtained in this study.

Recommendations for Further Research

The model needs to be improved. When a vegetation cold spot is not available around an area of interest, then a calculated cold spot based on reference ETshould be used, Weather data from a weather station are inputs for this model. Improvements may be done to use less weather data or forgo using all ground data.

The model needs to be completely automated, including downloading and processing of data. In addition, the model can be modified to be a Web-based software package. Then, a person can just click at the website and get the information he/she wants.

The model was calibrated and validated for the US-Mexico border areas. It needs to be improved and evaluated for other areas.

The calculation of heat into the reservoir (G) should be improved and G should be resolved as a function of turbidity of a reservoir. The method of calculating the daily E (the scaling up of a midday energy balance to a daily energy balance) will need to be refined for the pixels over the reservoir because it may be different over the reservoir compared to the land.

Rresearch Bbenefits

This model can improve the international water delivery management to evaluate the evaporation loss from reservoirs. For example, we have trained the cooperator Mr. Ramiro Lujan at the Mexican Section of the International Boundary and Water Commission to run the model and estimate the evaporation loss of Mexican reservoirs (See the sub-report at the end of the main report).

In addition, this model may be used for reservoir or lake evaporation estimate for other areas. This model can also estimate other energy components at a reservoir or lake, namely, the heat flux to water and sensible heat flux to air.

The model can be used for estimating evapotranspiration on land surfaces also. This can help farmers, water managers, and researchers to know the water use by crops and other plants.

The model can be used to determine the savings in lake evaporation water loss that would occur if water is stored longer in upstream reservoirs at higher elevation which will have less lake evaporation through out the year.

Documentation for the model theory, installation and compilation procedures was prepared. More documentation for help in finding weather data was also prepared. A Website was made which provides access to the remote sensing tool and the documentation. The documents and Website enable other researchers and related users to access, use, and revise the model.
http://hydrology1.nmsu.edu/Lake%20evaporation/Lake%20Evaporation%20Remote%20Sensing%20Model.htm

Two journal manuscripts have been submitted for publication:

1.
Wang, J., T.W. Sammis, and V. P. Gutschick. The sensitivity and accuracy of a Remote Sensing Evapotranspiration algorithm (RSET) using Aster satellite data. Submitted to Remote Sensing of Environment.

2.
Wang, J., T.W. Sammis, and V. P. Gutschick. Review of satellite remote sensing use. Submitted to Journal of Applied Remote Sensing.

Two conference papers have been published and presented:

1.
Wang, J., T.W. Sammis, and V. P. Gutschick. A Remote Sensing Model Estimating Water Body Evaporation. 2008 International Workshop on Earth Observation and Remote Sensing Applications. June 30-July 2, 2008. Beijing, China.

2.
Wang, J., T.W. Sammis, and V. P. Gutschick. A Remote Sensing Model Estimating Reservoir Evaporation. 2008 IEEE International Geoscience & Remote Sensing Symposium. July 6-11, 2008 | Boston, Massachusetts, U.S.A.

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SUB-REPORT FROM MEXICAN COOPERATOR:

IMPROVING WATER MANAGEMENT ALONG THE U.S.- MEXICAN BORDER: ESTIMATION OF WATER EVAPORATION AT A SURFACE WATER BODY USING A REMOTE SENSING MODEL

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NARRATIVE SUMMARY
Remote sensing techniques, models, and instruments are constantly developed and applied to many branches of human knowledge, providing multiple valuable services and products for social, agricultural, environmental, and hydrologic sciences.

During the last five decades there has been significant growth in the field of remote sensing, particularly in soil and water related sectors. Leading governmental agencies, academic institutions, and research organizations have created and developed useful remote-sensing models and techniques to improve the conservation and management of water resources.

A New Mexico State University (NMSU) team developed a satellite remote sensing model based on the Surface Energy Balance Algorithm for Land (SEBAL) method of energy budget, validated for the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) imaging instrument. It uses Moderate Resolution Imaging Spectroradiometer (MODIS) satellite data, land information from terrestrial weather stations, and water balance calibrations of evaporation to estimate water evaporation at surface water bodies.

This paper describes some runs and results of the NMSU model at an international water reservoir during the summer months of 2007. Final results indicate that the NMSU

model can provide useful information about evaporation water estimates. This model constitutes a valuable additional tool that could potentially contribute to improve water management along the U.S. - Mexican border.

INTRODUCTION

Available fresh water resources are steadily decreasing at many regions of the world, mainly due to the rise of human population growth at an unprecedented rate (Rogers 2008). One of the regions experiencing a constant increase in population and economic activities is the U.S. – Mexican border. An aggravating additional factor could be climate change. According to some climate prediction models, many parts of the planet will experience important climate variations with recurrent drought and water scarcity episodes, and, among these, the U.S. Southwest and north of Mexico will probably suffer severe and increasing water shortages (Christensen, J. H. at al. 2007). The combined effects of adverse climate change, growing human activities and population, would only exacerbate water stress at the region.

A recent episode clearly exemplifies the difficulties posed by water scarcity at the U.S. - Mexican border region. The drought experienced during the 1992-2005 period created a serious conflict of water allocation between the United States and Mexico that was resolved by the International boundary and Water Commission (IBWC) after complex binational negotiations and agreements (IBWC 2002, 2003). An important factor contributing to water delivery problems was the amount of water being lost through evaporation at reservoirs located along the international reach of the Rio Grande.

Satellite remote sensing is a major component of the methodology, with multiple and valuable applications upon diverse topics related to economic, social, geophysical, and environmental sciences. Since its inception and with subsequent developments, this technique has been a powerful tool to monitor and evaluate global surface changes caused by natural and anthropogenic activities and interactions. Specific applications of remote sensing models are: forest mapping, fire assessment, landscape and cultivated land classification, biomass mapping, crop type and production data gathering, water stress assessment, drought and flood impacts evaluations, location and extent of water bodies, etc.

Remote sensing can be used to monitor water loss by lake evaporation and water used by crops. Any measure to reduce or eliminate water losses along the border will contribute to preserving the existing limited water resources of the area, allowing managers to cover the expected water demands of incoming years for human and natural purposes, and ultimately, to reduce potential conflicts between the two countries. Better estimates of water evaporation at surface water bodies will improve the use and management of water along the U.S. - Mexican border. Remote sensing tools provide a useful tool to estimate lake evaporation and determine the water savings that can occur by storing water up stream in reservoirs at higher elevations where lake evaporation is less.

Remote sensing systems detect, analyze, and process the radiation emitted or reflected by objects or areas being subject to observation. Remote sensing data is processed and analyzed with computer software, and a number of commercial and open-source applications exist to solve the surface energy balance alghortihms for mapping regional lake evaporation and crop evapotranspiration using remote sensing data (Gowda , et al. 2007).

Other methods of estimating lake evaporation include the inflow-outflow water balance method (exposed-area method), pan measurement, and eddy covariance methods, which are time- and labor-intensive (Wang , et al. 2008). However, these traditional methods can not incorporate the spatial variability of a lake or reservoir evaporation (Wang , et al., 2008). Many of the past and current satellite remote sensing methods make use of spectral signatures or simple indices such as the Normalized Difference Vegetation Index (NDVI) and do not use a physically based algorithm (Wang , et al., 2008).

A NMSU team formed by Dr. Junming Wang, Ted W. Sammis, and Vince Gutschick developed a remote sensing model to calculate crop water use (i.e., evapotranspiration, ET) and water body evaporation (E) in a reliable, fast and practical way, making use of specific satellite information.

RESEARCH OBJECTIVES

The objectives of this project are:

To estimate water evaporation values at a surface water body located at the U.S. - Mexican border during the summer season of 2007, by using the remote sensing model developed by the New Mexico State University (NMSU) team.

To evaluate the model input data gathering process and the model running procedure.

RESEARCH METHODOLOGY/APPROACHES

This paper describes using the NMSU model to estimate water evaporation for a surface water body.

The basic hypothesis is that application of the NMSU water evapotranspiration model at a U.S - Mexican border surface water body, using MODIS satellite data and imagery, as well as land weather information, can provide useful water evaporation values to be used for better water management along the referred border.

One of the most appropriate and useful tools for forest health management, agricultural planning, and water management is the Moderate Resolution Imaging Spectroradiometer satellite data (MODIS), developed by NASA. MODIS data offer some useful characteristics: it has a 250 –1000 m spatial resolution and 1-2 day temporal resolution. MODIS data are provided freely and cover up to 36 spectral bands, more than many other satellites. The satellite data are available as raw data, and also as processed data, where the raw data is corrected for atmospheric effects and algorithms are used to produce specific products such as surface temperature, reflectance and images (Wang , et al., 2006).

The NMSU evaporation model using MODIS satellite data is based on a surface energy balance algorithm for land (Bastiannssen , et al. 1998). The model limits errors associated with surface temperature measurements from the satellite and is consequently a more operational useful model for determining ET than other methods (Wang , et al., 2008). The limitation is that a hot and a cold spot must be found in the satellite image. The model was modified in order to handle MODIS input and was calibrated and validated using reservoir data.

Model Theory

The model is based on the principle that water evaporation can be derived from energy balance. Energy balance is a function of net solar radiation (R_n), sensible heat flux (H), latent heat flux (water evaporation, λE), and Water Heat Flux (G). This can be expressed by the energy balance formulas:

$$\lambda \mathbf{E}_{ins} = \mathbf{R}_n - \mathbf{H} - \mathbf{G}, \text{ and}$$

$$\mathbf{ET}_{daily} = (\lambda \mathbf{ET}_{ins} / \lambda \mathbf{ET}_{r,ins}) \mathbf{x} \ \mathbf{ET}_{r,daily}$$

where λE_{ins} is instantaneous latent heat flux, ET_{daily} is the amount of daily evapotranspiration, $\lambda ET_{r,ins}$ is instantaneous reference latent heat flux for an cultivated field, and ET_{daily} indicates daily reference evapotranspiration, determined by the United Nations`s Food and Agriculture Organization (FAO) Penman-Monteith Method (Wang , et al. 2006), using data from the nearest weather station.

Net radiation (R_n) is a function of reflectance and solar radiation (R_s) as well as thermal radiation; it can be estimated from solar radiation alone, using an empirical regression. Soil (or water-body) heat flux (G) is a function of the normalized difference vegetation index, NDVI. Sensible heat is a function of NDVI, temperature, reflectance, solar radiation, and wind speed. For the summertime E estimate, the accuracy of the method is 1.5 mm/day. The evapotranspiration accuracy is 85 % (Wang , et al. 2008). All data, satellite imagery and land records, have to be obtained for the summer months of a given year.

The main objective of this model, namely, improving international water management, will be properly achieved if it is applied to water bodies located near or along the international boundary between the United States and Mexico. There exist several major reservoirs at the U.S.-Mexican border area. Amistad Reservoir, located along the Rio Grande, is one of the most important reservoirs along the U.S. - Mexican border. It is managed by the International Boundary and Water Commission between the U.S. and Mexico (IBWC).

Amistad Reservoir is located at the Rio Grande, 12 miles northwest of Del Rio, Texas. ($29^o28.18$ N and $101^o3.14$ W; surface elevation: 340 m (Fig. 15). The reservoir was formed in November 1969 by the construction of Amistad Dam, intended to provide flood control, water conservation, irrigation, hydroelectric power, and recreation to the area. It covers approximately 64,900 acres (263 km^2) and contains a water volume of about 5,658,500 acre–feet (6.98 km^3), with significant cultivated areas lying downstream along the Rio Grande (Fig. 16).

Data gathering process of the model

The NMSU model requires two types of data: satellite and land data. Satellite information input data are: surface temperature and reflectance. Land information input is solar radiation, humidity, and wind speed.

Satellite data

The MODIS web site provides periodic images taken by NASA`s Terra satellite of the entire Earth`s surface as it passes every day from north to south across the Equator in the morning. Satellite images for the Amistad Dam region were obtained by the NASA web site during the months of June, July, and August (summer time) of 2006 and 2007.

Satellite images showed cloudy skies above the selected area during most days of June, July, and August of both years, a fact that prevented pinpointing Amistad Dam. Also, satellite trajectories were not directly passing above Amistad Dam during most sunny days of the summer season of 2007, making impractical any temperature or reflectance readings for the selected area. During the summer of 2007, the days of good satellite images for Amistad Dam were the 4th and 14th of June, and the 5th and 7th of August.

Having selected the appropriate satellite images, all related 1-km resolution MODIS LB1 raw data were downloaded. The model automatically inputted the desired bands of the data. The model has user-friendly interfaces. As an example, a reflectance and temperature image of the site, with adjacent blank entries to be filled in with land weather information is shown in Fig. 17.

Land information

As mentioned earlier, the model requires land-specific information at or near the selected water body: solar radiation (L/hr), humidity (%), wind speed (km/h), and reference daily evapotranspiration (mm/day). Land information was obtained from two specific weather station Web pages.

Humidity and wind speed data were obtained from a weather station near the Amistad Dam. This station is a private weather station in the vicinity of Kickapoo Caverns at Brakettville, Texas, with Code Number MKCPT2, located approximately 37 miles from Amistad Dam in an east-north-east direction. Latitude/longitude coordinates are: 29.61 N, -100.47^{o} W.
(http://www.wunderground.com/weatherstation/ListStations.asp?selectedState=TX).
Kickapoo weather station does not provide solar radiation and ET_r data. Humidity and wind speed readings at MKCPT2 station for the selected days, are indicated in Table 2.

Solar radiation (R_s) and daily reference ET (ET_r) are parameters with small variations inside a latitude band encompassing Amistad Dam and Las Cruces, New Mexico, which are located about 3^o apart (Amistad Dam latitude: 29^o28.18 N, Las Cruces N.M. Latitude: 32.34^o N). Therefore, a weather station located at Las Cruces N.M. can provide reliable values for R_s and (ET_r). Solar radiation and ET_r data was provided by weather station Derry, with code number 65. Input data from this station was obtained for each selected date from the web sites: http://weather.nmsu.edu/cgi-shl/cns/uberpage.pl?selected=2 and http://weather.nmsu.edu/cgi-shl/cns/oldformat.pl, as shown in Table 3.

Calculation process
All above data were plugged into the ET Model, which performed the expected calculations.

The next step was to determine minimum and maximum temperature in a square area encompassing Amistad Dam. This was done by pinpointing "cold" and "dry" pixels near the dam, representing vegetation and bare soil temperatures. First, the "cold" pixel was pinpointed at the upper left and lower right sides of an imaginary square inside of which the dam was located. After pinpointing both sides of the rectangle, the data were fed into the program using a "setting cold point" command, which incorporates mean "coldest" temperature and site coordinates of the area into the model. A similar procedure was performed for "dry" pixels to find the mean "hottest" temperature. After plugging in all temperature data, the model calculates the evapotranspiration (ET) or E (Fig. 18).

Final Results
Water evaporation values on June the 4th and 14th, and August the 5th and 7th of 2007 were 6.08, 8.53, 5.67, and 5.11 mm/day, respectively. Final results indicate a mean evaporation of 6.35 mm/day for Amistad Reservoir during the summer sample days of 2007. The complete specific data set and final results for all dates of are shown in Table 4.

Site coordinates, temperature, and ET information were finally fed into a related program, the Model program, which yields a graphic image of the data, allowing a quick and visual comprehension of the numerical information. An example of the graphic image provided by the model is shown in Fig. 19.

PROBLEMS/ISSUES ENCOUNTERED
To obtain reliable data satellite information, good clear land satellite images from the selected area need to be obtained. Satellite orbits must pass near or right above the pinpointed area. In the summer, frequent cloudy sky conditions can reduce opportunities to obtain good satellite images of the selected Amistad reservoir site.

Cultivated or forested areas close or near the site of interest are needed in order to calculate cold points. Temperature readings would be difficult to obtain if there are not cultivated or forested areas nearby the site.

Weather stations with good solar radiation, daily ET_r, wind and humidity data are necessary to obtain reliable final results. If weather stations are not sufficiently close, computations can not be completed.

Pinpointing the selected site on the temperature image can be difficult if site coordinates are not well defined.

RESEARCH FINDINGS
During the summer season of 2007, cloudy skies were a frequent occurrence above the Amistad reservoir area, which lies along the U.S. – Mexican border.
Local weather conditions (i.e. clear skies) at the selected site (Amistad reservoir) allowed the use of satellite images during the days of 4 and 14 of June, and 5 and 7 of August of 2007. E values at Amistad Lake during the days of 4 and 14 of June, and 5 and 7 of August of 2007 were 6.08, 8.53, 5.67, and 5.11 mm/day, respectively.

CONCLUSIONS
Most of the existing evapotranspiration models require multiple data sources, involving extensive spatial and temporal information. A simplified and practical evapotranspiration model will contribute to enhance water evaporation estimates.
The NMSU ET Remote Sensing Model using MODIS data constitutes a valuable and practical tool to estimate water evaporation at surface water bodies. It provides useful practical numerical values about water evaporation.

The NMSU ET Remote Sensing Model could be used by U.S. and Mexican water management agencies to estimate evaporation at surface water bodies along the border; being a practical way to check water evaporation losses determined by regular approved methods, contributing to an improved water management at the area.

A binational operating protocol for the use of the ET Remote Sensing Model by U.S. and Mexican agencies could be developed. The method could be included into the official accepted procedures actually used to determine water evaporation losses along the U.S. - Mexican border.

RECOMMENDATIONS FOR FURTHER RESEARCH
The ET Remote Sensing Model requires minor adjustments and refinements to improve water evaporation estimates for surface water bodies located particularly in arid or desert areas where water evaporation is intense and cultivated or forested areas are scarce or nonexistent.

RESEARCH BENEFITS
Applying the NMSU ET Remote Sensing Model to determine water losses at surface water bodies along or near the US-Mexico border, would improve binational water management, contributing to a better use and conservation of the existing limited water resources of the region, and resulting in the benefit of its inhabitants.

Developing a binational protocol for the use of the model can enhance international cooperation among governmental agencies and academic institutions of both countries. The model represents an additional tool for policy and decision making by water management agencies of different countries.

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