INVESTIGATING GREAT SALT LAKE TEMPERATURE WITH THE ADVANCED VERY HIGH RESOLUTION RADIOMETER: INITIAL SURVEY AND CONSIDERATIONS

by

Erik T. Crosman

Meteorology 6410

Remote Sensing of the Envoronment

Department of Meteorology

The University of Utah

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TABLE OF CONTENTS

CHAPTER 1: INTRODUCTION

MOTIVATION

The temperature of Utah's Great Salt Lake (GSL) responds to regional weather and climate through the interplay of local heating and cooling, evaporation, passage of storms, and the distribution of precipitation throughout the watershed of its closed basin. In turn, the thermal characteristics of the Lake and nearby playas (salt flats) contribute to the development of lake breezes, salt breezes, lake-breeze fronts, and lake-effect snowstorms that affect the populace and water resources of northern Utah.

As discussed by Hostetler (1995), lakes continually respond to climactic conditions over broad spatial and temporal scales: lake temperature is governed by the balance of heat inputs and outputs and lake stratification depends strongly on the seasonal atmospheric forcing. Bates et all. (1993), Rodionov (1994b), Scott and Huff (1996), Grover and Sousounis (2002), and Laird and Kristovitch (2002) examined the sensitivity of the Great Lakes to atmospheric forcing. Bussieres and Shertzer (2003) and Rouse et all. (2003) have studied the impacts of atmospheric forcing over a broad range of scales upon the Great Slave Lake, Canada. Linkages between local and remote climate forcing have been noted for many terminal water bodies. Rodionov (1994a) studied linkages between climate forcing and anthropogenic influences upon the largest inland body, the Caspian Sea. Anthropogenic dessication of the Aral Sea combined with climate variability have been linked with regional climate changes in that region by Small et al. (2001ab). They used in situ observations and satellite radiometer data to infer increases in temperature of the Aral Sea from 1960-1997 in spring while the temperatures dropped significantly in the fall.

The immediate response of GSL temperature to atmospheric forcing can be quite clear. Rapid cooling of the lake is often observed with the passage of strong cold fronts. Since no long-term records of lake temperature exit, it has not been determined to what extent lake temperature is sensitive to regional and remote climate forcing. The most obvious interactions between GSL temperature and the overlying atmosphere are distinctly different in the warm season compared to the cool season. During the warm season, temperature contrasts arising from the differences in surface characteristics between the GSL/playas and the surrounding mountains and upland valley drive local winds toward the lake and playas at night and away from them during the day (Whiteman 1990; Segal et al. 1997; Whiteman 2000; Stewart et al. 2002; Horel 2003). Lake breeze and salt breeze fronts (Zumpfe and Horel 2005; Rife et al. 2002) develop sharp discontinuities in moisture, temperature and wind that affect local air quality and on occasion contribute to severe weather, such as the 11 August 1999 tornado in Salt Lake City (Dunn and Vasiloff 2001).

As discussed by Gwynn (2002), scientific research and awareness of the GSL and its environs as an important ecosystem has increased in recent years. While data on the physical state of the Lake have been collected manually for many years as part of the routine sampling programs undertaken by researchers at the United States Geological Survey, Utah Geological Survey, and Department of Natural Resources, the near surface temperature probe installed off Hat Island by John Horel in September 1998 was the first long-term automated record of lake temperature. As the lake level dropped in recent years, the sensor is now several kilometers from the nearest water. A similar sensor was installed in September 1999 off Gunnison Island, although it is also above water level now. Given the relatively poor spatial and temporal resolution of in situ observations of Lake temperature, there is significant value in using long-term records of advanced very high resolution radiometer (AVHRR) data from the NOAA polar-orbiting satellites to obtain a data set useful for estimating the variability of Lake surface temperature on a variety of temporal and spatial scales. Several decades of spatially high-resolution lake temperature data derived from satellite thermal radiances will advance knowledge of the physical processes that control the spatial and temporal variations in the surface temperature of the GSL and nearby playas and how these variations may in turn affect local weather and climate.

While the Great Salt Lake has been an "object of observation" from the beginning of remote sensing, no "comprehensive study of the lake through remote sensing" has been conducted to date (Zastrow and Ridd 2002). Spatial variability in Lake surface temperature has been examined for selected case studies and short field programs by several investigators (Baskin 1998; Zastrow and Ridd 2002; Rich 2002). Zastrow and Ridd (2002) contrasted thermal imagery from Landsat in 1996 and 1994 and noted a number of temperature patterns related to lake depth and freshwater sources. Rich (2002) used data from drifter bouys over a 48 day period during summer 1991 to assess surface currents and temperature. Strong easterly currents to the west of the gap between Antelope and Fremont Islands were attributed to a counterclockwise circulation arising from Ekman pumping. She hypothesized that prevailing southerly flow over the south arm of the Lake let to the Ekman pumping and also caused surface lake temperature to be warmer (colder) on the east (west) side of the Gilbert Bay during June 1991.

The difficulties encountered in remote sensing of surface temperature are not trivial. Thermal retrievals are complicated by instrument noise and calibration errors, atmospheric attenuation of emitted radiation, cloud contamination, and surface effects like water roughness and emisivity changes due to salinity and suspended particulate in the water (Robinson et al. 1984, Martin 2004, Yokoyama 1993, Campbell 1996, Kondratyev and Filatov 1999). Since the late 1970's, satellite-derived sea surface temperature (SST) algorithms have been developed and rigorously validated against bouy and ship in situ temperature data to obtain global satellite-derived SST's with biases between .1-.3 degrees C (McClain et all 1985, Li et all 2001, D. Oesch 2003). Regional studies have also validated satellite SST retrieval algorithms with varying results. In some cases the global algorithms were modified to improve retreivals in specific regions (Kizu and Sakaida 1995, Shenoi 1999), while in other cases the global algorithms remained effective when applied regionally (Mason et al. 1999, Li et al. 2001, Oesch 2003) In the GSL region, it is clear that the processing methodology appropriate for the open ocean or the coastal sea surface temperature may need to be adapted to obtain GSL temperature data of the highest possible accuracy.

The purpose of this thesis is to present an initial study of the suitability of using AVHRR lake temperature retreivals to study Great Salt Lake temperature. This study seeks to answer the following questions:

1). What physical mechanisms control the spatial characteristics of Great Salt Lake surface temperature observed in AVHRR imagery?
2). What are the sources and magnitudes of errors in estimating lake temperature from satellite radiance data?
3). Given our knowledge of error uncertainty in GSL surface temperature, what thermal features can we determine to be physical versus a construction of remote sensing limitations?

To answer these questions, AVHRR-derived images of Great Salt Lake surface temperature have been studied over several months in combination with weather conditions around the Great Salt Lake. In addition, a vast literature review has been conducted on possible factors influencing thermal remote sensing retrievals over the Great Salt Lake. An overview of previous studies of thermal characteristics of lakes using AVHRR imagery is also given. The theory and procedure of using AVHRR raw radiance data to derive lake temperature and apply cloud masks are discussed in Chapter 2. Chapter 3 gives a review of the meteorological and limnological factors influencing lake surface temperature. Finally, in Chapter 4 the possible impact of remote sensing considerations on lake temperature retrievals are presented. The study is briefly concluded and possible future work are outlined in Chapter 5.

PREVIOUS STUDIES USING AVHRR IMAGERY TO STUDY LAKE TEMPERATURE

Satellite-derived SST's have been rigorously validated from ship and bouy temperature data, and have been widely used since the launch of AVHRR/2 on NOAA-7 in 1981. Use of satellite-derived AVHRR data has been used more sporatically for use in monitoring lakes, but has gained more interest in recent years as the accuracy and availabiliy of AVHRR data set has increased. Several factors make AVHRR data over lakes less reliable than over the ocean. First, data may be less temporally consistent, as cloudy periods may completly exclude temperature data over a lake for days or even weeks at a time. Over the ocean, there are typically always regions that are cloud free due to its large surface area. Second, cloud masking is more difficult over lakes because uniformity tests applied over the ocean cannot be applied over lakes due to the high spatial variability of temperature within inland water bodies. Third, ice cover across large northern lakes blocks measurements in those areas during the winter months. Fourth, validation information for lakes is much less available than over the ocean, which makes it more difficult to quantitatively use the satellite-derived data. Finally, the effects of water turbidity and surface effects like waves and foam (and for the Great Salt Lake, salinity) on the water emissivity are somewhat uncertain over lakes and coastal regions.

Nevertheless, use of AVHRR imagery for studying the thermal characteristics of lakes has proved useful. The surface temperature dynamics and seasonal variability of the Great Lakes, Lake Baikal, Great Slave Lake, Great Bear Lake, and Lake Lagoda and other European Lakes have been studied (Oesch et all 2003, Hook et. all, 2003, Bussieres and Schertzer, 2003, Rouse et. all, 2003, Schwab et. all, 1999, Bolgrien et. all, 1995). Over the Laurentian Great Lakes, AVHRR imagery is processed daily to derive highly accurate lake surface temperature data. This data has been used for several application, including daily estimates of average surface water temperature and the seasonal dynamics of thermal features in the Great Lakes (Schwab et. all, 1999; Bolgrien and Brooks, 1992; Li et. all 2001). In addition, the surface temperature dynamics and seasonal temperature variability of several large and medium-sized European lakes and Canadian MacKenzie River basin lakes have been examined with AVHRR imagery, and more recently operational processing of AVHRR data has been implemented at the University of Bern (Bolgrien et. all, 1995; Oesch et. all 2003; Bussieres and Schertzer, 2003). Many features of lake temperature structure, including upwelling, thermal bars, and the influence of river runoff into the lake on lake temperature patterns have been documented. As satellite data has become more widely used and available, it is likely that interest in describing the interannual variability of lake temperature using remote sensing data will increase. However, the data must be calibrated to in situ observations and error sources must be well understood and minimized to derive meaning from such a data set.

The only known study which looked a the surface temperature of a saline lake was over Lake Eyre, Australia by Barton and Takashima (1986), and a discussion of some of those results will be presented in later sections. One advantage to remote sensing over the Great Salt Lake is that the climate of the region is semi-arid, meaning that annual sunshine and hence uncontaminated lake temperature pixels are more common over the Great Salt Lake than over most other lakes, particularly during the summer and fall months. One disadvantage is that the Great Salt Lake has larger variations in turbidity and surface characteristics (foam, surface waves, etc) than most freshwater lakes, and the collective effects of these on emissivity and hence temperature retrievals are not certain. In addition, the presence of a warm, low-moisture airmass over the region during the warmer months of the year may make the application of the global SST algorithms unsuitable over the Great Salt Lake (Barton and Takashima, 1986).

Using AVHRR imagery to investigate the Great Salt Lake has many applications beyond studying climate variability. AVHRR visible imagery has been used in several studies over salt playas to describe the seasonal variability of playas (Bryant, 2002; Roshier and Rumbachs 2004). Playas are extremely climatically sensitive to changes in regional aridity (Bryant, 1999). The AVHRR visible channel imagery over the Great Salt Lake region could be used to describe playa surface area in relation to variations in the infrared thermal channels, as well as provide estimates of the evaporative flux across the Great Salt Lake and surrounding playas. In addition, lake turbidity can be derived from AVHRR visible channel data. In some lakes, this turbidity gives some measure of the phytoplankton in the water. While AVHRR imagery does not provide accurate chlorophyll pigment concentrenatations, it can give estimates of concentrations when the water is transparent and little other particulates exist (Semovsky 2000). Additional research is required to determine whether or not chlorophyll estimates can be made over the Great Salt Lake using AVHRR imagery during calm spring and summer weather conditions.

The use of a long-term record of AVHRR-derived temperatures for monitoring climate change in lakes is essentially nonexistent, although recent work has hinted at the use of this data for climate change studies. However, the sea and lake satellite-derived temperature data sets are approaching the point where accuracy and temporal coverage are sufficient to determine a climate signal from the data (Zhang et. all 2004) As stated in Bussieres and Schertzer (2003),

> "*Observation of year-to-year variations in the seasonal water temperature trends of water bod ies would be of interest to the modeling of climate change." Studying year-to-year variations requires in the seasonal water temperature trends of water bodies requires a continuous infrared data set at sufficient resolution, proper calibration, of the data to account for drift in sensor performance, accurate ground georeferencing, and improvement in accuracy...AVHRR data, and the Moderate Resolution Imaging Spectroradiometer (MODIS) sensor data from sat ellites like terra and aqua can probably meet most of these requirements*."

Casey and Cornillion (1999) found that SST's with high accuracy (to within .2 ^{o}C) could be used to detect global climate change within the two decade period of SST's that is currently observable. Since SST retreivals currently have accuracy to about +/-0.5 ^{o}C, global climate signals may soon be possible (Martin 2004). Over the Great Salt Lake, regional climate changes may be observed with much higher error levels, given that lake is very shallow and responds quickly to temperature anomalies over the land, and to wet and dry spells through the interplay of evaporation and precipitation within its closed basin.

CHAPTER 2: DATA AND METHODS

In this section we describe the processing methodology used to obtain AVHRR satellite-derived temperature data over the GSL. The data obtained from each AVHRR channel is discussed and visual examples of data obtained in each channel are presented. Then the theory for deriving lake surface temperature from raw satellite radiances is presented, beginning with converting observed satellite radiances to brightness temperature and subsequently using split-window techniques and various assumptions to derive the sea surface temperature. This section concludes with a description of the cloud masking scheme used to identify and remove cloud-contaminated pixels.

The data for this initial study looking at the thermal characteristics of the GSL used preprocessed CoastWatch files, in which the surface temperature was already processed for the GSL using the techniques described herein. In addition, land and coastline masks are created in the postprocessing by CoastWatch. The images presented in this work are also generated using the CoastWatch software visualization package.

DESCRIPTION OF AVHRR CHANNEL DATA

NOAA polar-orbiting satellites have been equipped with an advanced very high resolution radiometer(AVHRR/2) since the launch of NOAA-7 in 1981. AVHRR/2 measures reflected radience in two visible spectral bands and emitted radience in two or three thermal infrared channels. Beginning with the launch of NOAA-15, the AVHRR/3 instrument has provided six channels, with an additional channel (channel 3a) in the visible-near infrared region. For AVHRR/2, the five channels have effective wavelengths around .63 μ (channel 1), .86 μ (channel 2), 3.7 μ (channel 3), 10.8 μ (channel 4), and 11.5 μ (channel 5). The AVHRR/3 added channel has an effective wavelength of 1.6 μ (channel 3A). Detailed information on both AVHRR/2 and AVHRR/3 can be found in the NOAA POES and KLM Data User's Guide (http://www2.ncdc.noaa.gov/docs/podug/index.htm and http://www2.ncdc.noaa.gov/docs/klm/).

AVHRR/3 channels 1 and 2 are recieved from NOAA polar-orbiting satellites for all daytime passes. Both channels are centered near the longer wavelengths of the visible spectrum. The reflectance data from the AVHRR channels 1 and 2 can be used to determine the turbidity of the GSL. Strong scattering by turbid waters reflects much light, which will show up as enhanced reflectance on the AVHRR images. The scattering particles that determine the turbidity of water in the GSL include phytoplankton, brine shrimp and brine fly eggs, and a host of organic and inorganic sediments. The reflectance of the lake is not only a function of the quantity, physical properties, and size distribution of suspended particulate matter, but also of the temperature of the water and the angle of the sun and the satellite. Under certain conditions, a distinct gradient in reflectance exists between the north and south arms of the Lake, hypothesized to be caused by the different scattering and absorption properties of the algae in the different parts of the lake. However, during other conditions the reflectance of the entire lake is relatively uniform. Figures 1 and 2 show the albedo of the GSL for days when the lake showed high and low reflectivity variations between the north and south arms. More research is needed to explain these desrepancies in the visible channels.

The primary use of the AVHRR channel 1 and 2 data in describing the thermal characteristics of the GSL is for cloud masking. Because of the relatively high reflectance of most clouds compared to water, most clouds can be detected by a simple threshold test for albedo over the lake. However, during windy conditions the enhanced albedo of lake water increases the spectral reflectance such that clear lake surface may be classified as cloud. This problem will be addressed in the cloud masking section.

AVHRR channel 3a is a near IR channel which measures reflectance and has replaced the standard Channel 3 channel during daytime passes of NOAA-17. NOAA-16 remains with the use of channel 3 during daytime and nighttime operations. Channel 3a is used primarily for snow detection because snow is spectrally distinct from clouds in this channel. Its exact utility in cloud masking over the GSL needs further study. However, the albedo of the lake is typicallcally about half the values noted in channels 1 and 2 (Figure 3); thus, the threshold for cloudy pixels may be able to be set to a lower value to aid in cloud detection in this channel.

Channel 3b is used during daytime and nighttime and receives both reflected solar radiation and emitted infrared radiation during the daytime hours. This channel is very valuable in detecting hot spots for fire detection, and is also used cloud mapping and cloud property retrievals. There has been some debate as to whether channel 3a or channel 3b are more valuable during daytime operations (Rosenfeld et. all, 2004). This channel is also used for cloud detection, and cloud tests using this channel are discussed in later sections. An example of channel 3 is illustrated in Figure 4. This channel is not used for deriving sea surface temperature because of sun glint problems.

Channels 4 and 5 are centered in the thermal infrared and are used for deriving lake or sea surface temperature. The theory of split-window techniques employed uses the difference in retrieved brightness temperatures in these two channels to account for attenuation of emitted infrared energy by water vapor. Water vapor attenuation is slightly greater in channel 5, meaning that the retrieved temperature is cooler. The attenuation is theoretically corrected for using split-window techniques, which will be described in the next section. A derived estimate of lake temperature is then produced using the brightness temperatures sensed in the channels 4 and 5. Figure 5(a-c) gives an example of clear-sky brightness temperature for each channel over the GSL as well as the derived lake temperature using methods described next.

HOW SEA SURFACE TEMPERATURES ARE DERIVED:

For this study, post-processed AVHRR data was obtained directly from CoastWatch through a special program initiated prior to the 2002 Winter Olympics to obtain lake temperature data for forecasting guidance during the the 2002 Winter Games in Salt Lake City, Utah. However, additional AVHRR raw satellite data was also obtained via anonymous ftp through the NOAA/NESDIS Comprehensive Large Array-data Stewardship System (www.saa.noaa.gov). The full resolution data was downloaded in High Resolution Picture Transmission (HRPT) format. The level 1b HRPT data are unpacked binary with each satellite scan line containing raw channel 1-5 radiances. The entire processing procedure for converting raw satellite radiances into temperature at the lake surface is now described, although in practice this was mostly done in the post-processing stage at CoastWatch.

The raw satellite radiances are recieved at local HRPT recieving stations. For the Great Salt Lake, the recieving station beginning in 1999 is Monterrey, California. Prior to that date, other stations located in Alaska and Virginia had to be used to get AVHRR imagery over the GSL. Because the GSL is near the end of their coverage area, the retrieved swaths occur at relatively large satellite zenith angles (> 60^{o}). This may cause large errors in the derived lake temperature as there is angular dependence on the emissivity emitted from the lake surface. This issue will need to be explored further before a long-term record of GSL temperatures prior to 1999 are created. For this study, however, only HRPT data from the Monterrey HRPT receiving stations is considered, which has typical satelite zenith angles less than 45^{o}.

Pre-processing is conducted at each NOAA recieving station which creates earth location files and appends calibration and satellite navigation files to the raw HRPT data stream. Coastwatch Format Sofware Utilitities and Library were obtained by the University of Utah in 2002 for visualizing Coastwatch HDF files. The CoastWatch utilities also have the capability to import and process level 1b HRPT data.

Each AVHRR instrument is calibrated before launch using a evacuated test chamber at a cold space temperature of 288 K. Even so, errors are introduced by the instument, which depend on factors such as internal operating temperature and radiometer age. However, nonlinear corrections (Brown et all. 1993) can be applied to the HRPT data stream that results in an improvement in the accuracy of the retrieved brightness temperature. The nonlinear correction method was derived from a detailed reanalysis of linear calibration procedures previously done at NOAA to calibrate the AVHRR instrument (Walton 1998) and will not be discussed further here.

Next, the brightness temperature of for each thermal channel is then computed using the Planck equation:

$$T(E) = (C_2 V)/(\ln(1 + (C_1 V^3)/E))$$

where T is the temperature (K) for the radiance value E, V is the central wave number of the channel (cm-1), and C_1 and C_2 are constants. The central wave number is a function of temperature for each channel and can be found in the NOAA Polar Orbiters Data User’s Guide. In the CoastWatch software, a single central wavenumber is used for the 270 to 310K wavenumber range.

Finally, sea surface temperature is approximated from the raw channel radiances using linear and nonlinar multichannel algorithms. These algorithms and the theory behind them are discussed next.

SPLIT-WINDOW TECHNIQUES

The theory behind split-window techniques for estimating SST’s is relatively simple. Absorption of surface radiation with corresponding emittance of atmospheric radiation results in the radiation emitted from the surface being absorbed and reemitted by gases and particles in the atmosphere (McMillan 1984). As shown in McClain et al. (1984) using the mean value theorem of calculus yields the following radiative transfer equation,

$$I_b = B_i(T_s)\tau_i + B_i(T_i)(1 - \tau_i) \tag{1}$$

where I_b is the observed satellite radiance, T_i is top of atmosphere radiance, T_s is the sea surface temperature, B(T) is the Planck function at a given temperature T. The transmissivity of the atmosphere is represented by τ_i.

The atmospheric absorption in the thermal infrared channels occurs at low levels in the atmosphere, where water vapor is concentrated. The influence of aerosols on atmospheric transmissivity is too variable to be modeled explicity and how aerosol effects are included in deriving SST’s will be described later. Because atmospheric absorption occurs primarily at low levels, a constant mean atmospheric temperature T_i is assumed for all channels. Because water vapor is the primary absorber in these channels, transmittance can be approximated by the following equation, where k_i is the absorption coefficient and X depends on the amount of water vapor present. (2)

$$\tau_i \cong 1 - k_i X \tag{2}$$

assumes that transmitance is a linear function of water vapor amount, a simplification that is not always true. (1) is then expanded in a Taylor series in terms of temperatuere

$$T_i - T_s = k_i X(T - T_s) \tag{3}$$

Where T_s is the sea surface brightness temperature and T_i the brightness temperature at the given spectral interval. Combining (3) with another identical equation for another spectral interval j, gives

$$T_s - T_i = [k_{i} / (k_j - k_i)](T_i - T_j) \tag{4}$$

The outcome of (4) is that is the difference between the observed temperature and what is observed in channel i is linearly related to the difference between the brightness temperature between two different window channels i and j. This is the basis of the operational nonlinear SST equations in use today. McClain (1985) used regression of (T_s-T_i) against (T_i-T_j) to obtain a surface temperature given the retrieved radiances. A series of diverse temperature/humitity profiles were used to simulate brightness temperatures that correspond to a given atmospheric profile (with the lowest level of the vertical profile assumed to be the surface temperature). However, this technique was found to have substantial bias as the derived radiances were not calibrated to surface measurements, in part because available in situ measurements are derived from bouys below the ocean surface. This problem was solved by multiplying the result of (4) by constants derived empirically by regressing the results of (4) with in situ observations. The multi-channel algorithms (4) now took the form of (5), where the C_1, C_2, and C_3 are constants that were deter-

$$SST = C_1 T_4 + C_2(T_4 - T_5) + C_3 \tag{5}$$

mined by least-squares regression of satellite SST retreivals with in situ matchup observations (Walton 1998, Martin 2004). The form of the split window algorithm in (4) and (5) assumes that T_s is independent of satellite zenith angle and total columnar water vapor. Zenith angle correction terms have since been added to the linear SST's to better approximate atmospheric attenuation (Martin 2004).

For very moist or dry conditions, the linearization of the water vapor effect in (4) and (5) can be undesirable. Nonlinear multichannel algorithms for satellite-derived SST's have been developed in the past decade with the goal of accounting for cases where the atmospheric transmittance cannot be approximated accurately with (2). The method developed for this purpose was called the cross product sea surface temperature (CPSST) (Walton 1988, Walton 1998). The current operational nonlinear (NLSST) equations (6) includes a term to correct for a varying satellite zenith angle, as well as implicitly accounting for the nonlinear influence of water vapor amount in the atmosphere on SST retrievals (Martin 2004).

$$SST = C_1 T_4 + C_2 T_{sfc}(T_4 - T_5) + C_3(T_4 - T_5)(\sec(\theta - 1)) + C_4 \quad (6)$$

The constants C_1-C_4 are again determined empirically using statistical regression methodology. The constant T_{sf} is the "first guess value" of SST and is taken from climatology or the linear MCSST approximation (Barton 1992, Oesch 2003). These constants represent our best method of correcting the derived product to match with our observational data set, and as such they correct for a number of error sources that are not well understood.

The basic methodology of deriving SST's using MCSST and NLSST techniques has changed very little in the last 15 years. Most deviation from the above equations has been in the form of redefining the constants C_n using more detailed in situ validation. Another important point to note about derivating the operational SST alogorithm coefficients is that the retreived radiences sense the top 30 μ of the water column but are validated against "bulk" bouy measurements from instruments located 1 m below the water's surface. Complications arising because of this assumption will be discussed later.

PROCESSING : CLOUD MASKING

A cloud masking file is appended to the CoastWatch files obtained by the University of Utah. The cloud masking algorithm and the theory behind the various tests is presented in this section.

The cloud identification process varies from day to night, and uses a combination of data from AVHRR visible and thermal channels to screen for clouds. The cloud detection tests are based on the facts that clouds are more reflective and are colder than the ocean, and the ocean is nearly uniform in temperature and reflectivity over large regions (Stowe et all. 1999). Cloud detection tests include threshold tests and uniformity tests. Threshold tests identify individual cloudy pixels that are more reflective or colder than the ocean surface, while uniformity tests examine the thermal or reflective variability of a rectangular array of pixels to determine cloudy regions. There are also tests that identify specific cloud types using retrieval data from several channels, and tests that compare pixel values to climatology (Martin 2004).

Cloud masking procedures have all but eliminated SST errors introduced by cloud contamination over the oceans (Robinson et. al. 1984). However, the full set of cloud tests used over the ocean is too restrictive over the GSL. Most notably, uniformity tests are not applied over the

Lake because of the highly nonuniform nature of the surface temperatures. Because of this, more cloud contamination can be expected in satellite-derived temperature over the GSL than over the ocean. However, because the GSL region is so small, the test thresholds on various cloud tests may be able to be set to more stringent values than over large oceanic regions, making possible improvements in the cloud tests that are used.

Cloud masking of AVHRR thermal imagery is conducted primarily through the Clouds from AVHRR (CLAVR) algorithm, developed to classify clouds for 4-km GAC AVHRR data. Different algorithms are conducted for daytime and nighttime imagery, with daytime cloud masking of the highest quality because of the use of the differing reflectance properties of lake and cloud surfaces (Stowe et. all, 1999). The set of CLAVR tests used over the GSL are a subset of the CLAVR algorithms developed for cloud masking over the Great Lakes (Maturi and Pitchel, 1993). Similar cloud masking procedures have been used over European lakes (Oesch, 2003). Recent developments in the CLAVR algorithm have led to the development of an improved cloud masking algorithm, the CLAVR-x. The CLAVR-x contains improved thresholds on existing tests as well as some additional tests. The CLAVR-x will not be described in this section, but future improvements in the cloud-masking capabilities are likely once the CLAVR-x is implemented over the GSL (CLAVR-x user's manual, 2004).

The scientific basis and evaluation of CLAVR-1 cloud masking techniques are given in Stowe et. all (1999). In the visible and near IR spectrum (0.3-3 microns--AVHRR channels 1,2, and 3a), the incident solar radiation dominates the thermally emitted radiation. The amount of reflected solar radiation depends on the physical and chemical composition of the Earth's surface and atmosphere. In the thermal infrared, from about 3-20 microns (AVHRR channels 3b, 4 and 5), earth-atmosphere components can be discriminated through their absorption and reemission properties.

During the daytime the differing reflectance properties of the land and water in the visible spectrum (channels 1 and 2) are used to detect clouds over these areas. Tests that use the various reflectance properties of clouds and land or water surfaces include "contrast signatures" (differing composition of system components) and "spectral signatures"(component that reflects beeter in one part of the visible spectrum than another). Additionally, spatial signatures (how uniform is the reflectivity field) can be used to discriminate between clouds and land/ocean features. In the thermal IR, attenuation of the outgoing radiation is primarily due to water vapor and aerosols.

A number of tests have also been developed in the thermal infrared spectrum, and these tests are crucial for cloud masking, expecially for thin clouds during the day and for all clouds at night. In this spectral region the contrast, spectal, and spatial signatures still apply, but the differing absorption and reemission properties of clouds, land and water surfaces are used to distinguish between clouds and water surfaces. The nighttime cloud detection tests depend on channels 3-5 differences. Emissivity of clouds and ocean are such that the difference T_5-T_3 will be positive in cloudy regions and negative in clear regions. In addition, the difference T_4-T_5 is used to help detect the presence of thin cirrus clouds during both day and night.

Each cloud test has a threshold value, determined through theoretical and empirical methods, that is used to determine whether a pixel is water or cloud. For instance, if the reflectivity of any pixel is over 20% then it is classified as cloud, otherwise that particular test classifies the pixel as clear. Many of the various test thresholds currently used by CoastWatch cloud masking procedures could possibly be refined to more accuratly mask clouds over the GSL once detailed knowledge about the reflectance and emission properties of the region are known. However, this requires careful examination of cloud and surface channel retrievals for an entire range of temperature, season, and solar/satellite zenith angles--a significant task that has not yet been applied. Perhaps a more feasible improvement would be to apply some of the improved thresholds derived for the CLAVR-x algorithm. The following text lists the tests currently used over the GSL and the threshold values.

Cloud Tests used on the Great Salt Lake (Adapted from Maturi and Pichel 1993)

CLAVR Test	DAY/NIGHT	Channel Used	Threshold
Reflective Gross Cloud Test (RGCT)	DAY	1,2	> 10%*
Reflectance Ratio Cloud Test (RRCT)	DAY	2	0.9 < R > 1.1*
Channel 3 Albedo Test (C3AT)	DAY	3,4,5	> 3 %
Four Minus Five Test (FMFT)	DAY/NIGHT	4,5	function(T4)
Thermal Gross Cloud Test (TGCT)	DAY/NIGHT	4	< 271*
Uniform Low Stratus Test (ULST)	NIGHT	3,5	function(T4)
Cirrus Test (CIRT)	NIGHT	3,5	function(T4)

* Values used can and will likely be modified in the future to enhance cloud detection over the Great Salt Lake

Reflective Gross Cloud Test:
Channels used: channel 2 (over ocean) and channel 1 over land. Both channels are useful over th Great Salt Lake for detecting clouds. The reflectance of the GSL is generally less than 10%, except in regions of sun glint. The reflectance of most thick clouds is much higher, such that any pixels with a reflectance over 10% are classified as cloudy.

Reflectance Ratio Cloud Test:
This test computes the ratio of channel 1 to channel 2. If this ratio is between 0.9 and 1.1, the pres ence of clouds is indicated. For cloud-free ocean this ratio is typically less than .75. Over vege tated land surfaces, the ratio is typically greater than 1.2. However, for desert regions, RRCT is typically in the cloud range.

Channel 3 Albedo Test:
This test is useful for detecting thin clouds or partial cloud contamination from sub-pixel sized clouds. Channel 3b detects reflected solar radiation as well as emitted thermal radiation. This cloud test estimates the emitted infrared radiation by using channel 4 and 5. Reflected radiation in channel 3 is subtracted from the total emitted radiance. The reflected component is then used to estimate a equivalent isotropic albedo. Cloud-free land surfaces are appreciably brighter than the ocean. Thus, the threshold albedo value for cloud detection is much lower over water than over land.

Channel 4 Minus Channel 5 Test:
Used to detect thin cirrus which can have larger differences that are caused by water vapor attenu ation. This test is employed primarily to detect thin cirrus which can produce larger FMFT values than are possible than water vapor attenuation alone. This results from ice particles having a lower emissivity (higher transmissivity) at 10.8 than at 11.9 microns, creating a FMFT signature similar to that of water vapor. However, if cirrus are thick enough, then the difference can be greater than

that caused by water vapor alone. The threshold for this test is is a function of channel four bright ness temperature. There may be issues with this test when a temperature inversion exists and over land (threshold needs to be higher).

Thermal Gross Cloud Test:
This test examines the channel 4 temperature versus a threshold of 271 K over the ocean--which corresponds to 0 degrees Celcius surface temperature. Over the GSL, the test needs to be changed to about -4 degrees C during midwinter, as subfreezing lake temperatures are common since the high salinity lowers the lakes freezing point about 5 degrees.

Uniform Low Stratus Test:
This is an important test at night. Used to detect low stratus clouds that have a similar brightness temperature as the surface. Works because clouds are more reflective and atmosphere more trans parent at 3.7 microns than at 10.8 and or 11.9 microns.

ULST threshold = exp (a + bT4) -1.0

a = -9.375, b = .0342, and threshold varies from 273 to 300 K.

Cirrus Test:
The cirrus test is defined as the difference between channel 3 and 5 temperatures divided by the channel 5 temperature

While not integrated into the operational LST algorithm, pixels viewed with an satellite zenith angle greater than 53 degrees leads to less accurate SST retrievals, since there is more atmospheric attenuation of surface emittance for longer path lengths. This could be a problem with IR satellite data prior to 1999, when the Great Salt Lake region tended to be covered near the edges of the archived LAC swaths.

In addition to cloud-masking, georeferencing of the satellite images is necessary as there is some inherent error in assigning pixel values to a location on a grid. The images of the GSL can be accurately georeferenced by using several known physical features: 1) The Southern Pacific Railroad (SPRR) causeway, which divides the north and south arms, and 2) shoreline locations where there is steep bathymetry, such as just to the east of Antelope Island and Freemont Islands.

PROBLEMS WITH CLOUD MASKING OVER THE GREAT SALT LAKE

Cloud tests which use the infrared channels are very sensitive to variations in the relative differences of the channel brightness temperatures. Typical values of various channel differences over the ocean (e.g., channel 4 minus channel 5 test) may not always be true over the highly saline GSL. In addition, there are channel-dependent changes in emissivity due to the salinity and turbidity of the GSL. In multiple instances, areas of the lake have been falsely detected as cloud using these difference tests. Figure 6 shows how much of the GSL was classified as cloudy during a clear day in January 2005. The lake temperature in this region was around zero degrees Celsius in this image, and it is believed that the FMFT algorithm yielded values in the “cloudy” range due to the presence of ice in the lake water.

In addition, turbidity in the lake shows large variations, which directly impacts the visible channel albedo and the brightness temperature in different channels. Thus, reflectance thresholds for clouds over the GSL will vary with the limnology and meteorology. Figure 7 illustrates how a windstorm over the lake on April 13th, 2005 increased the channel 1 reflectivity of the GSL such that all tests which use reflectivity measurements classified the lake as cloudy. In addition, the relative values of the channel four and five brightness temperatures were decreased to a value outside of those values typically expected over the oceans. Figure 8 shows the results of the FMFT in which most of the north arm was falsely classified as cloud. The values of the FMFT are also given, and show that the difference in the channel four and five temperatures was reduced in that region.

Cloud contamination from thin cirrus clouds that are not detected by any cloud tests can have a noticable impact on derived GSL temperatures. Figure 9 shows a GOES visible and infrared image and the derived SST product for a day with an approaching thin cirrus shield. Notice how the thinnest cirrus near the leading edge of the cloud shield is not masked as cloud, but has the effect of lowering the lake temperature several degrees.

An improved set of threshold values derived for the unique Great Salt Lake region cloud tests and/or an improved understanding of the conditions in which certain tests falsely detect the lake surface are required to enhance the accuracy of cloud detection.

CHAPTER 3: FACTORS INFLUENCING GREAT SALT LAKE TEMPERATURE

The surface temperature of the GSL depends on complex interactions between wind, latent/sensible energy exchanges, lake stratification, bathymetry, transparency, and vertical and horizontal water motions. Determining the primary mechanisms controlling lake surface temperature variability is difficult as the relative magnitude of the parameters described above are not well-understood in the GSL. In particular, the lake stratification regime and the circulation patterns in the lake are not spatially or temperally documented. Analysis of AVHRR imagery in combination with the knowledge of key physical mechanisms influencing lake surface temperature will hopefully yield conclusive information on temperature variability in the lake.

While the relative importance of the various physical mechanisms is not certain, a limnological review shows that GSL surface temperature is controlled primarily by the following factors:

1). Solar heating and infrared cooling of surface water and playa bottom in very shallow waters
2). Wind forcing resulting in 1) mixing with cooler or warmer subsur face water and 2) upwelling/downwelling
3). Lake currents and associated advection of cooler/warmer waters
4). River inflow (seasonal)
5). Evaporative cooling
6). Seiching

A description of the limnology surrounding each of these controls on the surface temperature and some possible examples of the effect of these controls on lake surface temperature as seen in AVHRR imagery of the GSL follows.

SOLAR HEATING AND INFRARED COOLING

The most obvious factor influencing the surface lake temperature of the GSL is solar heating and infrared cooling. Under calm winds during the summer months the lake surface can warm very rapidly. This is due in part to the fact that the specific heat capacity of GSL is about 20% less than for fresh water. In addition, the many shallow bays and shoals along the lake heat up very rapidly as the sediments and ground under the shallow water absorb energy and transfer it to the water above. The very warm bay water becomes less dense, and is able to float out over the lake surface as a warm, lighter density current. In addition, at night the bays are able to cool more rapidly, but if the enough fresh water has entered the bay, the water will enter the lake as a less dense but possibly cooler surface layer. Figure 10 gives examples of some cases where strong heating of the lake resulted in very warm temperatures in the shallow areas of the lake, with areas of warmer water possibly entering the lake as a less dense, warm surface layer.

Because of the high salinity of the GSL, it has a specific heat that is significantly less than that of pure water. Millero et all (1973) gave a method to determine the specific heat at surface pressure for water at temperature T and salinity S can be calculated as:

$$C_p(S,T,0) = C_p(0,T,0) + S\left(-\left(7.6444 + 0.107276T - 1.3839 \times 10^{-3}T^3\right)\right) + S^{\frac{3}{2}} 0.17709 - 4.0772 \times 10^{-3}T + 5.3539 \times 10^{-5}T^2$$

$$C_p(0,T,0) = 4217.4 - 3.720283T + 0.1412855T^2 - 2.654387 \times 10^{-3}T^3 + 2.093236 \times 10^{-5}T^4$$

Using a typical temperature of the GSL, 295 K, we will compare the specific heats of freshwater and high salinity water. The value for parctical salinity S is essentially the same as PSU, or practical salininty units. Thus, the PSU of the ocean is around 35 psu. The GSL, which is typically 3-5 times saltier than the ocean should have a psu of at least 160. Comparing the specific heats for the GSL $C_p(160,295,0)$ versus for freshwater $C_p(0,295,0)$ calculated at this temperature, we get

$C_p(0,295,0) = 4182$ J/kg deg K

$C_p(160,295,0) = 3451$ J/kg deg K

which illustrates that the GSL heat capacity is around 20% less than for fresh water at a temperature of 20^{o} C. This partially explains why the GSL is able to heat up and cool off so rapidly from daytime to night.

BACKGROUND ON OPTICAL PROPERTIES AND SOLAR HEATING:

The optical properties of lakes influence the surface heating of the water since the clarity of the water determines the absorption of solar energy as a function of depth. In addition, turbidity fluctuation influnce the emissivity of saline water, an important consideration for satellite retrievals over the GSL discussed later.

The optical properties of lakes are a function of the water quality and the incoming direct and indirect solar radiation. The annully averaged incoming radiation is a function of latitude and average cloud cover across the region of interest. Reflection at the surface of the water decreases how much radiation actually enters the water and also has a dependence on the angle of incoming radiation as well as the turbidity of the water itself. Reflectance from lakes according to Hutchinson (1957) varies from 10% in winter to 7% in summer. The average annual reflectance and Mono Lake, California was found to be about 7% (Mason 1967). Reflectance values in visible channels 2 and 3 over the GSL have been onserved to range from 4 to 14% in observed AVHRR imagery.

Most information regarding the transparency of saline lakes is derived from Secci discs, which measure the water's optical transparency. The Secci disc transparency measurements are highly variable, ranging from a a few centimeters to over 10 meters. The lowest transparency occurs when there are large amounts of suspended organic or inorganic sediments in the water. These conditions occur when there are phytoplankton blooms or sediments are suspended through wind forcing. Shallow saline lakes, like the GSL, tend to have low transparency due to wind forcing that continully mixes bottom sediments to the surface. Transparency measurements in shallow, polymictic Sasketchewan salt lakes were generally below 1 meter due to the constant wind stress. In highly productive saline lakes, phytoplankton have been found to reduce transparency to a few centimeters (Hammer, 1984).

Transparency measurements are typically correlated with turbidity measurements. Decreased transparency usually corresponds to an increase in the amount of suspended organic or nonorganic particles. However, Kulby (1982) found that salinity had a negative correlation with turbity measurements in California playa lakes, with the lowest turbidity in those lakes with the highest dissolved solid concentrations (Hammer, 1984).

The apparent colour of playa lakes is highly variable, as illustrated by the differing colour of the north and south arms of the GSL. The colors are typically caused by variable concentrations of different algae species. The red color of the north arm of Great Salt Lake is caused by the abundance of the Artemia phytoplankton.

For the GSL, significant daily to seasonal lake transparency variations are noted due to irregular strong wind events, seasonal algae blooms, and river particlate discharge. EPA water qualtity tests found that the turbidity of the GSL between 1.0 and 95.0 NTU's , or nepheliometric turbidity units. The variations in suspended solids was found to be between 0 and 500 mg/liter (EPA Utah Water Quality Data). These variations in lake turbidity could lead to variations in lake surface emissivity (discussed later) and how rapidly the lake surface temperature heats up, particularly under stratified, calm weather conditions. Because the variability of the light penetration will influence how the absorption of incoming solar radiation is distributed in the water column, under stratified lake conditions, lower transpariencies would tend to cause intense heating of the top of the water column, since little radiation would be transmitted and absorbed in the subsurface water.

LAKE MOTION: INTRODUCTION

Limited research has been conducted looking at water movement in saline lakes (Hammer, 1986). However, the circulation and mixing patterns in any lake are coupled to a number of well-understood physical forcing mechanisms that influence surface lake motion and hence temperature.

Turbulent motions in lakes can be classified as waves or currents. The relative importance of these two mechanisms in mixing lake waters varies from lake to lake. Wind and heat loss or gain at the lake surface are the two primary mechanisms which drive lake motion. Wind drives both surface and internal (called seiches) waves as well as surface currents. Density differences in horizontal and vertical directions create density currents.

WIND FORCING: MIXING, UPWELLING/DOWNWELLING, THERMAL TEMPERATURE STRUCTURE, DENSITY STRATIFICATION

In large, deep freshwater lakes, it is typically not difficult to determine the causes of areas of cold or warm water and circulation in the lake. However, the temperature variability patterns found in the GSL need to be carefully analyzed in order to determine the lake dynamics responsible for any observed anomalies. It is difficult to distinguish between wind-driven dynamics (i.e., upwelling/downwelling, areas of enhanced/supressed mixing), and variations in surface heating of the water as described earlier. Better understanding of lake stratification would help somewhat to alleviate this problem.

The GSL is a shallow lake with a large wind fetch, so the entire lake column is believed to mix whenever there is a moderate wind event across the lake. This mixing event is known as polymictic (Kalff 2002). Thus, the vertical temperature profile becomes isothermal after each large wind event. However, because of the large number of shallow bays and shoals surrounding the lake and the generally shallow nature of the lake, lake surface temperature responds quite rapidly to surface warming and cooling. During the Spring and Summer, intense surface heating of the surface water layer above cooler, deeper water creates thermal stratification of the water column. At the same time, weaker winds leads to less mixing and hence the ability of the temperature stratification to become semi-permanent during the summer months. Thus, during quescient periods a seasonal thermocline may develop that is not regularly mixed out (Kalff 2002).

LAKE STRATIFICATION

Shallow saline lakes tend to be polymictic, or have frequent mixing of the entire water column. Many shallow saline lakes mix “continually or daily, such that water temperature lag the air temperatures by only a few degrees. The exact mixing regime of the GSL is not well known. After the construction of the SPRR railroad causeway in 1959, the differences in brine sainities between the north and south arms of the Lake created a permanent dense brine layer in the bottom of the south arm. Stephens and Gillespie (1976) describe this condition, with a dense brine layer below 7.5 meter depth that was not disturbed even during storms. This condition of permanent density stratification remained intact in the south arm until mid-1991, at which point the permanent stratification dissappeared due to declining water levels and the creation of a breach in the causeway that allowed water from the two arms of the lake to mix. Between 1959 and 1991, the GSL was classified as meromictic, meaning that is was characterized by partial or incomplete mixing.

Since 1991, the GSL has been polymictic, although periods of lake stratification of days to weeks or longer probably occur. Stratification in very shallow (less than 4 meters) Saskatechewan lakes was investigated by Hammer and Haynes (1978), and it was found that periods of stratification were confined to brief periods of light winds and hot temperatures. However, because of the higher density of saline lakes, the mean depths required for stratification are less for saline lakes than for fresh water lakes, since a given amount of wind energy will overturn less depth of a denser substance. In Sasketchewan freshwater lakes, prolonged stratificaton of the lake only occurs when the mean depth is over 10 meters. Somewhat less depth is required for saline lakes to become stratified, and it is well established that saline lakes “may exhibit very sharp vertical stratification in temperature" (Northcote and Hall 1983). Even though the mean depth of the GSL is under 5 meters, it is likely that the deeper parts of the lake are thermally stratified whenever there are periods of relatively light winds. Interestingly, Stephens and Gillespie (1976) found that the temperature of the surface mixed layer (above 7.5 meters) had temperatures that ranged from 10 ^{o}C in April to 27 ^{o}C in July, while the temperature of the dense, stabe brine layer ranged from 10 ^{o}C to as much as 24 ^{o}C in August. Hammer hypothesized that the warm deep water temperatures were caused by heliothermic heating--the fact that the deeper water with higher salinity has a lower specific heat and thus warms more rapidly for a given amount of solar insolaton. Interestingly, this would require relatively clear lake conditions for light to penetrate appreciably to 7.5 meter depths. This so-called "greenhouse effect" in the lake is called heliothermal heating (Egorov, 1993). Another consideration for the GSL is that years with a much larger that average runoff would likely see stronger stratification due the increased density stratification caused by the lighter freshwater runoff over the more saline lake water, and would also see an enhanced "greenhouse" effect which would impact lake temperature.

During periods of weak winds in the fall cooling period, convection is the primary mixing energy contributor. A study of the saline Lake Mahoney in British Columbia found that convective mixing accounted for 72% of the mixing energy on the lake during the fall (Ward et all, 1990).

EFFECT OF WIND FORCING ON GREAT SALT LAKE TEMPERATURE

It is likely that wind is by far the most important factor is determining GSL surface temperature after solar heating (and perhaps convective mixing during the fall). Thus, differential mixing caused by variations of wind speed and direction across various regions of the lake, interacting with lake topography and thermal stratification, would produce the largest variations in lake surface temperature. Studies on the hypersaline Dead Sea and the less saline lake Kinneret have found that variations in the local thermally-driven flows result in variations in mixing depth and consequently surface water temperature across the lake. Pan et. all (2002) found there were large variations in the temperature field in lake Kinerret due to differential winds and land-lake temperature contrasts. This effect appears to be quite notable in the GSL. Figure 11 shows the differences in temperature believed to be caused by differences in the mixing depth between various regions of the lake due to variations in the wind speed observed over the lake surface. Note the cooler regions in the south and north arms hypothetically caused by more intense mixing due to locally stronger winds in those areas. Figure 12 shows the effect of strong winds on mixing the water column and inhibiting the warming effect of the sun on the surface temperature of the Lake during a windy day.

The depth of mechanical mixing by wind-driven waves in a lake depends on the 1) density stratification , and 2) the wind speed. During periods where there is a varying vertical temperature profile this would lead to changes in surface temperature that would be a function of the depth of mixing that was occuring. Surface gravity waves, or simply waves, are definitely the most important water movement in the GSL. Because of its large surface area, the Great Salt Lake is capable of having large waves when the wind is strong. Wetzel (1938) derived an empirical relationship for estimating maximum wave height

$$H = 0.332\ F^{0.5}$$

where F is the maximum fetch in Kilometers. For the Great Salt Lake F is roughly 30 kilometers east-west across the lake, and over 60 kilometers along a northwest to southeast axis in the South arm of the lake south of the causeway. The corresponding maximum wave heights are:

$$H\ (30\text{ km}) = 1.8\text{ meters}$$

$$H\ (100\text{ km}) = 2.6\text{ meters}$$

While these numbers may be an overestimate in the GSL due to its shallow nature and the fact that saline water is more dense than fresh water, it becomes clear that waves with a height nearly 1/4 as high as the maximum depth of the GSL are possible in extreme northwest wind events. Because the density of the GSL water is approximately 30% higher than fresh water, it takes more wind forcing to move the denser water (Hecht).

In addition to direct turbulent mixing of the water column by the wind, the wind also drives larger-scale Ekman transport of lake water and associated upwelling/downwelling within the water column. At this point, determining the upwelling of water due to direct wind stress and that due to Ekman pumping effects in the GSL has been unsuccessful because of the high variability in wind speeds and direction around the lake. Consequently, determining a true upwelling signal versus an "enhanced mixing" signal in the thermal fields of the GSL has been impossible. Figure 13 shows an example of what appears to be upwelling to the north of the SPRR causeway on April 16th, 2005. However, there are some uncertainties as to whether the apparent cold signal is really upwelling of colder water, or some construction of the remote sensing limitations, such as an unknown change in emissivity that influenced the satellite retrieval over that region.

Despite its shallow nature, seiches are still a common occurence in the GSL. Exactly how these may impact lake temperaure is unknown, but could be a possible explanation for the upwelling signatures noted along the SPRR causeway. Gauge stations on the lake have found the presence of seiches happening every few days due to wind forcing over the GSL throughout the year. Over the south arrm, the period of the oscillation was found to be around 6 hours (Lin, 1976).

LAKE CURRENTS

Infrared thermal images have been used extensively to study the circulation regimes in lakes because the temperature variability found in the images gives information to the location of gyres, thermal fronts, and areas of upwelling in lakes.

Wind forcing, Ekman pumping, and density currents in combination with the lake bottom topography and river inflow to the lake all contribute to the net GSL circulation. However, the primary forcing mechanisms of lake circulation in the GSL remain the subject of speculation and confliciting theories. The circulation pattern of the GSL may vary from year to year depending on such factors as density stratification, runoff volume, and the area and depth of the Lake. This is thought to be partly because of the complex topography surrounding the lake and the uneven distribution of wind intensity and direction across the lake. However, studies investigation lake movement in lake Kinneret, Israel, show that the lake-breeze has little impact on the larger lake circulation (Pan et all, 2002). Rich (1991) proposed that the prevailing circulation in the south arm should include a counter-clouckwise rotation driven by the Coriolis force, assuming a prevailing southwesterly wind for the GSL. Others have proposed that lake circulation is driven primarily by river inflow (Katzenburger and Whelan, 1975) while others studies have theorized that wind was not a significant factor in driving lake circulation (Gwynn, 1978). The possible impact of river inflow and lake circulation on the surface temperature of the GSL needs to be monitored further using satellite imagery before any hypotheses can be proven. One difficulty using the remote sensing approach is that distinguishing lake currents in the thermal imagery may prove difficult at best. Figure 14 shows possible cold river inflow from the Weber river; however, the signal has appeared so sporatically in thermal images of the GSL that the exact cause of the cold signal is still uncertain.

Accurate lake bathymetry information, crucial to determining lake circulation will become available over the GSL for the first time this summer (Baskin, personnal communication). This will aid in future determinations of lake circulation and areas where we might expect upwelling in the lake.

CHAPTER 4: IS THE IMAGE REALLY WHAT WE THINK IT IS? ERROR SOURCES AND OTHER REMOTE SENSING CONSIDERATIONS:

Accurate satellite-derived sea surface temperature (SST) or lake surface temperature (LST) data requires careful consideration of instrument errors and atmospheric and surface processes that influence the upwelling radiation field (Figure 15). Sources of error include instrument noise, poor sensor calibration, cloud contamination, atmospheric absorption and emission, air-sea interaction, and surface characteristics including water surface roughness and turbidity (Brown 1993, Martin 2004). Even within the atmospheric window region, the radiation reaching a satellite radiometer has been "contaminated" by atmospheric absorption and subsequent re-emission of radiation (McMillin, 1984), and compensating for the atmosphere's effects when deriving SST's from satellite radiance retrievals was the focus of much research during the first two decades of remote thermal sensing (Anding and Kauth 1970, McClain 1984, McMillin 1984, Walton 1988).

Water vapor, clouds, and atmospheric aerosols all influence the upwelling radiation field. Water vapor attenuation is corrected for in the SST operational algorithms using split-window techniques described earlier (McClain 1984, Walton, 1988). Because attenuation of water vapor changes between different infrared bands, several bands can be used in combination to determine and correct the contributing atmospheric radiance. In this section we will give an overview of the primary error sources in deriving GSL temperature: surface emissivity effects, atmospheric attenuation, and instrument noise/calibration.

SURFACE EMISSIVITY EFFECTS

Emissivity variations of as little as .01 correspond to as much as one degree change in surface temperature retrievals for a completely dry atmosphere (Wen-Yau et. all, 1987). In reality, atmospheric moisture acts to decrease the sensitivity of the satellite to the emissivity variations such that for a moist water vapor sounding, an emissivity variation of .01 yield only around 0.2 ^{o}C change in retrieved temperature. In the following discussion, we will show how a satellite underestimating or overestimating emissivity when calculating SST will lead to surface temperatures that are warmer or colder than the true surface temperature, respectively.

Because emissivity is relatively constant across the world's lakes and oceans, the influence of emissivity variations on the satellite-derived surface temperatures has been largely neglected (Gnatt, 1988). A constant emissivity value between .98 and 1.00 is generally used for deriving SST's with high accuracy. Brown et. all (1996) investigated the surface emissivity of the ocean skin layer using a ship-bourne interferometer in the Gulf of Mexico. Studies on the infrared characteristics of ocean waters have supported that assumption, finding that the observed range of ocean reflectivity/emissivity are so small that they influence sea surface temperature retrievals very little.

However, for large zenith angles, strong winds, or highly turbid saline water, variations in emissivity may contribute to significant errors in sea surface temperature fields (Smith et all, 1997, Gnatt, 1988). Research on surface emissivity in turbid coastal regions of the ocean have noted that emissivity variations may be significant when mapping surface temperatures in coastal areas (Gantt, 1988, Wen-Yao et. all., 1987). Because of the unique environment of the hypersaline GSL, an overview of the influence of salinity, turbidity, and wind on the emissivity characteristics is crucial for determining possible error sources for satellite-derived infrared temperature retrievals over the lake. Under calm conditions, emissivity variations are not expected to contribute more than 1 degree C to the total error of satellite-derived temperatures over the GSL. Under turbid and very windy conditions, as well as at larger satellite zenith angles, the errors may be larger and more research is needed to quantify the possible impacts.

The emissivity from a plane water surface is given in Masuda et all (1988). The authors note three factors which influence surface emissivity:

1). Surface roughness and surface foam or slicks
2). Changes in refractive index due to salinity, chlorinity, and temperature changes
3). Zenith angle dependence

In addition, several studies have concluded that emissivity is also a function of the turbidity of water, i.e., the amount of suspended organic or inorganic matter. We will discuss each of these influnces in more detail below. First, however, we present a first-order approximation of temperature changes in association with changes in emissivity.

Consider greybody irradiance described through Stephan-Boltzmanns law:

$$E = \varepsilon\sigma T^4$$

Where E is the greybody irradience, ε is the emissivity, σ is the Stephan-Boltzmann constant and T is the temperature of the greybody. In most sea surface temperature retrievals, the emissivity is considered a constant value between 1.0 and .98. For our discussion here, let us assume that the emissivity is assumed to remain some constant value ε_o when calculating satellite brightness temperatures. Second, assume that the emissivity of the surface being remotely sensed does not really have an emissivity of ε_o, but instead has a true emissivity ε_o plus some correction factor, $d\varepsilon$, such that the true emissivity of the surface ε_t being sensed is $\varepsilon_t = \varepsilon_o + d\varepsilon$. Because of this fact, the satellite-derived temperature using the value of e0 will not be the actual temperature of the emitting object, which is in reality emitting at a true temperature $T_t = T_o + dT$. To determine the influence of the differences between the satellite observed temperature and the true temperature, consider the equation

$$\varepsilon_t \sigma T_t^4 = \varepsilon_o \sigma T_o^4$$

which yields the relation

$$\frac{\varepsilon_t}{\varepsilon_o} = \frac{T_o^4}{T_t^4}$$

we can see from this relation that when the true emissivity et is lower (higher) than the emissivity eo being assumed in deriving satellite observed brightness temperatures, the true brightness temperature of the emitting surface is warmer (colder) than the satellite-derived brightness temperature. Substituting into the relation above yields

$$\frac{\varepsilon_o + \partial\varepsilon}{\varepsilon_o} = \frac{T_o^4}{(T_o + \partial T)^4}$$

$$(\varepsilon_o + \partial\varepsilon)[(T_o + \partial T)^4] = \varepsilon_o T_o^4$$

$$\partial\varepsilon[(T_o + \partial T)^4] = \varepsilon_o T_o^4 - \varepsilon_o[(T_o + \partial T)^4]$$

$$\partial\varepsilon = \frac{\varepsilon_o T_o^4 - -\varepsilon_o[(T_o + \partial T)^4]}{(T_o + \partial T)^4}$$

The relationship described in the last equation is shown in Figure 16. describing how large of an emissivity change is associated with a temperature difference dt between a given satellite-derived temperature T and the true surface temperature T. Generally speaking, a 1 ^{o}C change in temperature is associated with around a .01 change in emissivity. This theoretical argument is important for determining how much of the temperature variability seen in the GSL could be due to emissivity variations in the water, and will be examinied in the following discussion.

The variables that most influence emissivity can be listed in the following equation (modified from Gantt with additions, 1987), where (1) is the influnce of salinity on emissivity, (2) is the influence of temperature on emissivity, (3) and (4) are the influence of inorganic and organic suspended particles on emissivity, and (7) and (8) are the influence of zenith angle and surface slicks on emissivity. The influence of (6) and (7) will not be discussed, at they are very small compared to the other terms and relate to the use of radiometer bandwidth and operating wavelength on the emissivity variations.

$$dE = \underbrace{\frac{dE}{dS}dS}_{(1)} + \underbrace{\frac{dE}{dT}dT}_{(2)} + \underbrace{\frac{dE}{dC_i}dC_i}_{(3)} + \underbrace{\frac{dE}{dC_o}dC_o}_{(4)} + \underbrace{\frac{dE}{d\Delta\lambda}d\Delta\lambda}_{(5)} + \underbrace{\frac{dE}{d\lambda}d\lambda}_{(6)} + \underbrace{\frac{dE}{d\phi}d\phi}_{(7)} + \underbrace{dE_{slicks}}_{(8)}$$

EFFECT OF TEMPERATURE ON EMISSIVITY

Term (2), which describes how emissivity changes with changing temperature, can be partly answered using the mathematical description looking at greybody changes in emissivity with changes in temperature as described above. Figure 16 shows the differences observed in emissivity for a given change in temperature as a function of temperature. In these examples, we have graphed the relationship for temperatures of 275 and 300 degrees K. These figures show that emissivity has a relatively small dependence on temperature variation. Figure 17 shows that there is still a noticable variation in emissivity across the range of observed temperatures for salty and fresh water. These emissivity variations are not directly accounted for in deriving global SST’s, although they are probably partially accounted for in deriving temperature-dependent bias-corrections for the global SST algorithms. Because of the high salinity of the GSL, the variations in spectral emissivity over the lake temperature range are surely somewhat different than that shown in Figure 17.

EFFECT OF SALINITY ON EMISSIVITY

Friedman (1969) compared the thermal characteristics of ocean water and pure water and found that salinity concentration does have a small influence on the emissivity of water. Increasing dissolved ions in the water changes the optical properties by shifing absorption lines and increasing the index of refraction. One important consideration is that the concentration of significant ions in the GSL is not the same as ocean water, with the GSL having higher values of chorine ions than is found in the ocean. Friedman’s work assumes relative concentrations of chlorinity and salinity in their calculations. It is unknown what effects this might have on the results obtained for the GSL.

Friedman's experiments detected the shift in absorption bands due to increasing the salinity of the water by using a grating spectrometer to measure the transmittance of three saline solutions. The first solution was half as salty as typical ocean water, the second contained salt concentrations comperable to ocean water, and the third contained twice the concentration of salt as the typical ocean solution. Figure 18 shows the relative transmittance of the different salt water concentrations in the 10.5-12 micron region. The shift of the relative concentration lines with different salt concentrations show an absorption band shift that appears somewhat constant with changing frequency. Thus, transmissivity appears to be a linear function of the salinity content in the thermal infrared region. A procedure was developed from these transmissivity measurements to calculate the index of refraction and extinction coefficient for oceans water. Figure 19 shows the effects of relative salt concentrations on the reflectance of ocean water. What is apparent is that linear corrections for water reflectance for a hypersaline lake like the GSL may be necessary in determining the emissivity of the lake. Friedman recommends linear extrapolation for other values. In addition, an additional correction factor is necessary for the extinction coefficient in the ir region between 11.8 and 15 microns to correct for changes in intensity of the absorption band as salt concentration increases.

Barton (1986) found that the water surface emissivity dependence on zenith angle (described in next section) and salinty for Lake Eyre, Australia, using the data given by Friedman, which is approximately as highly saline as the GSL, results in less than 1 degrees K error in surface temperature approximation. Figure 20. shows salinity versus emissivity for various zenith angles. Note that for large zenith angles, the salinity effect would become much larger, as was the case for the pure versus saline water shown in Masuda et all (1997).

The salinity effect is typically neglected in sea surface temperature retrievals. However, research in turbid oceanic regions, such as coastlines or in regions where large rivers meet the ocean, has found that the influnence of suspended organic or inorganic materials in salt water also influence the emissivity, and may need to be included in sea surface temperature retrievals along coastal waters (Wen-Yao et. all, 1987, Gantt, 1988). For the GSL, which has salinities much higher, the errors due to neglecting the salinity effect are larger than that between pure and ocean water, but still thought to be less than a degree K.

EFFECTS OF ZENITH ANGLE ON EMISSIVITY

Masuda et all (1997). compare the emissivity from pure and sea water for typical wavelengths for AVHRR thermal ir channels. It is seen that emissivity is a function of zenith angle and salinity. Within a zenith angle of 50 degrees, the differences in emissivity between fresh water and ocean water are less than 0.1%. This effect is basically due to the fact that at large satellite zenith angles, less radiation is emitted from the water surface in that direction. In addition, limb darkening due to a longer atmospheric path results in more attenuation of the outgoing radiation. For larger zenith angles, the difference becomes much larger(not shown). For zenith angles less than 60 degrees, the corresponding temperature changes are very small (order 0.1 degrees K).

EFFECTS OF SURFACE ROUGHNESS/SLICKS/FOAM ON EMISSIVITY

Several studies have found that the surface emissivity is weakly dependent on the roughness of the sea surface. Waves on the water influence the upwelling radiation field primarily by what is called shadowing, which is illustrated in Figure 21. Radiation leaving the water surface is blocked by the adjacent wave crest, such that a satellite at a far zenith angle recieves less radiation. In addition, ir radiation is reflected by the waves surfaces, altering the outgoing radiation field. Wu and Smith developed a rigorous model to determine the influence of rough sea surfaces on the emisivity of the ocean. Theri results corresponded quite well with results obtained by Masuda et all, with relatively small effects of wind speed on surface emissivity for zenith angles less than 50 degrees and winds less than 15 m/s (See Figure 6). However, the effect of the wind increases for zenith angles over 50 degrees.

Foam or oil on the water surface is known to influence emisivity, but their influence on the surface reflectance has not been documented quantatatively. In the visible region foam is known to have a strong influnce on the reflectance, resulting in the presence of a "white cap," or region of high reflectivity.

It is unknown what effect various matter floating on the GSL might influnce emissivity. Areas of concentrated Brine Fly or Brine Shrimp eggs or other organic matter may act as enough of a film on the lake to influence emissivity. In addition, the presence of phytoplankton and other organic and inorganic matter suspended in the lake may also influence emissivity, as is described in the next section.

EFFECTS OF TURBIDITY ON SURFACE EMISSIVITY

Possibly the largest error source for GSL temperature retrievals are variations in the surface emissivity of the lake due to organic and inorganic suspended sediments in the lake. More research is needed to further describe the emissivity differences that might be attributed to the types and concentrations of suspended particulates in the GSL.

The effect of turbidity on the emissivity of ocean water has been examined in several studies. The following description of the influence of suspended particulates in the lake summarizes the findings of Gantt(1988) and Wen-Yau et. all(1987). However, before we begin the description, one important point must be made: Studies have found that freshwater emissivity does not have sensitivity to sediment loading (Davies et all, 1971), whereas sea water does. It is concievable, then that the high salinity of the GSL serves to magnify the results of the lake turbidity on emissivity.

The magnitude of the influence of turbidity on GSL surface emissivity is uncertain. It is possible that the upwelling of turbid water shown in Figure 13 near the causeway of the GSL may be responsible for false temperature anomalies produced by sharp emissivity variations. Gantt (1988) hypothesized that there is a relationship between salinity and the sensitivity of lake surface emissivity to increasing organic and nonorganic sediment concentrations. If this is true, emissivity over the highly saline GSL may be highly sensitive to particulate loading in the lake. However, to our knowledge the relationship between salinity, turbidity, and emissivity has not been described or quantified. For our purposes, we will apply the documented emissivity characteristics of sea water to our discussion of their possible impacts on temperature retrievals over the GSL, with the caution that these results may not be strictly applicable to a hypersaline environment.

The following description summarizes the findings Liu-Yau shown in Table 1. Emissivity for various temperatures and organic and inorganic sediment concentrations are listed. The inorganic material consisted of grey-colored tidal creek clay-silt sediment, and the organic material was crushed marsh peat. Emissivity values were obtained using a Barnes 8-14 micron PRT-5 radiometer on prepared sea water with salinty concentrations of 35 ppm. What we see is a general decrease in emissivity as the amount of organic and inorganic matter is increased from clear sea water to 100,000 mg/l of suspended particulate. This decrease is illustrated in Figure 22. We can see that emissivity decreases more rapidly as the water becomes highly turbid. Values of turbidity in the 60,000 to 100,000 mg/L are often observed in the Yellow River in China, while values of less than 200 mg/L are typical in the Delaware Bay region of the Atlantic Ocean. Values for the GSL obtained from EPA water quality tests conducted in the south arm of the GSL show that total suspended sediment mass is still relatively low, in the 50-500 mg/L range. Within this range, there should be little impact on the derived lake surface temperatures assuming that the information in Figure 22 is applicable to the GSL. However, it is likely that under strong winds these numbers increase exponentially, and the turbidity effect may be significant,

As noted earlier, the effect of water vapor in the atmosphere is to diminish the impact of emissivity on the surface temperature retrievals. In a very dry atmosphere, however, the atmospheric influence would be minimal, and the surface emissivity variations would influence the satellite temperature retrievals. Given that emissivity decreases by approximately .01 for every 1 degree K change in emissivity, a change in turbidity from clear to extremely sediment-loaded water (100,000 mg/L) would result in a decrease of emissivity from around 0.975 to 0.945. Assuming that the satellite uses a constant value for emissivity when calculating brightness temperatures, that would result in a maximum negative bias of over 3 ^{o} C in derived surface temperature. The "causeway effect" described above has occurred with biases larger that 4 ^{o}C, during periods when sediment loading was unlikely to be anywhere near 100,000 mg/L. Thus, assuming that the emissivity variations above can be applied to the GSL, some of the the variations in temperature must be physical and not caused by emissivity variations.

ADDITIONAL SATELLITE RETRIEVAL ERROR SOURCES

In this section we describe instrument considerations and atmospheric attenuation on the outgoing infrared radiation field. Before that is decribed, the issues that result from the skin and atmospheric effects are discussed.

The coefficients used in the linear and nonlinear SST algorithms discussed in Chapter 2 are derived to compensate for a multitude of error sources (the largest of which is the aerosol effect) without quantifying or determining the source of the error. The inflence of the largest attenuator, water vapor, is quantitatively corrected for using the split-window techniques. On a global scale, deriving these coefficients using a global surface matchup bouy data set has been highly successful (Li et al. 2001). However, in some regions the globally-derived SST algorithm coefficients will yield inaccurate SST results. Shenoi found that application of the global SST algorithms contained unacceptable biases of 1-2 degrees C in the north Indian Ocean. Results were greatly improved, however, through the derivation of local calibration coefficients.

HOW APPLICABLE ARE THE OPERATIONAL SPLIT-WINDOW SST's EQUATIONS OVER THE GREAT SALT LAKE? ATMOSPHERIC AND SKIN EFFECTS

It has been found that large SST estimation errors appeared when large differences in temperature existed between air temperature and bouy SST (Robinson et al. 1984, Prata 1990, Yokoyama 1993). The continental location of the GSL makes it more susceptable to large air-water temperature differences than maritime locations. These errors are thought to be primarily caused by the atmospheric and cool skin/warm layer effects discussed in the next section.

Barton and Takashima (1986) investigated the surface temperature of the hypersaline Australian Lake Eyre and found that use of the standard lake surface temperature algorithms as discussed earlier were not applicable to the Lake Eyre region because of the large land-lake temperature difference due to advection of warm, dry air form the land over the relatively cool lake water during the day and the strong cooling of the air above the water at night. This is due to the fact that a large vertical temperature profile exists, which is not accounted for in the standard SST equations. These equations use typical marine temperature and humidity profiles when calculating the transmissivity (water vapor attenuation) with the split-window techniques. In addition, a constant mean atmospheiric temperature is used to derive SST and they assume that the air temperature at the lake-atmosphere interface is identical to the lake temperature. The atmospheric profile assumed in the global models is derived using a "geographically diverse set of atmospheric profiles... to obtain sets of surface temperatures corresponding to atmospheric brightness temperatures." These, in turn, are regressed with predicted bouy temperature data to obtain a "temperature-dependent bias correction." (McClain, 1985). However, the set of atmospheric temperature profiles used over the globe are probably not that representative of the GSL region, where very large temperature differences occur between the Lake and the air surface. Consequently, the bias correction may be completely different over the GSL than over the oceans. Instead, Barton and Takashima (1986) used a transmission model which accounted for the atmospheric profiles of temperature and moisture observed in a typical atmospheric sounding over the Lake Eyre region. The differences between the surface temperatures calculated from the transmission model and the standard split-window calculations are shown in Table 2. In general, the differences are relatively small, although occasionally they can be as large as 2 degrees C. Over the Great Salt Lake, the errors are thought to be somewhat smaller, as the land-lake temperature difference over the GSL is typically less than that over the extremely hot and arid (annual rainfall 15 cm, average summer high temperature 100 C) Australian desert. The temperature bias caused by this effect is both positive and negative, depending on the thermal profile of the water body and the temperature of the air relative to the lake.

Another possible error concern over the GSL is what is called the cool skin and warm-layer effect. The source of this error lies in the fundamental difference between where the satellite recieves radiation from, the thin sea surface layer, and the 1 meter depth at which most bouy observations are collected. Because of this observational discrepancy, the regression equations used deriving sea surface temperatures have been effectively callibrated the SST's to the bulk ocean temperature, not the thin surface layer they are actually measuring. If a constant of proportionality existed between the measured skin temperature and water a 1 m depth, then the correction would be universal and accurate. However, because of the high variability of the thermal profile in the top meter of the ocean's surface, the relationship between the ocean skin temperature and bulk 1-m temperature can complicate temperature retrievals (Robinson et al. 1984, Martin 2004).

The cool skin occurs at the molecular interface of a water surface with air and is caused by the net cooling effects of net longwave radiation and sensible and latent heat fluxes. It is always present, such that the top few millimeters of the water surface is always 0.1-0.5 degrees less than the water directly below.

The warm layer is a diurnal phenomenon, and can be on the order of several degrees or more, and is caused by the combined effects of net longwave radiation and sensible and latent heating.

Several authors have noted these effects as a possible source of error biases in satellite-derived data sets. Yokoyama et. all (1993) found that the standard deviation in errors of sea surface temperature retrievals over Mitsu Bay, Japan was decreased from .59 to .34 when bouy match-ups on days with large air-water temperature differences were removed when deriving equations to estimate SST. Since there is no network that actually measures skin temperature or the temperauture profile of the top meter of the ocean surface, quantifying the errors introduced through these methods remain theoretical, since bouy measurements are used for validation studies. Regardless of the effect of the cool skin and warm-layers on the suitability of the global SST equations for derived GSL temperature, the physics of heating and cooling of the surface water layers is of interest in determining the causes of observed temperature variability on the retrived images.

Over the oceans, the temperature difference between the bulk water temperature a meter below the surface and the surface skin temperature are typically less than a degree or two due to turbulent mixing and a modified marine boundary layer. However, over the GSL, the skin and warm-layer effect is amplified compared to the ocean because of the lake's higher salinity, lower clarity , and semi-arid continental location. The key physical factors believed to be enhancing these effects are

1). Large land-water temperature differences
2). Enhanced evaporative cooling
3). Lower specific heat capacity of the hypersaline water, which allows it to warm and cool more rapidly for given energy inputs/outputs
4). Decreased turbulent mixing due to the higher density of the lake water

One factor that may act to decrease the skin and warm layer effects over the GSL is that the wind speeds over the lake are rarely completly calm due to the presence of thermally-induced flows (J. Horel, personal communication). However, because it is more difficult to produce waves on the GSL because of the higher density of lake water, intense cooling at the surface due to the skin effect may occur even in the presence of some wind.

More research is needed to quantify the possible skin and warm layer surface temperature variations over the GSL. The lake surface layers warm very rapidly in the summer, in excess of 10 degrees in one day. It is unknown how large of cool skin is might be produced by the combined effects of a cold overlying air mass, evaporative cooling, and calm winds.

INSTRUMENT LIMITATIONS

Instrument noise and calibration is not believed to be a significant souce of error in the satellite retrievals over the GSL. Satellite calibration techniques, which determine fundamental retreival accuracy, have been the topic of significant research (Brown et al. 1985, Weinreb et al. 1990, Brown et al. 1993). Nonlinear calibration techniques have decreases instrument noise to less than 0.2 degrees Celsius (Brown et al 1993). Compared to other possible error sources, instrument effects can be neglected.

ATMOSPHERIC AEROSOL

Volcanic aerosol released into the stratosphere by eruptions like El Chicon (1982) and Pinatubo (1991) cause significant attenuation of satellite radience retreivals, leading to a cool bias of the AVHRR-derived SST's. These global-scale events are compensated for in the operational SST algorithm coefficients, which are derived by comparing bouy SST's to satellite-derived brightness temperatures. It is unknown whether increased aerosol levels in the lower atmosphere over the southern half of the GSL due to lacal anthropogenic sources will lead to a noticible bias in the derived lake temperature. Local aerosol contamination could also occur from an natural sources (i.e., forest fires, dust storms). These would lead to a cool bias for the satellite-derived SST's.

CLOUD CONTAMINATION

Clouds are filtered from SST retrievals through a variety of cloud masking tests. However, thin high clouds or low clouds with near-surface temperature are more difficult to remove, expecially in regions where large thermal gradients exist or along shorelines. (Martin 2004). Figure 9 illustrated how cirrus clouds can contaminate lake surface temperature retrievals. In addition, cloud detection is much more difficult over land and ice than over open water. Because thermal uniformity tests cannot easily be applied to the GSL without losing much of the shoreline thermal structure, there is increased potential for cloud contamination problems with a GSL SST temperature climatology than over open ocean (Martin 2004). Cloud contamination would invariably lead to a cool bias of satellite-derived SST's.

SUMMARY AND RECOMMENDATIONS:

Analysis of the thermal characteristics of the GSL show promise in defining the spatial variability of surface temperature of the lake. Four primary factors (not including sensor limitations) can influence the accuracy of satellite-derived temperature using these algorithms:

1). Vertical lake temperature structure (skin effect)
2). Land-lake temperature contrasts and improper vertical profile of temperature and moisture (atmospheric effect)
3). Aerosol attenuation
3). Emissivity variation due to suspended sediments
4). Emissivity variation due high salinity

A summary of all the possible error sources and their estimated magnitudes in deriving GSL surface temperature using AVHRR imagery are shown in Table 3. Further research is needed to determine the relative impacts of each of these on the retrieved temperatures of the GSL. In any case, a validation study of the lake temperature retrievals with in situ measurements is required to determine the accuracy of applying global SST algorithms to the GSL, as these algorithms may be unsuitable for the GSL environment. If in situ validation studies find the global split-window algorithms inaccurate in describing GSL surface temperature, here are some suggestions of methods that could be applied to compensate for those errors:

If the primary cause was due to 1) or 3), the constants applied to the split-window algorithm would need to be recalculated or estimated to provide more accurate results. There could be some difficulty in doing this over past time periods even if present observations were made available (i.e., bouy in the lake), as past lake temperature observations are not regularly available for validation and lake level, etc, changes the limnology of the GSL on a year-to-year basis.

If the primary cause of the errors was 2), then the global SST algorithms would need to be replaced with a transmission model that better approximates water vapor attenuation over the GSL under large lake-land temperature differences.

Emissivity variations due to the high salinity of the lake (4) could be compensated for by simply adding a correction term to the operational equations. If the primary cause of errors is due to emissivity variations (3) due to varying suspended organic and inorganic matter, compensating for these errors would be challenging. However, since turbidity and emissivity is directly related to the reflectance of the water, the visible channels could be used to roughly determine the lake reflectance, and then the emissivity variations could possibly be accounted for.

Because any large errors in derived lake temperature require large land-lake temperature differences and high lake turbidity, a generally accurate temperature algorithms would likely be created using a regression equation to fit GSL temperature retrievals to in situ observations-- perhaps reducing lake surface temperature errors to less than 1 degree C on averaged time scales. In this case, error sources would be reduced as the regression would implicitly compensate for much of the error introduced by the error factors listed above. This data set would have important applications for understanding climate change and land-lake-atmosphere interactions in the GSL region.

NEW REFERENCES

Assivar, R., and H. Pan, 2000. Simulations of the summer hydrometeorological process of Lake Kinneret. *Journal of Hydrometeorology*, **1,** 95-109**.**

Barton, I., 1985. Transmission model and ground-truth investigation of Satellite-derived sea sur face temperatures. *Journal of Climate and Applied Meteorology*. **24**, 508-516.

Baskin, R., 1990. Determination of ground-water inflow locations in Great Salt Lake, Utah, using the thermal infrared multispectral scanner. M.S. Thesis, University of Utah.

Barton, I., and T. Takashima. An AVHRR investigation of Surface emissivity near Lake Eyre, Australia. *Remote Sensing of Environment,* **20**, 153-163.

Beletsky, D., J. Saylor, and D. Schwab, 1999. Mean Circulation in the Great Lakes. *J. Great Lakes Research*, **25**(1), 78-93.

Beletsky, D., and D. Schwab, 2001. Modeling circulation and thermal structure in Lake Michi gan: Annual cycle and interanual variability. *Journal of Geophysical Research*, **106**, 19745-19771.

Bennett, J., 1974. On the dynamics of wind-driven lake currents. *Journal of Oceanography*. **4,** 400-414.

Bolgrien, D., N. Granin, and L. Levin, 1995. Surface temperature dynamics of Lake Baikal observed from AVHRR images. *Photogrammetric Engineering and Remote Sensing,* **61**(2), 211-216.

Bolgrien, D. and A. Brooks, 1992. Analysis of thermal features of Lake Michigan from AVHRR satellite images. *Journal of Great Lakes Research*, **18**, 259-266.

Brown, P. Minnett and W. McKeown, 1996. Observations of the infrared radiative properties of the ocean-implications for the measurement of sea surface temperature via satellite remote sensing. *Bulletin of the American Meteorological Society*, **77**(1), 41-51.

Bryant, R. and M. Rainey, 2002. Investigation of flood inundation on playas within the Zone of Chotts, using a time-series of AVHRR. *Remote Sensing of the Environment*, **82**, 360-375.

Bryant, R. Application of AVHRR to monitoring a climatically sensitive playa. Case study: Chott El Djerid, Southern Tunisia. *Earth Surface Processes and Landforms*, **24**, 283-302.

Davies, J., P. Robinson, and M. Nunez, 1971. Field determinations of surface emissivitie and tem perature for Lake Ontario. *Journal of Applied Meteorology*, ??, 811-819.

Fairall, C., E. Bradley, J. Godfrey, G. Wick, J. Edson and G. Young, 1996. Cool-skin and warm layer effects on sea surface temperature. *Journal of Geophysical Research*, **101**, 1295-1308.

Friedman, D., 1969. Infrared characteristics of ocean water (1.5-15 microns). *Applied Optics*, **8**, 2073-2078.

Gantt, R., 1989. The effect of turbidity on the infrared emissivity and thermal mapping of coastal water. University of Delaware. Ph.D. Dissertation.

Goetz, S., S. Prince, S. Goward, M. Thawley, and J. Small, 1999. Satellite remote sensing of pri mary production: an improved production efficiency modeling approach. *Ecological Model ing*, **122**, 239-255.

Hammer, U. 1986. *Saline Lake Ecosystems of the World*. Monograph Biologicae, Volume 59. H. Dumont. Ed. Dr. W. Junk Publishers.

Hook, S., F. Prata, R. Alley, A. Abtahi, R. Richards, S. Schladow, and S. Palmarsson, 2003. Retrieval of lake bulk skin temperature using Along-Track Scanning Radiometer (ATSR-2) data: A case study using Lake Tahoe, California. *Journal of Atmospheric and Oceanic Technology*, **20**, 534-548.

Kalff, J., 2002. *Limnology: Inland Water Ecosystems*. Prentice Hall.
Li, A. 1976. *A survey of the physical limnology of the Great Salt Lake*. Utah Division of Water Resources completion report.
Masuda, K., T. Takashima, and Y. Takayama, 1988. Emissivity of pure and sea waters for the model sea surface in the infrared window regions. *Remote Sensing of the Environment*, **24**, 313-329.
Mogilev, N., and R. Gnatovsky, 2003. Satellite imagery in the study of Lake Baykal surface tem peratures. *Mapping Sciences and Remote Sensing*, **40**(1), 41-50.
Oesch, D., A. Hauser, S. Wunderle, 2003. Operational processing of NOAA-AVHRR data for mapping lake surface temperatures in an alpine environment: feasibility and validation. From SPIE International Symposium Remote Sensing 2003, Barcelona, Spain.
Pan, H., R. Avissar, and D. Haidvogel, 2002. Summer circulation and temperature structure in Lake Kinneret. *Journal of Physical Oceanography*. **32**, 295-313.
Roshier, D., and R. Rumbacks, 2004. Broad-scale mapping of temporary wetlands in arid Austra lia. *Journal of Arid Environments*, ??, 249-263.
Rueda, F.,S. Schladow, S. Monismith, and M. Stacey, 2003: Dynamics of a large polymictic lake I : field observations. *Journal of Hydraulic Engineering*, **129**(2), 82-91.
Rueda, F. and S. Schladow, 2003: Dynamics of a large polymictic lake II: numerical simulations. *Journal of Hydraulic Engineering*, **129**(2), 92-101.
Richardson, T., C. Gibson, and S. Heaney, 2000. Temperature, growth and seasonal succession of phytoplankton in Lake Baikal, Siberia. *Freshwater Biology*, **44**, 431-440.
Schmugge, T., W. Kustas, J. Ritchie, T. Jackson, and A. Rango, 2002. Remote sensing in hydrol ogy. *Advances in Water Resources*, **25**, 1367-1385.
Semovski, S., 2000. Remote sensing methods in studies of Lake Baikal enviromnent. *Lake Baikal: A Mirror in Time and Space for Understanding Global Change Processes*. Minoura, K, Ed. Elsevier, 186-199.
Shevyrnogov, A., A. Kartushinsky, and G. Vysotskaya, 2002. Application of satellite data for investigation of dynamic processes in inland water bodies: Lake Shira (Khakasia, Siberia), a case study. *Aquatic Ecology*, **36**, 153-163.
Sidran, M., 1981. Broadband reflectances and emissivity of specular and rough water surfaces. *Applied Optics*. **20**, 3176-3183.
Smith, W., R. Knuteson, H. Revercomb, W. Feltz, H. Howell, W. Menzel, N. Nalli, O. Brown, J. Wen-Yau, L., R.. Field, R. Gantt, and V. Klemas, 1987. Measurement of the surface emissiv ity of turbid waters. *Remote Sensing of the Environment*, **21**, 97-109.
Stramma, L. P. Cornillon, R. Weller, J. Price, and M Briscoe. Large diurnal sea surface tempera ture variability: satellite and in situ measurements. *Journal of Physical Oceanography*. **16**, 827-?.
Wen Yau, L., R.T. Field, R. Gantt, and V. Klemas, 1987. Measurement of the surface emissivity of turbid waters. *Remote Sensing of the Environment*, **21**, 97-109.
Wick, G., W. Emery, and L. Kantha, 1996. The behavior of the bulk-skin sea surface temperature difference under varying wind speed and heat flux. *Journal of Physical Oceanography*, **26**, 1969-1988.
Wu, X. and W. Smith, 1997. Emissivity of a rough sea surface for 8-13 microns: modeling and verification. *Applied Optics*, **36**(12), 2609-2619.
Zhang, H. R. Reynolds, and T. Smith, 2004. Bias characteristics in the AVHRR sea surface tem perature. *Geophysical Research Letters*, **31**, L01307.

OLD "NEW" REFERENCES:

Anding, D. and R. Knauth, 1970: Estimation of sea surface temperature from space. *Remote Sensing of the Environment*, **1**, 217-220.

Barton, I.J., 1992: Satellite-derived sea surface temperatures--A comparison between operational, theoretical, and experimental algorithms. *Journal of Applied Meteorology*. **31**, 433-442.

Barton, I.J., 1995: Satellite-derived sea surface temperatures: current status. *Journal of Geophysical research, **100**, 8777-8790.*

Brown, O.B., J.W. Brown, and R.H. Evans, 1985: Calibration of advanced very high resolution radiometer infrared observations. *Journal of Geophysical Research*, **90**, 11667-11677.

Brown, O.B., J.W. Brown, and R.H. Evans, 1993: Calibration of advanced very high resolution radiometer infrared channels: A new approach to nonlinear correction.*Journal of Geophysical Research, **98**, 18257-18256.*

Kizu, S. and F. Sakaida, 1996: A new set of MCSST equations for NOAA-9/AVHRR. *Journal of Oceanography*. **52**, 235-249.

Li, X., Pitchel, P., Clemente-Colon, Krasnopolsky, and Sapper, J., 2001. Validation of coastal sea and lake surface temperature measurements derived from NOAA/AVHRR data. *International Journal of Remote Sensing*, **22**, 1285-1303.

Mason, C.S., Petrie, B., and B.J. Topliss, 1999: Satellite measurements of sea-surface tempera ture: An application to regional ocean climate. *NAFO Sci. Council Studies*. **32**, 45-54.

Maturi, E.M., and W.P. Pichel, Cloud masking for CoastWatch satellite imagery. in Oceans '93 Proceedings Volume II, 1993, 369-374.

Martin, S., 2004: *An Introduction to Ocean Remote Sensing.* Cambridge University Press, 426 pp.

May, D.A., M.M. Parmeter, D.S. Olszewski, and B.D. McKenzie, 1998: Operational processing of satellite sea surface temperature retrievals at the Naval Oceanographic Office. Bull. of American Meteorological Society, **79**(3), 397-407.

McClain, E.P., W.G. Pichel, and C.C. Walton, 1985: Comparative performance of advanced very high resolution radiometer-based multichannel sea surface temperatures. *J Geophys. Res*., **90**, 11587-11601

McMillin, L.M., and Crosby, D.S., 1984: Theory and validation of the multiple window sea sur face temperature technique. *Journal of Geophysical Research*, **89**, 3655-3661.

Prata, A.J., 1990: Satellite-derived evaporation from Lake Eyre, South Australia. International *Journal of Remote Sensing*. **11**, 2051-2068.

Robinson, I.S., N.C. Wells, and H. Charnock, 1984: The sea surface boundary layer and its rele vance to the measurement of sea surface temperature by airborne and spaceborne radiome ters. *International Journal of Remote Sensing.* **5**, 19-45.

Shenoi, S.C., 1999: On the suitability of algorithms for the retrieval of SST from the north Indian Ocean using NOAA/AVHRR data. *International Journal of Remote Sensing*, **20**, 11-29.

Schwab, D.J., Leshkevich, G.A., and G.C. Muhr,1998: Automated mapping of surface water tem perature of the Great Lakes. *Journal of Great Lakes Research*, **25**, 468-481.

Walton, C.C., 1988: Nonlinear multichannel algorithms for estimating sea surface temperature with AVHRR satellite data. *Journal of Applied Meteorology*, **27**, 115-124.

Yokoyama, R., S. Tanba, and S. Souma, 1993: Air-sea interacting effects to the sea surface observation by NOAA/AVHRR. *International Journal of Remote Sensing*, **14**, 2631-2646.

REFERENCES

Anati, D., 1997: The hydrography of a hypersaline lake. *The Dead Sea: The Lake and Its Setting.* T. Niemi, Z. Ben-Avraham, J. Gat Eds. Oxford Monographs on Geology and Geophysics No. 36. Oxford University Press. New York, NY. 89-103.

Baskin, R., 1998: Locating shoreline and submarine springs in two Utah lakes using thermal imagery. *Modern and Ancient Lake Systems*. J. Pitman and A. Carroll Eds. Utah Geological Association Guidebook 26. Salt Lake City, UT. 51-58.

Bates, G., F. Giorgi, S. Hostetler, 1993 : Towards the simulation of the effects of the Great Lakes on climate. *Mon. Wea. Rev*., **121**, 1373-1387.

Bussieres, N., and W. Schertzer, 2003 : The evolution of AVHRR-Derived water temperatures over lakes in the Makenzie Basin and hydrometeorological applications. *Journal of Hydrometeorology* . **4**, 660-672.

Bills, B., and coauthors, 2002: Geodynamics of Lake Bonneville. *Great Salt Lake: An Overview of Change* . J. W. Gwynn. Ed. Special Publication of the Utah Department of Natural Resources. ISBN 1-55791-667-5. 7-31.

Carpenter, D. M., 1993 : The lake effect of the Great Salt Lake: Overview and forecast problems. *Wea. Forecasting*, **8**, 181-193.

Cayan, D. R., K. Redmond, L. Riddle, 1999 : ENSO and hydrologic extremes in the western United States. *J. Clim*., **12**, 2881-2893.

Coll C. and V. Caselles, 1997 : A split-window algorithm for land surface temperature from AVHRR data: validation and algorithm comparison. *Journal of Geophysical Research*, **102**, 16697-16714.

Crosman, E., and J. Horel, 2005 : Climatology and variability of satellite-derived temperature of the Great Salt Lake. 2005 Aquatic Sciences Meeting, Amer. Soc. Limn. Ocean., Salt Lake City, February 2005.

Doran, C., J. Fast, J. Horel, 2002 : The VTMX 2000 Campaign. *Bull. Amer. Meteor. Soc*., **83**, 537-551.

Dunn, L. B., and S. V. Vasiloff, 2001: Tornadogenesis and operational considerations of the 11 August 1999 Salt Lake City tornado as seen from two different doppler radars. *Wea. Forecast ing*, **16**, 377-398.

Grover,E., P Sousounis, 2002 : The influence of large-scale flow on fall precipitation systems in the Great Lakes Basin. *J. Climate*, **15**, 1943-1956.

Gwynn, W. 2002: *Great Salt Lake: An Overview of Change*. J. W. Gwynn. Ed. Special Publica tion of the Utah Department of Natural Resources. ISBN 1-55791-667-5. 584 pp.

Hawkes, H. B., 1947: Mountain and valley winds - with special reference to the diurnal mountain winds of the Great Salt Lake region. Ph.D. dissertation, Ohio State University, 312 pp.

Higgins, R., J.-K. Schemm, A. Leetma, 2000 : Extreme precipitation events in the western United States related to tropical forcing. *J. Clim*., **13**, 793-820.

Horel, J., M. Splitt, L. Dunn, J. Pechmann, B. White, C. Ciliberti, S. Lazarus, J. Slemmer, D. Zaff, J. Burks, 2002a : MesoWest: Cooperative Mesonets in the Western United States. *Bull. Amer. Meteor. Soc.*, **83**, 211-226.

Horel, J., T. Potter, L. Dunn, W. J. Steenburgh, M. Eubank, M. Splitt, and D. J. Onton, 2002b : Weather support for the 2002 Winter Olympic and Paralympic Games. *Bull. Amer. Meteor. Soc.,* **83**, 227-240.

Horel, J., 2003: Terrain-forced mesoscale circulations. *Handbook of Weather, Climate, and*

Water: Dynamics, Climate, Physical Meteorology, Weather Systems, and Measurements. Edited by T. Potter and B. Colman. Wiley and Sons. 562-573.
Hostetler, S.W., 1995: Hydrological and thermal response of lakes to climate: Description and modeling: *Physics and Chemistry of Lakes*, 2nd ed. Lerman, A., Gat, J., and D. Imboden, Eds. Springer-Verlag Berlin. 63-82.
Isaacson, A., F. Hachman, R Robson, 2002: Economics of the Great Salt Lake. Great Salt Lake: An Overview of Change. J. W. Gwynn. Ed. Special Publication of the Utah Department of Natural Resources. ISBN 1-55791-667-5. 187-201
.Kondratyev, K., and N. Filatov, 1999: Limnology and Remote Sensing- A Contemporary Approach . Springer, London UK. 406 pp.
Kuehn, D., 2002: The brine shrimp industry in Utah. Great Salt Lake: An Overview of Change. J. W. Gwynn. Ed. Special Publication of the Utah Department of Natural Resources. ISBN 1-55791-667-5. 259-272.
Laird, N., and D. Kristovitch, 2002 : Variations of sensible and latent heat fluxes from a Great Lakes buoy and associated synoptic weather patterns. *J. Hydrometeor*., **3**, 3-12.
Laird, N., J. Walsh, and D. Kristovitch, 2003 : Model simulations examining the relationship of lake-effect morphology to lake shape, wind direction, and wind speed. *Mon. Wea. Rev*., **131**, 2102-2111.
Laird, N., and D. Kristovitch, 2004 : Comparison of observations with idealized model results for a method to resolve winter lake-effect mesoscale morphology. *Mon. Wea. Rev*.,**132**, 1093-1103.
Lall, U., and M. Mann,1995 : The Great Salt Lake: a barometer of low-frequency climatic vari ability. *Water Resources Research* , **31**, 2503-2515.
Laymon, C., Quattrochi, D., Malek, E., Hipps, L., Boettinger, J., and McCurdy, G.,1998: Remotely-sensed regional-scale evapotranspiration of a semi-arid Great Basin desert and its relationship to geomorphology, soils, and vegetation., *Geomorphol*., **21**(4-4),329-352.
Ludwig, F., J. Horel, C. D. Whiteman, 2004 : Using EOF analysis to identify important surface wind patterns in mountain valleys. *J. Climate and Appl. Meteor*., **43**, 969-983.
Malek, E., G. Bingham, G. McCurdy ,1990: Evapotranspiration from the margin and moist playa of a closed desert valley. *J. Hydrology* , **120** , 15-34.
Malek, E., Bingham, G. E., Or, D., Mccurdy, G. D., 1997: Annual mesoscale study of water bal ances in a Great Basin heterogeneous desert valley. *J. Hydrol.* , **191** 223-244.
Malek, E., McCurdy, G., and Giles, B.,1999: Dew contribution to the annual water balance in semi-arid desert valleys., *J. Arid Environ.* , **42** ,71-80.
Malek, E., Biltoft, C, Klewicki, J, and Giles, B.,2002: Evaluation of annual radiation and windi ness over a playa: Possibility of harvesting the solar and wind energies, *J. Arid Env*., **52**(4), 555-564.
Malek, E.,2003: Microclimate of a desert playa: Evaluation of annual radiation, energy, and water budgets components, *International Journal of Climatology* , **23** , 333-345.
Mann, M.E., U. Lall, B. Saltzman, 1995 : Decadal-to-centennial-scale climate variability: Insights into the rise and fall of the Great Salt Lake. *Geophysical Research Letters*. **22**, 937-940.
McCurdy, G., 1989: Radiation balance of a desert salt playa. M.S. Thesis. Utah State University. Logan.
Mesinger, F. and others, 2003: NCEP Regional Reanalyses. Symp. on Oberving and Understand ing the Variability of Water in Weather and Climate, Long Beach, CA., Feb 2003.
Moon Y., and U. Lall 1996 : Atmospheric flow indices and interranual Great Salt Lake Variabil

ity. *Journal of Hydrologic Engineering*, **1**, 55-62.
Oesch, D., A. Hauser, and S. Wunderle, 2003 : Deriving Lake Surface Temperature Variations of an Alpine Lake using NOAA - AVHRR Data. 30th International Symposium on Remote Sens ing of Environment. Honolulu, HI.
Onton, D. J., and W. J. Steenburgh, 2001 : Diagnostic and sensitivity studies of the 7 December 1998 Great Salt Lake-effect snowstorm. Mon. Wea. Rev., **129**, 1318-1338.
Physick, W., and N. Tapper, 1990 : A numerical study of circulations induced by a dry salt lake. *Mon. Wea. Rev*., **118**, 1029-1042.
Prata A.J., 1994 : Land surface temperature derived from the advanced very high resolution radi ometer and the along-track scanning radiometer 2. Experimental results and validation of AVHRR algorithms. *J Geophys. Res*., **99**, 13025--58.
Price J.C., 1984: Land surface temperature measurements from the split window bands of the NOAA 7 advanced very high resolutions radiometer. *J Geophys. Res*., **89**, 7231-7.
Rich, J., Great Salt Lake south arm circulation: currents, velocities, and influencing factors. Great Salt Lake: An Overview of Change. J. W. Gwynn. Ed. Special Publication of the Utah Department of Natural Resources. ISBN 1-55791-667-5. 171-183.
Rife, D, T. Warner, F. Chen and E. Astling. 2002 : Mechanisms for diurnal boundary layer circu lations in the Great Basin Desert. *Mon. Wea. Rev*., **130**, 921-938.
Rodionov, S., 1994: Global and Regional Climate Interaction: The Caspian Sea Experience. Kluwer Academic, Dordrecht, Netherlands. 241 pp.
Rodionov, S., 1994 : Association between winter precipitation and water level fluctuations in the Great Lakes and atmospheric circulation patterns. *J. Climate*, **7**, 1693-1706.
Rogers, E., and G. DiMego, 2001: Spring 2001 Change Package for MesoEta- CAFTI Briefing. Available at ftp://ftp.ncep.noaa.gov/pub/emc/wd20er/caftimay01/v3_document.htm
Rosen, M. R., 1994: The importance of groundwater playas: A review of playa classifications and the sedimentology and hydrology of playas. In: *Paleoclimate and Basin Evolution of Playa Systems,* The Geological Society of America Special Paper 289, M. R. Rosen (Ed.), 1-35.
Rouse, W.R., C. Oswald,J. Binyamin, P.D. Blanken, W.M. Schertzer, and C. Spence, 2003: Inter annual and Seasonal Variability of the Surface Energy Balance and Temperature of Central Great Slave Lake. *J. Hydrometeor*., **4**, 720-730.
Schmugge, T., W. Kustas, J. Ritchie, T. Jackson, A. Rango, 2002: Remote sensing in hydrology. *Advances in Water Resources*, **25**, 1367-1385.
Schultz, D. M., W. J. Steenburgh, R. J. Trapp, J. Horel, D. E. Kingsmill, L. B. Dunn, W. D. Rust, L. Cheng, A. Bansemer, J. Cox, J. Daugherty, D. P. Jorgensen, J. Meitin, L. Showell, B. F.
Smull, K. Tarp, and M. Trainor, 2002 : Understanding Utah winter storms: The Intermountain Precipitation Experiment. *Bull. Amer. Meteor. Soc*., **83**, 189-210.
Schwab, D.J., Leshkevich, G.A., and Muhr, G.C. 1992. Satellite measurements of surface temper ature in the Great Lakes: Great Lakes Coast Watch. *J. Great Lakes Res*. **18**(2), 247-258.
Scott, R. W., and F. A. Huff, 1996: Impacts of the Great Lakes on regional climate conditions. J. *Great Lakes Res.*, **22**, 845-863.
Segal, M., M. Leuthold, R. W. Arritt, C. Anderson, and J. Shen, 1997 : Small lake daytime breezes: some observational and conceptual evaluations. *Bull. Amer. Meteor. Soc*., **78**, 1135-1147.
Serreze M., M. Clark, A. Frei, 2001 : Characteristics of large snowfall events in the montane western United States as examined using snowpack telemetry (SNOTEL) data. *Wat. Res. Res*., **37**, 675-688.

Small, E., L. L. Sloan and D. Nychka. 2001a : Changes in Surface Air Temperature Caused by Desiccation of the Aral Sea. *J. Climate*, **14**, 284\u2013299.
Small, E., F. Giorgi, L. Sloan and S. Hostetler. 2001b : The Effects of Desiccation and Climatic Change on the Hydrology of the Aral Sea. *J. Climate*, **14**, 300\u2013322.
Steenburgh, W. J., S. F. Halvorson, and D. J. Onton, 2000 : Climatology of lake-effect snow storms of the Great Salt Lake. *Mon. Wea. Rev*.,**128**, 709-727.
Steenburgh, W. J., and D. J. Onton, 2001 : Multiscale analysis of the 7 December 1998 Great Salt Lake-effect snowstorm. *Mon. Wea. Rev*., **129**, 1296-1317.
Steenburgh, W. J., 2003: One hundred inches in one hundred hours: Evolution of a Wasatch Mountain winter storm cycle. *Wea. Forecasting*, **18**, 1018-1036.
Steinhorn, I., 1997: Evaporation estimate for the Dead Sea: Essential considerations for saline lakes. *The Dead Sea: The Lake and Its Setting*. T. Niemi, Z. Ben-Avraham, J. Gat Eds. Oxford Monographs on Geology and Geophysics No. 36. Oxford University Press. New York, NY. 122-132.
Stewart, J. Q., C. D. Whiteman, W. J. Steenburgh, and X. Bian, 2002 : A climatological study of thermally driven wind systems of the U.S. Intermountain West. *Bull. Amer. Meteor. Soc*., **83**, 699-708.
Stephens, D., 1990: Changes in lake levels, salinity, and the biological community of Great Salt Lake, 1847-1987. *Hydrobiologia*, **197**, 139-146.
Stowe, L., P. Davis, and E. McClain,1999 : Scientific basis and initial evaluation of the of the CLAVR-1 Global Clear Cloud Classification Algorithm for the Advanced Very High Resolu tion Radiometer. *J. Atmosph. and Oceanic Techn*., **16**, 656-681.
Tapper, N. J., 1991: Some evidence for a mesoscale thermal circulation over dry salt lakes. *Pale ography, Paleoclimatology, Paleoecology*, **84**, 259-269.
Topping, G., 2002: Great Salt Lake : An Anthology. Utah State University Press, Logan UT. 276 pp.
Wagner, F., 2003 : Preparing for a Changing Climate: Rocky Mountain/Great Basin Regional Climate Assessment. Utah State University. 244 pp. Available from http://www.cnr.usu.edu/ publications/book.pdf.
Walton, C. C., W. G. Pichel, J. F. Sapper, and D. A. May, 1998 : The development and opera tional application of nonlinear algorithms for the measurement of sea surface temperatures with the NOAA polar-orbiting environmental satellites. *J. Geophys. Res*., **103**, 999-1028.
Warner, T., 2004: Desert Meteorology. Cambridge University Press, Cambridge, 595 pp.
Whiteman, C. D., 1990: Observations of thermally developed wind systems in mountainous ter rain. *Atmospheric Processes over Complex Terrain*, Meteor. Monograph., No.45, Amer. Meteor. Soc., 5-42.
Whiteman, C. D., 2000: *Mountain Meteorology: Fundamentals and Applications*. Oxford University Press, 355 pp.
Wick, G., J. Bates, D. Scott, 2002 : Satellite and skin-layer effects on the accuracy of sea surface temperature measurements from the GOES satellites. *J. Atmos. Oceanic Techn*., **19**,1834-1848.
Zastrow, L., and M. Ridd, 2002: Satellite imaging and analysis of the Great Salt Lake. *Great Salt Lake: An Overview of Change*. J. W. Gwynn. Ed. Special Publication of the Utah Department of Natural Resources. ISBN 1-55791-667-5. 313-324.
Zumpfe, D., and J. Horel, 2005 : Lake-breeze fronts in the Salt Lake Valley. Submitted *J. Appl.* meterology.