6.3 ESTIMATING THE LAND-SURFACE RADIANT, TURBULENT AND CONDUCTIVE ENERGY BUDGETS USING SATELLITE SYSTEMS AND COMPLEMENTARY SYNOPTIC DATA

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1. INTRODUCTION

Since the advent of the meteorological satellite, a large research effort within the community of earth scientists has been directed at assessing the components of the land-surface energy balance from space. All of the energy available at the Earth's surface for heating and cooling the air (sensible heat), evaporating water from soil and vegetation (latent heat), and heating or cooling of the soil (soil conduction) is a response to the streams of solar and thermal radiation within the earth-atmosphere system. Solar radiation, the earth's only significant energy source, is partitioned into various energy fluxes at the surface. Solar radiation engenders a response in thermal-infrared (TIR) wavelengths at the surface that has important consequences for local and global energy balances.

Monitoring and modeling the energy balance at the earth's surface has applications in meteorology, climatology, hydrology, agriculture and many other disciplines. Before the age of the meteorological satellite, scientists in these fields relied heavily on data gathered from in-situ flux instrumentation or on parameterizations based on observations of standard meteorological variables, such as air temperature, humidity and clouds. Examples of in-situ instrumentation and parameterizations using standard meteorological observations to measure or estimate fluxes of solar and thermal infrared radiation are detailed in Campbell and Diak (2002).

We describe our integrated system to estimate the radiative and turbulent land-surface fluxes. The methods use space-based data sources, complementary in-situ synoptic data (rawinsondes), land-use information, a diagnostic model of the land surface and a mesoscale numerical forecast model. This system is now running in real time over the continental United States at a resolution of 10 km, producing daily and time-integrated flux components.

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2. SATELLITE ESTIMATES OF TERRESTRIAL RADIATION

2.1 SOLAR RADIATION

Estimating solar energy at the earth's surface from space data is one of the success stories of satellite meteorology. Satellite data can already provide data quality similar to ground-based measurements and at very high spatial resolution.

The radiative transfer models and associated equation sets for the evaluation of solar energy from GOES satellite data have been presented in detail in Gautier et al (1980), with updates and improvements in Diak and Gautier (1983). The method, developed for the first generation of GOES satellites (GOES-3 through GOES-7) has been successfully adapted to the current instruments, starting with GOES-8, having somewhat different visible sensor characteristics (Diak et al, 1996).

For brevity, only first principles will be described in this work. The methodology has recently been used by Bland and Clayton (1994), Lipton (1993) McNider et al. (1994), Jacobs et al. (2002) and others for calculation of solar insolation using GOES data. The physical parameterization is very simple, but has proven accurate under a variety of circumstances and computationally very efficient, an important consideration considering the large data volumes required for accurate estimates of insolation from geostationary satellites.

The simple physical model is based on conservation of radiant energy in the earth-atmosphere column. To detect the presence of cloud, a surface albedo field of the target area (within the GOES bandpass) is retained as a reference. This surface albedo estimate is made by using the GOES image closest to solar noon at each surface location within the area to calculate a surface albedo for each day (whether the location is cloudy or clear). The model for the calculation of the surface albedo accounts for the effects of ozone and water vapor absorption and Rayleigh scattering in the clear atmospheric path and within the GOES bandpass. A small empirical correction to account for aerosol and gaseous attenuations is also included. At the end of each day, the minimum surface albedo from the prior two weeks is selected at each geographic location under the presumption that it represents cloud-free conditions.

To detect the presence of clouds, this surface albedo is used along with the atmospherics and sun geometry of a given data point to estimate the digital brightness that the satellite would record if the point were clear. If the actual brightness of the data is at or below this clear "threshold", an appropriate clear model of the atmosphere is employed to calculate the surface insolation. This model of the clear atmosphere includes the effects of ozone absorption, Rayleigh scattering and also water vapor absorption, using simple bulk relationships for the entire solar spectrum (see Diak and Gautier, 1983 for details), as well as the corrections for aerosol and molecular attenuation already mentioned.

When clouds are detected via the measured brightness exceeding the estimated clear threshold, a cloud parameterization is invoked which is tuned for the middle and low cloud types that most influence the solar radiation at the surface. The parameterization generates a quadratic equation (see Diak and Gautier, 1983) in which the independent variable is the cloud albedo within the GOES visible bandpass. The cloudy radiation model used to estimate this cloud albedo, similar to the clear model, accounts for the same absorption and scattering processes within that bandpass. While satellite-based cloud information (e.g., Schreiner et al., 1993) could be utilized, the additional data burden is not strictly necessary, given that the insolation results are not very sensitive to the choice of the cloud altitude, but would offer some improvements.

After solving for the cloud albedo within the GOES bandpass, the cloudy solar radiation at the surface is calculated, taking into account the cloud albedo, and again the scattering and absorption processes mentioned above, now partitioned above and below the cloud (using bulk parameterizations applicable to the entire solar spectrum). It is also assumed that the cloud albedo calculated within the GOES bandpass can be applied to the entire solar spectrum. A cloud absorption term is also calculated as a linear function of the cloud albedo, with a maximum absorption of 7% of the incident flux at cloud top, and this term is also applied in the calculation of the surface insolation.

Currently at the University of Wisconsin-Madison, we are producing daily insolation totals for the North and South America using data from the GOES-8 (east) and GOES-10 (west) geostationary satellites at a resolution of 0.2 degrees. Fig. 1 shows an example of hourly satellite insolation estimates, compared to insolation measured at an AMERIFLUX site for 7 August 2002, Daily insolation maps for North and South America can be viewed and the data downloaded at http://www.soils.wisc.edu/wimnext/. The accuracy of the daily estimates is less than 10% versus pyranometers over all cloud conditions. Using these daily insolation estimates, we produce daily maps of potential evapotranspiration (ET_p) for the Midwestern United States (see http://www.soils.wisc.edu/wimnext/), and also use the hourly insolation estimates as forcing for the remote-sensing-based methods used to estimate land-surface fluxes to be discussed in following sections.

2.2 LONGWAVE RADIATION

The longwave method we developed is extremely simple, taking advantage of an existing real-time cloud product derived from geostationary satellite data, and employing only two pieces of additional information: values of the shelter-level (2-m) air temperature and vapor pressure (or dewpoint temperature used to calculate vapor pressure), obtained from hourly synoptic observations or a local source (Diak et al, 2000). Despite the simplicity, the accuracy of the method is competitive with other published results (see Schmetz, 1989; Ellingson, 1995).

The evaluation of clouds from GOES satellite data is based on the so-called "CO_2 Slicing" technique (Menzel et al. 1983; Schreiner et al. 1993), the outputs of which are the cloud top pressure and temperature, and the "effective" cloud fraction, F, which is the dimensionless product of the cloud areal fraction times the cloud infrared emissivity. This algorithm uses radiation measurements in three spectral channels in the CO_2 absorption band between 13µm and 15µm, and a channel in the infrared window at 11.2µm; these four channels have a 7-10 km resolution (depending on satellite viewing angle). The three channels in the CO_2 absorption band differentiate the cloud-top pressure and temperature, while the longwave infrared window channel is used to quantify the effective cloud amount. During the daytime, the visible imaging channel is also used to identify cloud presence. While the radiative transfer used to derive the cloud product from satellite sounder information is fairly complex, this complexity is transparent to the user of the product.

The method uses the GOES-derived cloud product to modify an empiricism for the downwelling longwave radiation for clear conditions (Wm^{-2}), estimated from Prata (1996), a formulation that estimates atmospheric emissivity to calculate downwelling longwave radiation. The range of atmospheric emissivity is only from about 0.60 to 0.95, depending most importantly on lower atmospheric moisture content.

The Prata formulation has a respectable RMS accuracy of between 10 and 14 Wm^{-2} in tests over various terrestrial regimes (Prata 1996) for half-hourly average fluxes. It was also the most accurate of several similar parameterizations that we investigated using pyrgeometer data taken at two Wisconsin agricultural locations in 1998. Many such simple algorithms for clear-air downwelling longwave flux have been developed in the past 40 years or so, employing various levels of empiricism (see the discussion of the historical development of such algorithms and accuracy comparisons in Prata 1996), and most produce relatively accurate results, considering their empirical or quasi-empirical nature. The general success of the methods for estimating clear-air longwave radiation relies on the fact that the majority of clear-air downwelling atmospheric longwave radiation to the earth's surface emanates from the lowest several hundred meters of the atmosphere (Prata 1996; Schmetz 1989 and others). This atmospheric region can be adequately characterized for longwave radiation estimates using low-level measurements of air temperature and humidity. The temperature structure of levels higher in the atmosphere is generally of secondary importance for clear-air longwave fluxes at the surface.

We use a general relationship for cloudy downwelling longwave radiation of the form,

$$LW_d = LW_{dc} + F\,(1 - e_{a(p)})\ . \qquad (1)$$

In Eq. 1, LW_d is the total downwelling flux including clouds (Wm^{-2}), LW_{dc} the clear-air flux from Prata (1996) and the factor F (dimensionless) is a "filling" factor. In Eq. 1, the clear-air emissivity deficit, (the difference between the clear-air emissivity [$e_{a(p)}$] and unity) is "filled" when there are clouds, e.g., $F > 0$. The factor we use is calculated using information in the GOES cloud evaluation, and is the product of the cloud temperature and cloud emissivity (see Diak et al., 2000).

This parameterization was tested in simulations using forecast-model-generated atmospheric profiles and a radiative transfer model (Diak et al., 2000). Comparisons with in-situ pyrgeometer data gathered in Wisconsin in 1998 produced a RMS accuracy of 20 Wm^{-2} for half-hourly average downwelling longwave radiation and less than 5 Wm^{-2} for 24-hour averages. Figures 2b and 2c show respectively a GOES visible image for the Wisconsin region and a corresponding map of downwelling longwave radiation calculated using the procedures detailed above. An initial use for these data was in a frost prediction system for Wisconsin (Diak et al., 1998).

3. TURBULENT FLUXES

The most common way to estimate evapotranspiration (*ET)* is to solve for the latent heat flux, *LE*, as a residual in the energy balance equation for the land surface,

$$LE = R_N - G - H\ , \qquad (2)$$

where R_N is the net radiation, G is the soil heat flux, and H is the sensible heat flux all usually given in Wm^{-2}. The quantity $R_N - G$ is commonly called the "available energy"; remote sensing methods for estimating these components are described in Kustas and Norman (1996). Using reliable remotely-sensed estimates of solar radiation, differences between remote sensing estimates and observed $R_N - G$ are within 10%.

Typically, energy balance remote sensing methods estimate the sensible heat flux in Eq. 2 through the evaluation of a surface-air temperature gradient at a single time, similar to the Bowen ratio method used in situ, only using the land-surface radiometric temperature or a derivative quantity as the lower boundary condition instead of a low-level air temperature. The aerodynamic resistance to heat transfer is largely defined by the aerodynamic roughness length and wind speed, and the land surface is treated as a single "effective" surface in contact with the atmosphere.

A robust modeling framework to address some of the limitations of first-generation methods was proposed early on in the application of satellite observations by Wetzel et al. (1984). Strictly speaking, the Wetzel et al. study was aimed at the estimation of soil moisture from remotely-sensed data, but an evaluation of surface fluxes is implicit in the scheme. The study recognized that using a time rate of change in $T_R(q)$ from a geostationary satellite, such as from the Geosynchronous Operational Environmental Satellite (GOES), coupled to an atmospheric boundary layer (ABL) model could mitigate some of the inherent problems arising from the use of single-time-level data, such as atmospheric corrections, emissivity and instrument calibration. By using time rate of change of $T_R(q)$, one reduces the need for absolute accuracy in satellite calibration, and atmospheric and emissivity corrections, all significant challenges.

Diak (1990) and Diak and Whipple (1993) implemented this approach, with a method for partitioning the available energy into *LE* and *H* by using the rate of rise of $T_R(q)$ from GOES and the growth of the ABL. The latter study included a procedure to account for effects of horizontal temperature advection and vertical motion above the ABL. The use of such an ABL energy closure method was first suggested by Diak and Stewart (1989) as a possible method to mitigate some of the problems created with the use of shelter-level variables as upper boundary conditions.

3.1 THE ATMOSPHERE-LAND EXCHANGE INVERSE MODEL (ALEXI)

Further refinements to these time-rate-of-change have been recently developed (Anderson et al., 1997; Mecikalski et al., 1999), that also use an energy closure scheme based on energy conservation within the ABL. The Atmosphere-Land-EXchange-Inverse (ALEXI) model uses a simple slab model of the time-development of the ABL in response to heat input to the lower atmosphere. A profile of atmospheric temperature at the initial time (usually from an analysis of synoptic data) serves as the upper boundary condition in atmospheric temperature. Through surface-ABL energy balance considerations and implementation of the two-source scheme for the land surface component of the model (Anderson et al., 1997, Mecikalski et al., 1999), ALEXI couples ABL development to the temporal changes in surface radiometric temperature from GOES and fraction vegetation cover (*NDVI*) from the Advanced Very High Resolution Radiometer (AVHRR). Using this ABL profile closure scheme, the methodology *evaluates* an air temperature near the top of the atmospheric surface layer (50m).

Most recently, this system has been merged with the CIMSS Regional Assimilation System (see Diak et al., 1998), so that the forecast model component of this system provides the required atmospheric profiles and a first guess at 50-m atmospheric variables to ALEXI. The method now is being used daily to evaluate fluxes over the continental United States at a resolution of 10 km for regions that are clear. The flux results are translated into moisture indices for the soil (evaporation) and vegetation (transpiration) components of the surface energy balance. For vegetation this quantity is "available moisture", a function that determines the ratio of actual to potential transpiration for a canopy. While it can be parameterized in different ways (see, for example Chen and Dudhia,1999 and Campbell and Norman, 1998), it is a quantity that is in common use in the SVAT components of numerical weather prediction models. For the soil evaporation, a similar quantity is derived from ALEXI flux estimates.

Under cloudy conditions, it is not possible to estimate surface fluxes using the ALEXI method, as a clear view of the surface is required to measure the surface radiometric temperatures. A time-continuous energy/water balance is still maintained under cloudy conditions, however, by using the satellite solar and longwave radiation estimates detailed previously, along with the latest moisture index results from ALEXI and AVHRR vegetation indices, in an aerodynamic-energy balance (a modified Priestley-Taylor scheme) estimate of vegetation transpiration. The moisture availability calculated for the soil is used to estimate soil surface evaporation under cloudy conditions. Figures 3a and 3b show daily integrals of net radiation and latent heat, respectively (both MJm^{-2}), for the continental United States on a day in July 2002. A typical summer value of daily incident solar radiation under clear skies in mid latitudes is about 28-30 MJm^{-2} at mid latitudes. The latent heat flux shown in Fig. 3b is a response to the net radiation shown in Fig. 3a and the moisture indices for the soil and vegetation carried by ALEXI. The low values of evapotranspiration in the eastern part of the United States are a response to the clouds and resulting low net radiation in that area. In the west, where the values of net radiation are uniformly high, low evapotranspiration values are caused by low vegetation amounts and low soil moisture content.

3.2 "DISAGGREGATING" ALEXI FLUXES TO SMALLER SPATIAL SCALES

Kustas and Norman (2000) found that sub-pixel variability in surface properties can result in large errors in pixel-average heat flux estimation using pixel-average inputs when there are significant discontinuities in surface conditions, and particularly with low wind speeds. A solution to the problem of spatial resolution has been introduced by Norman et al. (2002), who developed a scheme (called DisALEXI) for "disaggregating" ALEXI 5-10-km flux estimates to the 30-m scale using high-resolution NDVI and $T_R(q)$ data, and the local 50-m air temperature estimate provided by ALEXI as the important atmospheric boundary condition in temperature. Although this scheme makes use of energy conservation principles applied to ABL dynamics to deduce air temperature via ALEXI, it still does not consider local variability in mean air properties at the disaggregation scale.

Preliminary results are encouraging, however, suggesting that disaggregation of coarse spatial resolution *ET* output may be feasible using high-spatial-resolution data from aircraft, the Land Remote-Sensing Satellite (Landsat), the Advanced Space-borne Thermal Emission Reflectance Radiometer (ASTER) or the Moderate-Resolution Imaging Spectroradiometer (MODIS) (Norman et al., 2002). Figure 4 shows a comparison of fluxes measured in-situ for three days in July (during the SGP97 experiment) with fluxes estimated using the DisALEXI technique, and incorporating high-resolution thermal and vegetation (24-m) data taken by aircraft. For all flux components, the root-mean-square-deviation (RMSD) between model estimates and measurements is 38 Wm^{-2}. This compares well with the observational accuracy of 20-40 Wm^{-2} typically associated with eddy-covariance flux measurements (Twine et al., 2000).

One logistical drawback of the disaggregation method using data from Landsat or ASTER, however, is their temporal infrequency of coverage of a specific location (at best sixteen days). The images also need to be clear during the overpass, reducing the probability of getting usable data. MODIS provides a higher temporal frequency (daily), but with lower horizontal resolution (maximum of 250 m in the visible and near infrared and 1 km in the thermal infrared).

4. CONCLUSIONS AND FUTURE WORK

Estimating solar energy at the earth's surface from space data is one of the success stories of satellite meteorology. Satellite data can already provide data quality similar to ground-based measurements and at very high spatial resolution. Some improvements still may be possible, however, through a better use of other satellite data sources that can refine knowledge on the physical and optical properties of clouds for such insolation schemes. Any satellite or in situ data that could be used to estimate the cloud base altitude/temperature for the calculation of downwelling longwave radiation would be of some value to schemes based in satellite measurements.

Current efforts incorporating remote sensing data into SVAT modeling schemes that accommodate the fundamental differences between aerodynamic and radiometric temperatures and that are not sensitive to measurement errors should greatly enhance the prospect of quantifying *ET* at regional scales with remote sensing. The measurement errors with the largest impact on *ET* estimation are atmospheric and emissivity effects in converting satellite brightness temperatures to radiometric surface temperatures and assigning meteorological variables, primarily air temperature, for each satellite pixel from regional weather station observations. Due to limited spatial observations of atmospheric properties, the uncertainty in the surface-air temperature difference is likely to be several degrees resulting in unreliable *ET* estimation, which has significantly hampered many past modeling approaches.

Although the ALEXI approach described here, addresses most of limitations of thermal infrared methods of estimating surface fluxes, there is a drawback to these schemes in that the source of radiometric temperatures (GOES), and the atmospheric boundary layer closure schemes mandate an output resolution of about 5 to 10 km. For many applications, particularly evaluating *ET* for individual fields, these 5-10 km estimates are of limited use. Unfortunately, useable temporal changes of satellite brightness temperatures are only available from GOES at a minimum resolution of several km. Other satellites/instruments, such as Landsat and ASTER, have much finer spatial resolution, but have much coarser temporal coverage (at best 16 days). As stated, data from the MODIS instrument may offer additional possibilities for disaggregation and will certainly offer better quality estimates of vegetation amounts than have been previously possible.

Currently, the coupling of the CIMSS forecast system and ALEXI is in one direction, meaning that the forecast model provides initial atmospheric profile conditions and other meteorological inputs to ALEXI, but no there is no feedback of ALEXI to the forecast model. In the near future, the fields of available water produced by ALEXI, as well as satellite-estimated hourly insolation, will be used as forcing to this forecast model, completing the loop, and hopefully providing better inputs to ALEXI, as well as improved forecasts of the ABL structure.

Lastly, a prototype satellite interferometric sounding instrument is scheduled for launch in late 2005. This instrument, the Geosynchronous Imaging Fourier Transform Sounder (GIFTS), with thousands of channels in the thermal infrared spectral region, will provide new capabilities for estimating turbulent fluxes at the land surface. In simulation studies Diak et al. (1994) demonstrated that these data can provide a more direct estimate of the daily change in the heat content (fluxes) of the ABL than current methods (discussed), that rely on sophisticated land surface models to interpret surface radiometric temperatures and estimate fluxes. In this study, superior flux estimates were obtained using a relatively simple statistical (eigenvalue) analysis of the spectroscopic data versus land-surface fluxes.

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7. FIGURE CAPTIONS

Figure 1. Comparison of hourly satellite estimates of insolation (circles) with those

measured at an AMERIFLUX station (boxes) at Fort Peck, Montana on 7 August 2002.

Figure 2(a-b). Figure 2a is a GOES visible image for 1400UTC on 12 May 2000. Figure 2b is downwelling longwave radiation (WM-2) for the same region estimated through the methods outlined in this paper.

Figure 3(a-b). Figure 2a show a map of net radiation derived via the methods outlined in this paper for the continental United States for 26 August 2002 at a resolution of 10 km. Figure 2b is a map of latent heating for the same day, derived using the net radiation in Fig. 3a and the ALEXI methods outlined in this paper.

Figure 4. Comparison of in-situ measurements of net radiation (plus sign), latent heat (hexagons), sensible heat (diamonds) and soil heat (squares) with DisALEXI results for three days during the SGP97 experiment.

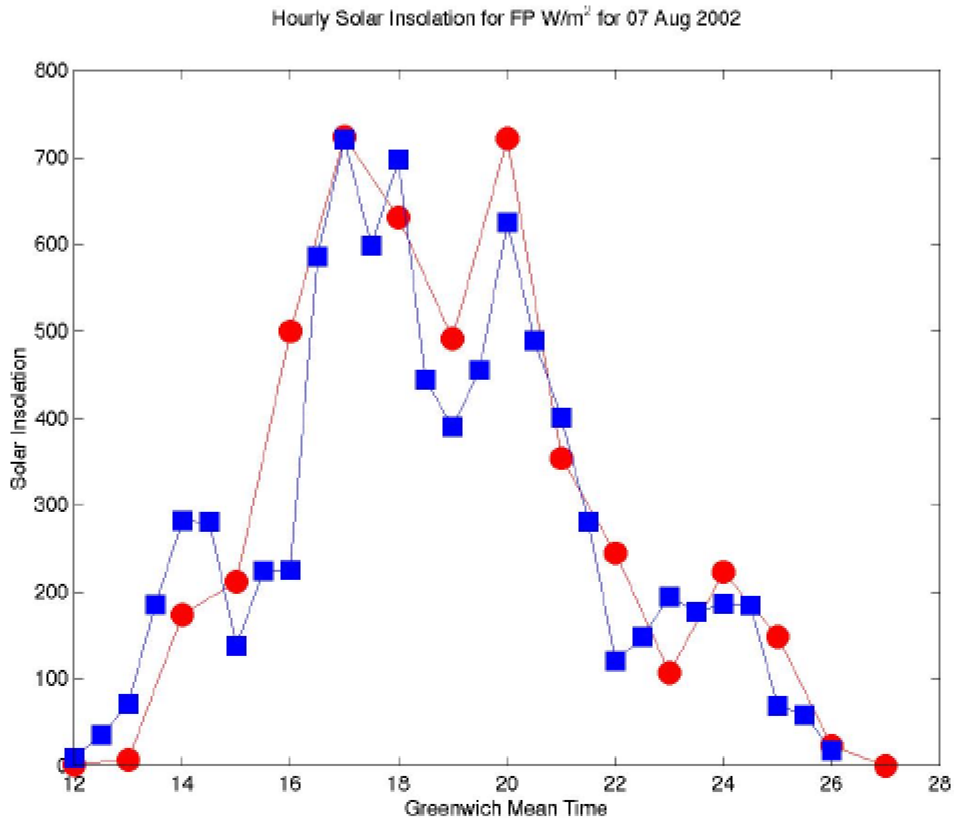


Figure 1

Figure 2a

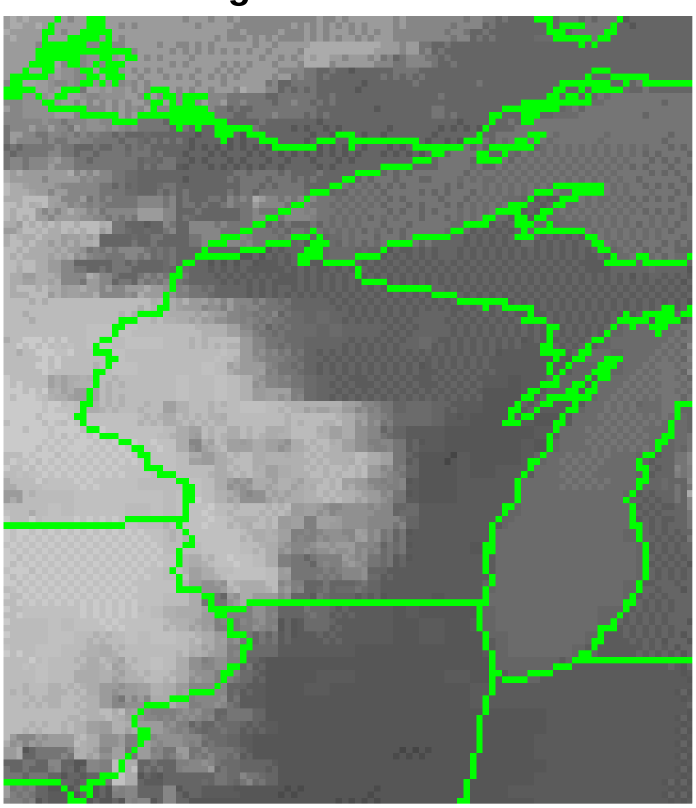
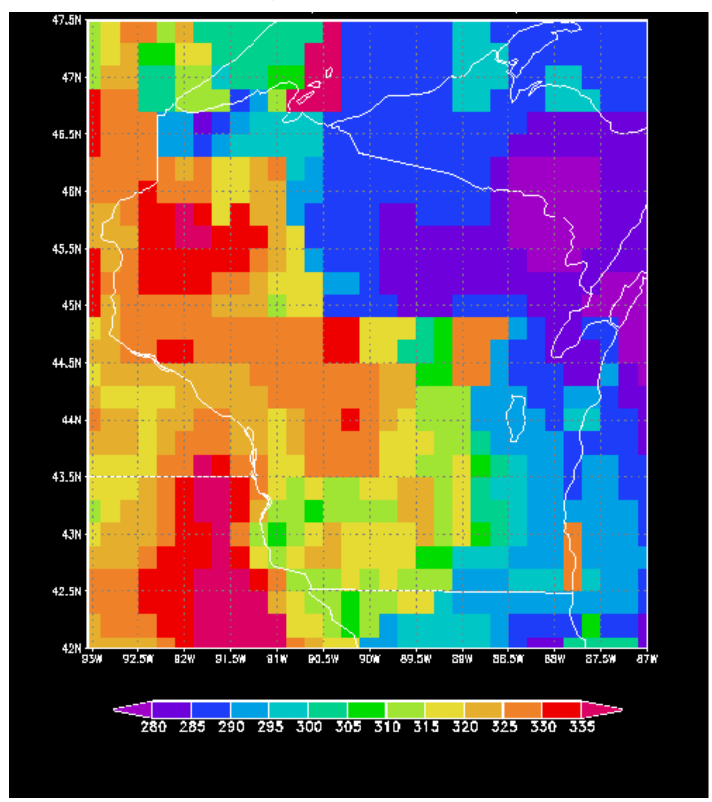


Figure 2b



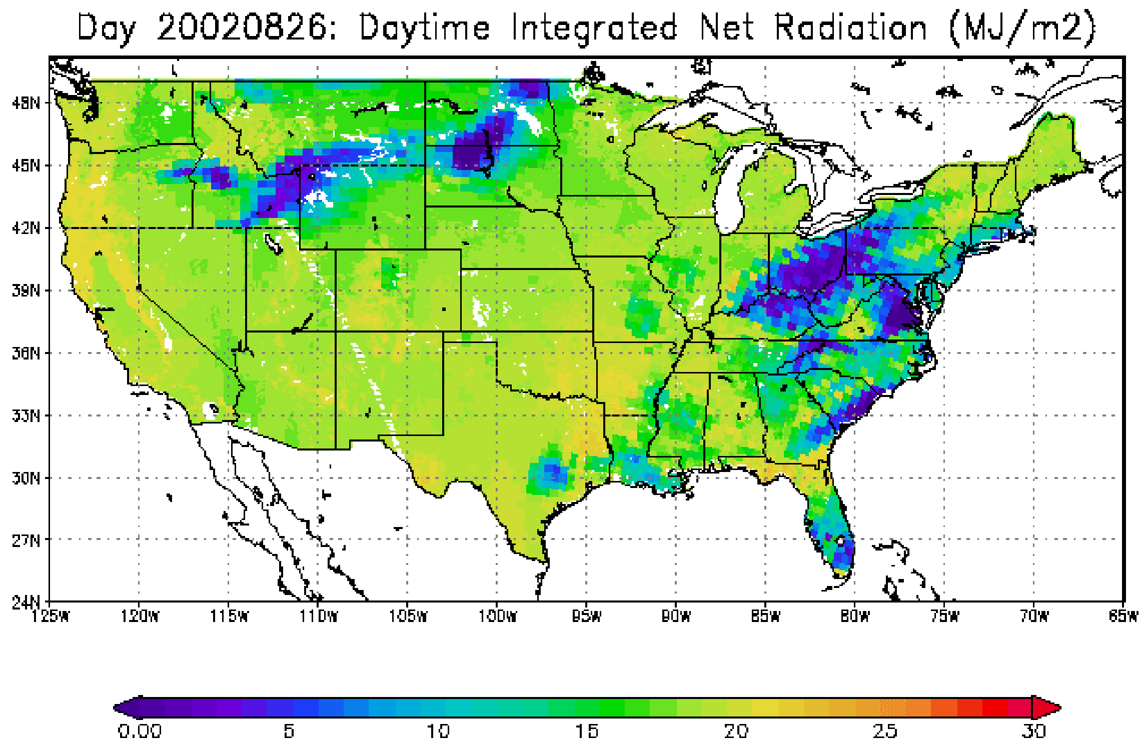


Figure 3a

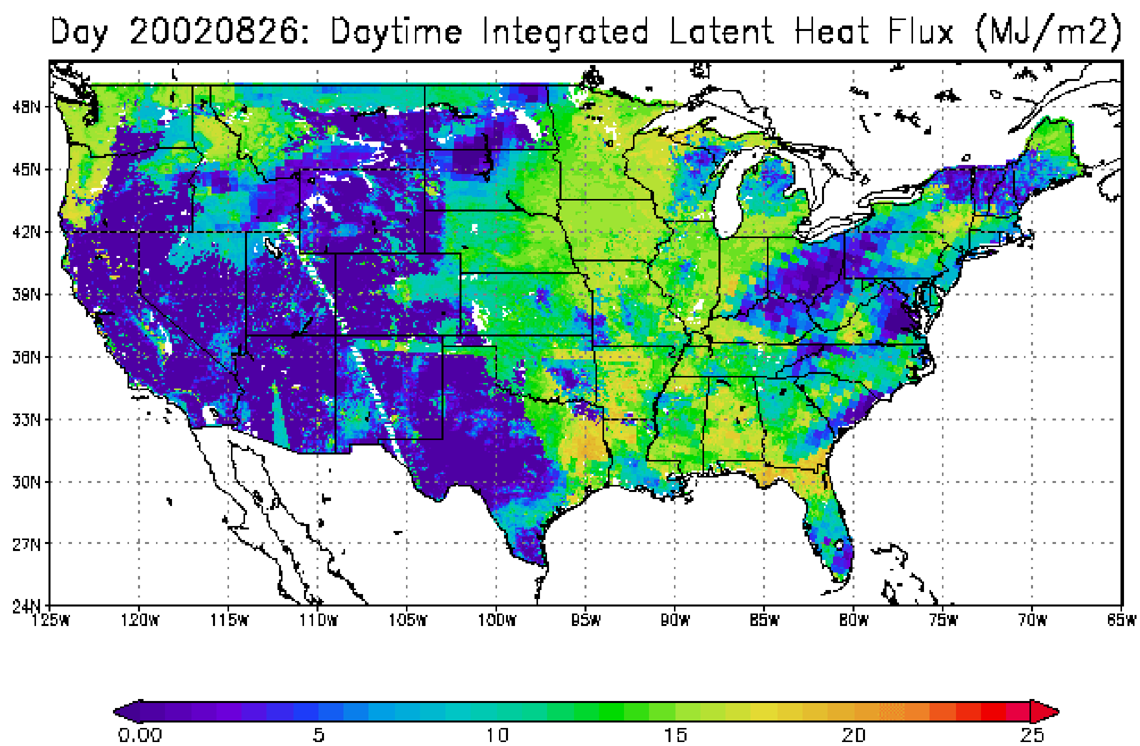
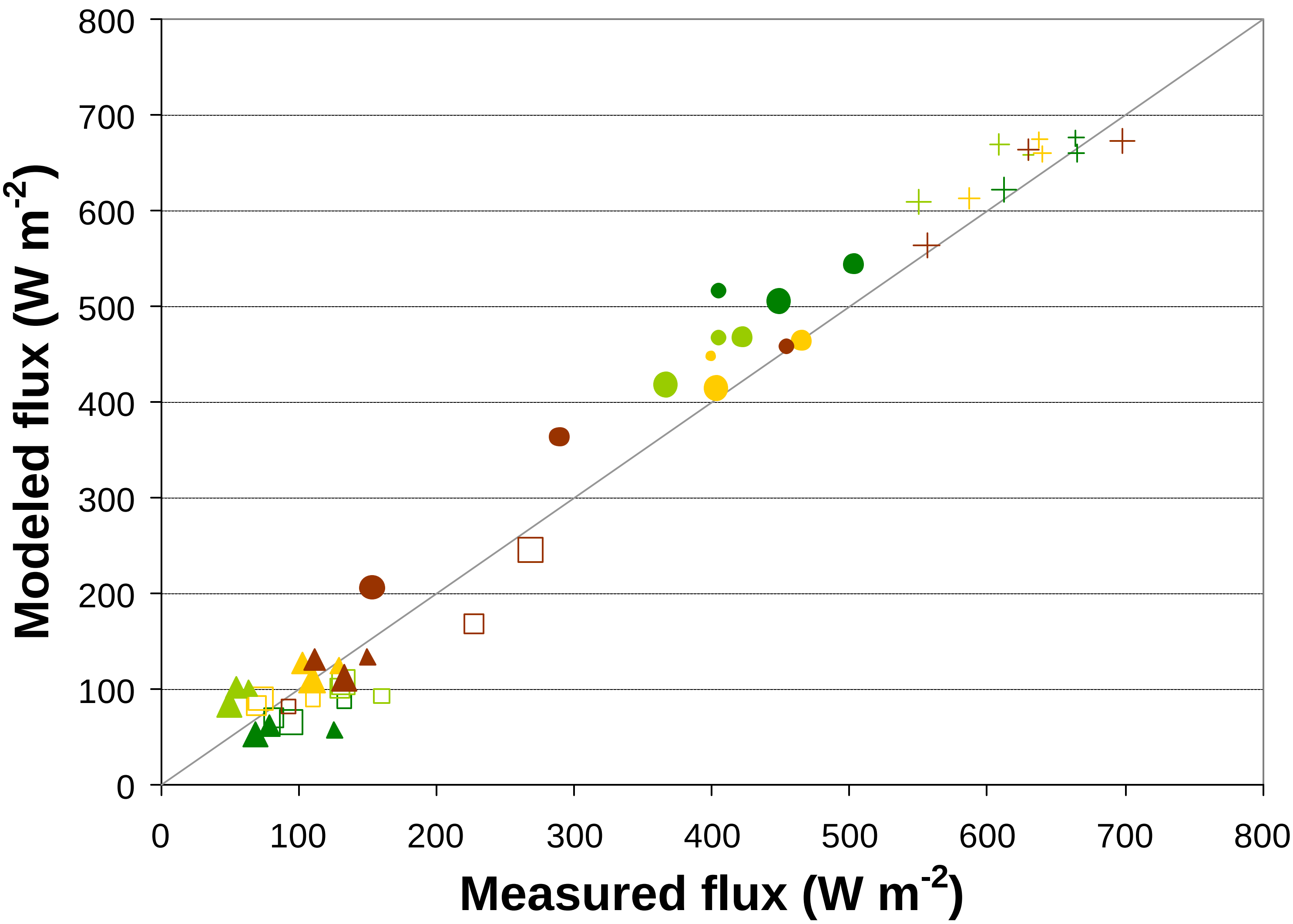


Figure 3b



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800
Measured flux (W m^{-2})
Modeled flux (W m^{-2})

Figure 4