EVALUATING THE INFLUENCE OF VEGETATION ON EVAPOTRANSPIRATION FROM WASTE ROCK SURFACES IN THE ELK VALLEY, BRITISH COLUMBIA

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ABSTRACT

Coal mines operated by Teck Coal Ltd., in the Elk Valley, British Columbia, utilize the method of surface mining, resulting in large waste rock piles that affect water quality. In order to limit the influence that these waste rock piles have on water-rock interaction, alternate management strategies are being explored. In this study, the influence of vegetation on evapotranspiration is examined, as potential benefits exist in using vegetation to reduce the infiltration and percolation of water into waste rock. During the 2013 growing season, energy and water balance components were measured using the eddy covariance technique at a bare waste rock surface, a waste rock surface with a vegetated grass cover, and a waste rock surface with a reclaimed forest cover. In addition, other water balance components were measured, such as soil water storage and precipitation. The placement of vegetation atop the waste rock pile allowed for increased evapotranspiration compared to the bare waste rock surface. From 23 May 2013 to 30 September 2013, the reclaimed forest and reclaimed grasses site experienced 305 mm and 272 mm of ET, respectively, while the bare waste rock site had only 140 mm of ET. This increase in evapotranspiration suggests less deep percolation at vegetated sites, estimated as 148 mm, 172 mm and 246 mm for the grass, forest and bare rock sites respectively. ET at the vegetated sites was dominantly controlled by the net radiation, while the near surface moisture was the dominant control on ET at the bare rock site.

Results from this study suggest future reclamation projects should consider placing surface vegetation as a potential method to reduce deep percolation into waste rock piles.

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LIST OF ABBREVIATIONS AND SYMBOLS

	Units	Description
D	kPa	the vapour pressure deficit
ET	mm	evapotranspiration
G	$W\ m^{-2}$	the soil/ground heat flux
H	$W\ m^{-2}$	the sensible heat flux
LE	$W\ m^{-2}$	the latent heat flux
LAI	$m^2\ m^{-2}$	the leaf area index
L_v	$J\ kg^{-1}$	the latent heat of vapourization
NP	mm	the net percolation
P	mm	precipitation
r_a	$s\ m^{-1}$	the aerodynamic resistance
Rn	$W\ m^{-2}$	the net radiation
r_s	$s\ m^{-1}$	the surface resistance
S	mm	the soil moisture storage
s	$kPa\ K^{-1}$	the slope of the saturation water vapour versus temperature curve
T	°C	the air temperature
w	$m\ s^{-1}$	wind velocity
z	m	the height above the surface
β		the Bowen Ratio, calculated as H/LE
κ	$m^2\ s^{-1}$	the molecular diffusion coefficient in air
$\rho_a C_p$	$J\ m^{-3}\ K^{-1}$	the specific heat of air
ρ_v	$g\ m^{-3}$	the density of vapour

γ	kPa K^{-1}	the psychrometric constant
’		prime, the instantaneous deviation from the mean
−		overbar, denotes a time average

CHAPTER 1: INTRODUCTION

1.1 Coal Mining in the Elk Valley

The Elk Valley, located in southeastern British Columbia, is the location of five open pit mines operated by Teck Coal Ltd. Coal mining in the Elk Valley began in 1897, with large-scale mining beginning in the 1960's (Lussier et al., 2003). These mines utilize the practice of surface mining, an efficient and widespread method also used globally, and in North America, for the mining of coal in the Appalachian region of the United States (Kerem et al., 2007). This process involves the removal of overburden (rocks, soil, and vegetation), followed by the use of explosives and heavy machinery in order to access the underlying coal seams (Hendryx et al., 2012). The unused, non-coal, materials (mine spoils) that were removed are then placed in the adjacent valleys, potentially leading to the burial of headwater streams (Lindberg et al., 2011). The five open pit mines in the Elk Valley create approximately 140 x 10^6 tonnes of waste rock each year, as estimated in 2003 (Lussier et al., 2003). Many environmental problems arise from the use of this mining technique, particularly in regards to the placement of large spoil piles in river valleys. These spoil piles have a large influence on water quality, especially downstream where studies have found consequences such as increases in salinity and contaminants (Lindberg et al., 2011). In 1995 an assessment was conducted for a mine effluent permit amendment that found soluble selenium was being transported from the mines into the Elk River (McDonald and Strosher, 1998). It was also discovered that the waste rock piles were allowing for easier passage of water from upslope locations, enhancing the ability for contaminants, such as soluble selenium, to be transported to surface waters

(McDonald and Stosher, 1998). In 2011, a research program was initiated in order to improve understanding of the integrated biogeochemistry and hydrology of spoils, their effect on watershed functioning, and the implications of different management strategies as an effective method to reduce selenium loading to the surrounding ecosystem. The program aims for a better understanding of the integrated pathways of water, solutes, and controlling biogeochemical mechanisms within spoils. One aspect of this research program is to examine the influence of vegetation on evapotranspiration from coal spoil surfaces. This aspect of the program will examine the potential benefits of placing vegetation on coal spoils as a method to limit infiltration and percolation into and through the waste-rock, as a potential mechanism to reduce water-rock interaction and contaminant loading.

1.1 Reclamation of Coal Mines

In order to reduce the amount of leaching, a reduction in the percolation of water (and associated constituents) through the waste rock piles would be required. One potential way to decrease the volume of water that is percolating through the piles is to increase the amount of evapotranspiration that is occurring, reducing the volume of water available for deep percolation. In order to increase the amount of evapotranspiration occurring, re-vegetation of the site would need to be completed.

In the Unites States the Surface Mining Control and Reclamation Act (SMCRA) was established in the late 1970's, placing importance on reclaiming land that is able to support native vegetation (Zipper et al., 2013). The general performance standard 525 (b)

(19) requires mine operators to "establish on the regarded areas, and all other lands affected, a diverse, effective, and permanent vegetative cover of the same seasonal variety native to the area of land to be affected and capable of self-regeneration and plant succession at least equal in extent of cover to the natural vegetation…". In order for mine operators to successfully complete the re-vegetation requirements, the SMCRA instituted general performance standard 515 (b) (5), stating mine operators must "…remove the topsoil from the land in a separate layer, replace it on the backfill area…except if topsoil is of insufficient quantity or of poor quality for sustaining vegetation, or is other strata can be shown to be more suitable for vegetation requirements, then the operator shall remove, segregate, and preserve in a like matter such other strata which is best able to support vegetation". However, unlike the United States, there is no regulation stating soil must be salvaged, or a suitable replacement used, when reclaiming coal mines in British Columbia. This means that at the coal mines operated by Teck Coal Ltd. in the Elk Valley, no soil layer is present on the surface of the spoil pile prior to seeding. It is therefore unknown how effective the vegetated covers will be at reducing the amount of water available for deep percolation, through evapotranspiration.

There are several key reasons why salvaging soil and placing it back on the mine spoil prior to seeding is considered advantageous. One of the main reasons is due to the low water holding capacity of bare waste rock (Showalter et al., 2010). Bare waste rock has a low moisture retention capacity due to both the lack of fine particles and the presence of coarse rock fragments (Ames, 1980). The coarse rock fragments lead to the presence of large voids, enhancing the leaching of precipitation through the spoil piles and reducing the amount of moisture available at the surface for vegetation growth (Ames, 1980). Conversely, natural soils contain organic matter, allowing for an increased soil water holding capacity and the presence of the near surface soil moisture that is required for vegetation growth (Zipper et al., 2011). The lack of organic matter in the waste rock also means that there are low levels of essential nutrients, such as nitrogen and phosphorus, for plant uptake (Showalter et al., 2010). When soil is unsalvageable a substitute, favorably made up of weathered mine spoil material that meets certain chemical standards, is used in its place to ensure standards are met (Showalter et al., 2010; Zipper et al., 2013).

1.3 Study Objectives

This study used a comparative approach to examine surface-atmosphere exchanges at a forested, grass covered and bare mine spoil sites for the 2013 growing season. The objectives of the study were to:

1. Determine the amount of evapotranspiration occurring at each of the sites
2. Evaluate the control of biological and environmental variables on the evapotranspiration occurring at each site
3. Assess the effectiveness of using a vegetated cover as a method to reduce percolation through waste rock piles

The results of this study will help aid in the development of improved reclamation strategies, as understanding the effect of vegetation on the partitioning of water is essential for their design. As well, this study will provide further scientific knowledge on the effect of mine spoils on the movement of water and associated constituents. There currently is no evapotranspiration data from coal spoils in various stages of vegetation at Teck Coal sites, so this study will address the lack of knowledge in this area and provide the framework for future studies.

1.4 The Surface Energy Balance

The evaporation of water from a given surface can be thought of in terms of the surface energy balance, allowing for a basic representation of energy exchange between the surface and atmosphere. When determining the surface energy balance both longwave and shortwave radiation are highly important variables, which can have a large influence on the amount of evaporation (Baldocchi et al., 2000). The longwave and shortwave radiation is used to calculate the latent and sensible heat flux densities, as well as soil storage and canopy storage flux densities (Baldocchi et al., 2000). These terms are used to make up the surface energy balance equation:

$$Rn=LE+H+G+S \tag{1.1}$$

where Rn is the net radiation (W m^{-2}), LE represents the latent heat flux (W m^{-2}), H is the sensible heat flux (W m^{-2}), G symbolizes the soil conductive heat flux (W m^{-2}), and S represents the canopy storage heat flux (W m^{-2}) (Baldocchi et al., 2000). Often the canopy storage heat flux is assumed to be negligible and is not included in energy balance calculations. The net all-wave radiation (Rn) is the most important energy exchange, as it represents the limit to the available energy (Oke, 1987). The latent and sensible heat fluxes describe convective exchange to or from the atmosphere (Oke, 1987). The latent heat flux (LE) and the transfer of vapour is partly governed by saturation vapour pressure, specifically the difference between the surface saturation vapour pressure and the atmospheric saturation vapour pressure (e_s-e_a) (Penman, 1948), and is expressed as:

$$LE = \rho L_v \kappa_v \frac{\Delta \rho_v}{\Delta z} \tag{1.2}$$

where ρ is the density (kg m^{-3}), L_v is the latent heat of vapourization (J kg^{-1}), z represents the height above the surface (m), and κ_v is a molecular diffusion coefficient in air for water vapour (m^2 s^{-1}) (Oke 1987). Conversely, the sensible heat flux (H) is governed by the difference between the surface temperature and the air temperature (Ts-Ta) (Penman, 1948):

$$H = -\rho c_p \kappa_h \frac{\Delta T}{\Delta z} \tag{1.3}$$

where c_p is the specific heat of air at constant pressure (J kg^{-1} K^{-1}), T is the air temperature (°C), and κ_h represents the molecular diffusion coefficient in air for heat (m^2 s^{-1}) (Oke 1987). Generally, Rn is positive when the surface gains energy, and the non-radiative fluxes (LE, H, G) are positive when they are directed away from the surface (Oke, 1987). The storage term (S) is positive when there is an accumulation of energy within the system. All of these flux terms will vary both spatially and temporally as a result of fluctuation in solar energy, soil and air temperature, soil moisture, soil texture, photosynthetic capacity and leaf area (Baldocchi & Vogel, 1997).

The Bowen ratio (β) describes the partitioning of radiation into sensible and latent heat fluxes, and is defined as:

$$H/LE = \beta = \gamma(T_s - T_a)/(e_s - e_a) \tag{1.4}$$

where γ is the psychometric constant (kPa K^{-1}) (Penman, 1948). When the value of β is less than one, a greater proportion of energy is being transferred to the atmosphere through the latent heat flux. When the value of β is greater than one, the sensible heat flux of energy is more dominant than the latent heat flux. (Monteith, 1965). Bowen ratio values that exceed one are common over boreal forests and occur when the surface conductance is small relative to the aerodynamic conductance (Baldocchi & Vogel, 1997). Negative β values occur when a downward flux of sensible heat leads to the evaporation of intercepted rainfall (Humphreys et al., 2003). This condition is observed in forested regions with a mild climate, where frequent, light rainfall occurs (Humphreys et al., 2003).

The surface energy balance and associated principles acts as the base for more complex evaporation models. The energy balance model examines the overall fluxes of energy between the surface and the atmosphere and does not specifically identify some of the main controls on these fluxes. It is for this reason that the surface energy balance model is often combined with other more specific models, such as the Penman-Monteith model, in order to provide a more detailed evaluation of evaporation (Monteith, 1965).

1.5 The Penman-Monteith Method

The Penman-Monteith model is likely the most widely used evaporation model, expanding on the energy balance principles to identify and incorporate important biotic and abiotic factors controlling evaporation (Baldocchi et al., 2000). The method combines both mass and energy balances, and is based on fundamental physical properties, making the method universally valid (Chen et al., 2005). The model uses the "big leaf" concept, treating the canopy as one "big leaf" with a bulk stomatal resistance and a bulk aerodynamic resistance (Raupach and Finnegan, 1988). The model also requires the input of several meteorological variables, limiting the use of this method to studies where these are available (Chen et al., 2005). Evapotranspiration measurements calculated by the Penman-Monteith method are considered to be very reliable and are often used as a reference to compare the results of other models (Chen et al., 2005) The Penman-Monteith model is expressed as:

$$LE = \frac{s(Rn - G - S) + \rho_{\mathrm{a}} C_{\mathrm{p}} D g_{\mathrm{a}}}{s + \gamma(1 + \frac{1/r_{\mathrm{a}}}{1/r_{\mathrm{s}}})} \tag{1.5}$$

where s is the slope of the saturation water vapour versus temperature curve (kPa K^{-1}), $\rho_{\mathrm{a}} C_{\mathrm{p}}$ is the specific heat of air (J m^{-3} K^{-1}), D is the vapour pressure deficit (kPa), r_a is the aerodynamic resistance (s m^{-1}), and r_s is the surface resistance (s m^{-1}). In order to determine the influence of the surface resistance on the latent heat flux, and thus evapotranspiration, and inverted Penman-Monteith equation is typically used as LE is measured directly (Humphreys et al., 2003). The stomata of leaves allow for, and control, the movement of water that is contained in the plant to the atmosphere. Given the important role that stomata play in evapotranspiration, stomatal resistance is considered to be a key governing factor of evapotranspiration rates. It is therefore important to examine the influence of stomatal resistance on evapotranspiration at each site, as well as the potential controls on the stomatal resistance, as they indirectly influence ET.

1.6 The Eddy Covariance Method

The eddy covariance technique allows for the *in situ* measurement of the transport of mass and energy, allowing for the calculation of turbulent fluxes within the atmospheric boundary layer (Burba, 2005). The transport of mass and energy throughout the atmosphere occurs as the result of the continual movement of turbulent eddies, and is described as a vertical flux (Oke, 1987). Each eddy contains a mass, a vertical velocity, and the volumetric content of various entities, with each having a mean value and a fluctuation (Oke, 1987). The turbulent flux of a given entity across a horizontal place is proportional to the mean covariance of the vertical velocity and the fluctuation of the concentration of the entity (Baldocchi and Meyers, 1991). The eddy covariance method measures the instantaneous covariance of the vertical wind speed and a given entity from the mean value (Oke, 1987).

The method is derived from the conservation equation, where the mean covariances between wind velocity components (*w*) and the mixing ratio of a chemical constituent (*x*) represent the fluxes (Baldocchi et al., 1988). The application of several assumptions, as well as averaging procedures according to Reynolds, allows the flux to be expressed as:

$$F = -\bar{\rho}_a \overline{w' \; x'} \tag{1.6}$$

where ρ_a is the density of dry air (Baldocchi et al., 1988). Given this base flux equation the equations to calculate the latent and sensible heat fluxes can be derived, and are expressed as:

$$H = C_a \overline{w' \; T'} \tag{1.7}$$

$$LE = L_v \overline{w' \; \rho'_v} \tag{1.8}$$

where C_a is the heat capacity of the air, w is the vertical wind velocity (m s^{-1}), T is the air temperature (°C), Lv is the latent heat of vaporization (J g^{-1}), and ρ_v is the vapour density (g m^{-3}). The prime symbol (‘) expresses deviation from the mean and the overbar denotes a time average (Oke, 1987).

The eddy covariance technique relies on multiple assumptions in order to accurately measure turbulent transport. One assumption is that the surface being measured is horizontally homogenous and flat, meaning no advection occurs, and there is no horizontal divergence of fluxes (Baldocchi, 2003). It is also assumed that the vegetation on the underlying surface extends upwind for an adequate distance, allowing for an acceptable fetch and footprint (Baldocchi, 2003). Another assumption is that the air density fluctuation is negligible, as a result of constant air density in the lower atmosphere (Oke, 1987). The presence of sloping hills or non-uniform terrain can cause a divergence or convergence of airflow over the surface, as the mean vertical velocity will be non-zero, causing the eddy covariance framework to become invalid (Baldocchi et al., 1988; Baldocchi, 2003). A solution for the violation of this assumption is the mathematical rotation of the wind coordinate system that would force the vertical wind speed to zero (Baldocchi, 2003). This would allow for the computation of flux covariances that are orthogonal to the mean airflow streamlines (Baldocchi, 2003). The presence of flat, horizontally homogenous terrain within the footprint of the eddy covariance tower would prevent the advection from occurring, and would allow the internal boundary layer to be fully adjusted (Baldocchi et al., 1988). When setting up the tower over a homogenous surface there is a general rule that 100 m of fetch is needed for every 1 m the tower is above the effective surface (Baldocchi et al., 1988). It has also been observed that while a surface may be horizontally homogeneous, the active exchange of heat and water vapour may lead to a fluctuation in the density of dry air, causing the mean vertical velocity to become non-zero (Webb et al., 1980). Webb et al (1980) developed an equation for a corrected vertical turbulent flux, however it was also noted that this correction is not needed for the sensible heat flux, and is generally small for fluxes of water vapour.

1.7 The Boreal Region

The Elk Valley, British Columbia, is located within the montane cordillera ecozone. However, the tree species present throughout the region, in particular on the study sites, are similar to species native to the boreal region. Given the similarities between the regions, and the large amount of literature examining boreal forests, this study will use the results from the boreal region for comparison with results from the Elk Valley. The boreal region is the world's second largest biome and represents the circumpolar region between 50 and 70 degrees North (Baldocchi et al., 1997). The northern geographical extent of the boreal forest is determined by the July 10 degree climatic isotherm (Saugier et al., 1997). Two of the key distinguishing features of this region are the short growing season and the extremely cold winter temperatures (Baldocchi et al., 2000). Over the course of the year the air temperatures can range from -70 to 30 degrees Celsius, while in most areas of the boreal region the growing season lasts less than 120 days (Baldocchi et al., 2000). Boreal regions often have low precipitation input, exemplified by the observation that the Canadian boreal region receives between 200 and 600 mm of precipitation per year (Baldocchi et al., 2000). Due to the geologically recent presence of glaciers, there is low species diversity throughout the boreal region, as well as a slow rate at which the species migrate and evolve (Baldocchi et al., 2000). The Canadian boreal region consists of evergreen and broad-leaved conifer forests, as well as fens and lakes (Baldocchi et al., 1997). Two of the main conifer species present in the boreal forest are spruce, grown on wet soils with limited drainage, and pine trees, often established on well drained and nutrient poor soils (Baldocchi et al., 2000).

The dark colour of the evergreen forest canopy, combined with the low solar angles, allows for the absorption of most incoming solar radiation (Baldocchi et al., 2000). During the growing season the ratio between the incoming solar radiation and the net radiation is between 0.70 and 0.80, meaning 70 to 80% of the incoming solar radiation is available to be partitioned into sensible and latent heat fluxes (Baldocchi et al., 2000). The latent and sensible heat fluxes over a boreal stand change based on variations in solar energy, soil and air temperature, soil moisture, photosynthetic capacity, and leaf area throughout the growing season (Baldocchi et al., 2000). Deciduous trees partition the available energy mainly into sensible heat in the period before leaf emergence, and mainly into latent heat once leafs have emerged (Blanken et al., 1997). Conversely, leafless conifer trees mainly partition the available energy into sensible heat. (Blanken et al., 1997).

While the comparison of results from the Elk Valley with results from the boreal region is useful, it is important to consider that these regions are subject to different amounts of precipitation. The boreal region is characterized by low precipitation (Baldocchi et al., 2000), while the montane cordillera region experiences higher amounts of precipitation, particularly during winter months. The higher precipitation in the montane cordillera region results in a different water balance than sites within the boreal region. At natural boreal sites a large amount of the precipitation is evaporated back into the atmosphere, with some studies resulting in a water balance where evaporation exceeds precipitation.

In the montane cordillera region the larger amount of precipitation allows for more water to be available for infiltration, as a lesser proportion is being evaporated compared to boreal sites. This is generally measured over the period of an entire year, due to the large influence of winter snowfall on the total precipitation. However, more water would also be made available for infiltration if a large amount of precipitation were to occur over a shorter period of time, such as during a large rainfall event. While this study only examines the months of May to September, the fact that the inclusion of winter data would result in a greater difference between total precipitation and total evapotranspiration should be noted. It should also be noted that the interannual and interseasonal variation in the pattern and amount of precipitation received at a particular site would result in different water balances over different study periods.

1.8 The Boreal Energy Balance

1.8.1 Boreal Forests

Many studies completed in the boreal region have found that transpiration from coniferous forests is largely controlled by canopy conductance (Cienciala et al., 1997). The characteristic low precipitation and low temperatures can limit the growth of trees in many boreal forests (Baldocchi et al., 2000). This can lead to the formation of canopies with low leaf area indexes (LAI), and therefore a reduced photosynthetic capacity and lower stomatal conductance (Baldocchi et al., 2000). This results in the exertion of a significant amount of resistance to transpiration and an overall reduction in the total evapotranspiration rate of the system (Baldocchi et al., 2000). Over boreal forests, transpiration rates are often found to be quite low despite the required supply of energy needed to drive transpiration (Saugier et al., 1997). Conifer stands utilize between one third to one half of the net radiation for ET (Baldocchi & Vogel, 1996). Saugier et al., (1997) suggested that some of the time less than one third of the incoming radiation was used for evapotranspiration, suggesting that latent heat fluxes over a boreal forest are weakly coupled to the amount of available energy. Over a boreal jack pine forest most of the incoming radiation was used for sensible heat fluxes, as opposed to latent heat fluxes (Baldocchi & Vogel, 1996). Baldocchi and Vogel (1996) found that at the boreal pine forest Bowen ratio values approached and exceeded 3, exemplifying the use of radiation for sensible heat fluxes as compared to latent heat fluxes. Other than the influence of the available energy it was also reported that changes in the vapour pressure deficit can significantly influence evapotranspiration as Baldocchi and Vogel (1996) found that over a boreal pine forest an increase in the vapour pressure deficit of the surrounding air lead to a negative feedback on evapotranspiration, resulting from stomatal closure.

The seasonal pattern of evaporation from a boreal coniferous forest is affected by changes in air and soil temperature, insolation, vapour pressure deficits, soil moisture and LAI (Baldocchi & Vogel, 1996). During the spring, evaporation rates over the conifers is relatively low due to the fact that the cold or still frozen soils resist root uptake of soil moisture (Baldocchi et al., 2000). This results in a low soil hydraulic conductivity, which leads to closing of the stomata (Baldocchi & Vogel, 1996). From May until August, the day-to-day variation in evaporation over the boreal conifer forest results from the presence of absence of clouds and the changes in atmospheric humidity associated with moving fronts (Baldocchi et al., 2000). Daily evaporation values generally peak at the end of June and during July (Baldocchi et al., 1997), coinciding with the summer solstice and peak net radiation (Amiro and Wuschke, 1987). Evaporation rates have been observed to decline once again starting in September at a boreal spruce forest, with the presence of frost reducing the evaporation rates to values similar to those measured in the spring (Betts et al. 1999). Arain et al. (2003) also observed a decrease in evapotranspiration rates during August and September at a boreal spruce forest, and Baldocchi et al. (1997) observed a gradual downtrend in the latent heat flux after a peak in July at a jack pine forest.

Baldocchi et al. (1997) reported that the daily evaporation from a jack pine forest in the Canadian boreal region ranged between 0.5 and 2.5 mm d^{-1}, with peak values approaching 3.5 mm d^{-1} after a period of significant rainfall. At a Manitoba jack pine forest examined by Ewers et al. (2005) a peak rate of 3.5 mm d^{-1} was observed. Moore et al. (2000) observed an average daily evapotranspiration rate of 1.1 mm d^{-1} at a Manitoba jack pine forest and an average daily rate of 1.6 mm d^{-1} at a Saskatchewan jack pine forest. The maximum daily evapotranspiration rate at a northern Saskatchewan black spruce forest was 3.5 mm d^{-1} during the summer. The average evapotranspiration rate at a southern Saskatchewan black spruce forest was 2.0 mm d^{-1} during the 1994 growing season. Moore et al. (2000) observed an average daily evapotranspiration rate of 1.9 mm d^{-1} at a Saskatchewan black spruce forest. At a Manitoba black spruce forest Ewers et al. (2005)

observed a maximum evapotranspiration rate of 1.3 mm d^{-1}. Also at a Manitoba black spruce forest Lafleur (1992) observed average daily evapotranspiration rates of 2.24 mm d^{-1} and 2.33 mm d^{-1} in 1989 and 1990, respectively. A Manitoba black spruce forest examined by Moore et al. (2000) experienced an average daily evapotranspiration rate of 1.3 mm d^{-1}.

Jarvis and McNaughton (1986) found that the coupling of the latent heat flux to available energy decreased as the aerodynamic resistance decreased and the surface resistance increased. This results in the latent heat flux of a dry canopy being dependent on and limited by surface resistance (Jarvis & McNaughton, 1986). The surface resistance in boreal forests is often large, and is affected by short term factors, such as partial stomatal closure due to soil moisture deficits and high vapour pressure deficits, and long term factors, such as low values of the leaf area index and less photosynthetic capacity (Baldocchi et al., 2000). Daytime mean surface conductance increased during the growing season and then declined rapidly during the fall at a black spruce forest examined by Arain et al. (2003). Values were observed to increase with air temperature, and decrease as available energy declined (Arain et al., 2003). Daily canopy conductance was observed to exhibit a diurnal pattern in response to changes in the vapour pressure deficit, with maximum conductance occurring when the vapour pressure deficit was small (Blanken et al., 1997). A similar result was observed by Arain et al. (2003), as surface conductance was found to decline midday, when the vapour pressure deficit was increased. Humphreys et al. (2003) observed that the vapour pressure deficit was the most important environmental variable influencing the canopy conductance, and decreased canopy conductance resulted in less evapotranspiration.

Lagergren and Lindroth (2002) found that both pine and spruce stands were influenced by vapour pressure deficits, a relationship observed in many boreal forests. Ewers et al. (2005) observed for both jack pine and black spruce forests the vapour pressure deficit was responsible for most of the variability in daily transpiration rates. Humphreys et al. (2003) observed that a Douglas-fir forest the high aerodynamic roughness of the canopy resulted in the forest being well coupled to the vapour pressure deficit. Arain et al. (2003) also observed a relationship between evapotranspiration and the vapour pressure deficit. This result was expected given the high aerodynamic conductance of the forest as a result of roughness. While not the dominant control, Bernier et al. (2006) also observed that variation in the vapour pressure deficit resulted in variation in transpiration rates at a jack pine forest.

Changes in soil moisture were also found to result in a significant amount of change in transpiration, at jack pine and black spruce sites examined by Ewers et al. (2005). Humphreys et al. (2003) also observed that low soil moisture appeared to reduce evapotranspiration. During a year with low precipitation, and subsequently low soil moisture, evapotranspiration was observed to be 8% lower at a black spruce forest and 7% lower at a jack pine forest that during a wet year (Zha et al. 2010). At a boreal aspen forest Barr et al. (2007) observed decreases in evapotranspiration as a result of the depletion of root zone soil water, and determined that soil water content was a control on ET.

Radiation was observed to exhibit the greatest control on transpiration at a jack pine forest examined by Bernier et al. (2006). Humphreys et al. (2003) observed that higher transpiration rates in the early summer were partly the result of greater amounts of radiation, along with warmer temperatures and a higher vapour pressure deficit. During summer days when the vapour pressure deficit was low Arain et al. (2003) observed that at a black spruce forest the latent heat flux clearly responded to changes in the net radiation.

Whether a canopy is wet or dry can also have a significant impact on the evapotranspiration rate (Baldocchi & Vogel, 1996). In general, ET rates from wet stands are larger than those seen in dry stands by factors of 50% or greater (Baldocchi et al., 2000). On clear days after a period of substantial rainfall there is an increased amount of available energy, and the evaporation rates from the wet stands approach and exceed the potential evaporation rates (Baldocchi & Vogel, 1996). In extreme cases the evaporation from a wet surface can exceed the limit set by the amount of available energy (Baldocchi et al., 2000). The rate of evaporation from a wet surface can be calculated using the rate of increase in the latent heat content of the air surrounding the canopy (Monteith, 1965). Wet leaf surfaces can become cooler than the surrounding air, causing sensible heat to be removed from the overlying air and thereby initiating evaporation through a non-radiative source of energy (Baldocchi et al., 2000). When a canopy is dry the control of water loss is through the physiological process of stomatal closure, which deceases with canopy conductance (Humphreys et al., 2003). When the canopy is saturated stomatal closure, and thus, the canopy conductance, no longer controls the transfer of water through evaporation (Humphreys et al., 2003). This means that the rate of evaporation is only controlled by the atmospheric humidity deficit, energy inputs, and the turbulence needed to transport the heat and water vapour between the atmosphere and the canopy (Humphreys et al., 2003). Another difference between dry and wet stands is the daytime Bowen ratio (Jarvis et al., 1976). The daytime Bowen ratio of a dry canopy generally ranges between 0.1 and 1.5 depending on the species and conditions, with some studies reporting values approaching 3 (Jarvis et al., 1976;Baldocchi & Vogel, 1996). For wet canopies the Bowen ratio generally ranges between -0.7 and 0.4, suggesting that the latent heat flux is more prominent in wet canopies than in dry canopies, as supported by numerous studies (Jarvis et al., 1976).

1.8.2 Boreal Grasslands

One option for the reclamation of mine spoils is the placement of grasses atop the spoil surface, rather than forests. It is therefore important to understand the energy balance of grasslands and the main controls on evapotranspiration from these sites. Grassland ecosystems have a large interannual variation in productivity, primarily resulting from changes in precipitation (Wever et al., 2002). Vegetation productivity and evapotranspiration are strongly related, so a large interannual variation in evapotranspiration is common over grasslands (Wever et al., 2002). Vegetation productivity is also related to the leaf area index, and a positive relationship between evapotranspiration and LAI has been observed over a Lethbridge grassland (Zha et al., 2010). The grasslands studied in Lethbridge undergo emergence and senescence, so seasonal changes in the leaf area index would result in seasonal changes in evapotranspiration. A significant positive relationship was observed between the leaf area index and evapotranspiration from July to September at the Lethbridge grassland (Zha et al., 2010). At a grassland site in Japan the latent heat flux increased with LAI until the canopy closure occurred when the LAI reached 3 (Li et al., 2005). Wever et al. (2002) found that there was a correlation between precipitation and evapotranspiration at a northern temperate grassland site, as well. The peak evapotranspiration rate over a year with above average precipitation was 4.5 mm d^{-1} , whereas the peak rate was only 3 mm d^{-1} during the years where there was average or below average precipitation (Wever et al., 2002). Zha et al. (2010) observed that during a dry year the annual evapotranspiration was 55% lower than during a wet year. Ponton et al. (2006) observed a maximum evapotranspiration rate of 5 mm d^{-1}, also at a Lethbridge grassland. The summer soil moisture was observed to be a dominant control on evapotranspiration at the Lethbridge grassland, along with spring soil temperatures (Zha et al., 2010. Li et al. (2005) observed at a grassland in Japan that variations in the latent heat flux, and thus evapotranspiration, were primarily dependent on changes in the amount of available energy. It was also observed that the relative contribution of available energy to the latent heat flux was larger than contribution from advective energy imposed by the vapour pressure deficit (Li et al., 2005). Canopy evapotranspiration is also largely dependent on the surface conductance, as larger surface conductance is associated with larger latent heat fluxes, and thus larger evapotranspiration (Li et al., 2005).

As well as influencing evapotranspiration, the amount of precipitation that fell on a grassland was also observed to affect radiation partitioning. At the Lethbridge grassland site, the only time in which the latent heat flux was greater than the sensible heat flux was during the year with above average precipitation, during the other study years the sensible heat flux dominated the energy balance (Wever et al., 2002). At a montane grassland in Austria the latent heat flux dominated the energy partitioning, consuming 54% of the net radiation based on midday means (Hammerle et al., 2008). The latent heat flux also dominated the energy budget throughout the growing season at a grassland in Japan (Li et al., 2005). After snowmelt the ground heat flux increased rapidly, consuming up to 25% of the net radiation. It was observed that when the leaf area index was low the latent heat flux, sensible heat flux, and ground heat flux all consumed a comparable fraction of the net radiation (Hammerle et al., 2008). As the leaf area index increased the fraction of the latent heat flux increased, the ground heat flux decreased, and the sensible heat flux decreased initially and then increased at higher LAI values (Hammerle et al., 2008). At the Austrian grassland, changes in the air temperature and the vapour pressure deficit were responsible for approximately 40% and 30% of the variation in the partitioning of net radiation, respectively. Diurnally, the latent heat flux roughly followed the net radiation at a grassland in Japan, with peak values observed one to two hours after the peak in net radiation (Li et al., 2005). It was also observed that the latent heat flux followed a similar pattern to the vapour pressure deficit.

CHAPTER 2: STUDY SITES AND METHODS

2.1 Study Sites

This study takes place in the Elk Valley, British Columbia, Canada, where Teck Coal Ltd. operates 5 coal mines. The Elk Valley is located in the southeastern corner of the province, west of Cranbrook and south of Calgary, Alberta. For this study, three sites were chosen: a reclaimed site with forest cover, a reclaimed site with grass cover, and a bare waste rock site.

2.1.1 Forest (Reclaimed) site

This study site (50° 09'55"N, 114° 52'44"W, 1705 m a.s.l.) is located on the Teck Coal Ltd. Fording River Operations mine in southeastern British Columbia. The site is located on eastward sloping terrain (21.6°) on the side of a mine spoil pile. A naturally growing forest non-uniformly covers the site in all directions. The forest consists of Lodgepole pine (*Pinus contorta*) and Engelmann spruce (*Picea engelmannii*) stands, with a mean height of 4.03 ± 0.09 m and a mean stand diameter of 6.23 ± 0.15 cm. The average leaf area index (LAI) of the site is 2.79 m^2 m^{-2}, with a standard deviation of 1.19 m^2 m^{-2}. The understory vegetation consists of bird's-foot trefoil (*Lotus corniculatus*), screw-moss (*Tortula sp.*), thread-moss (*Bryum sp.*), golden ragged-moss (*Brachythecium salebrosum*), redtop (*Agrostis gigantea*), great mullein (*Verbascum Thapsus*), and grasses. The top 1 m of waste rock on which the vegetation has grown consists of 73.8% coarse gravel and cobble, 19.5% sand, and 6.7% clay and silt.

2.1.2 Grasses (Reclaimed) site

This study site (49° 56’48”N, 114° 47’43”W, 2070 m a.s.l) is located on the Line Creek Operations mine site operated by Teck Coal Ltd., south of the Fording River Operations mine. The reclaimed grasses site is located on the slightly sloping (13°) south-southwest (SSW) side of a spoil pile. The surface of the site is dominantly covered by alfalfa (*Meducago sativa*), hard fescue (*Festuca trachyphylla*), thickspike wildrye (*Elymus lanceolatus*), and Festuca *rubra*, all of which are agronomic species. The native nodding thread-moss (*Pohlia nutans*) is also present, however this species does not cover a significant proportion of the site and is only seen in small quantities. The average LAI of this site is 1.82 m^2 m^{-2} , with a standard deviation of 0.79 m^2 m^{-2}. The top 1m of the waste rock consists of 54.7% coarse gravel and cobble, 38.9% sand, and 6.4% clay and silt.

2.1.3 Bare Rock site

The bare rock site (49° 56’58”N, 114° 47’38”W, 2146 m a.s.l.) is also located at the Line Creek Operations mine site, approximately 500m north of the reclaimed grass covered site. The site is located on gently (<5 °) sloping terrain facing the east-southeast direction. The surface is composed of bare waste rock, with no vegetation present. The top 1m of the waste rock at this site consists of 77.4% gravel and cobble, 18.2% sand, and 4.4% clay and silt.

2.2 Eddy Covariance

2.2.1 Measurements

To measure the exchange of energy and water vapour between the spoil surfaces and the atmosphere the eddy covariance technique was used (Baldocchi et al., 1988). The data used for the eddy covariance measurements of flux were collected with a combination of LICOR Li-7700 enclosed-path (vegetated sites) and Li-7500a open-path CO_2/H_2O analyzers (bare rock site) (LICOR Biosciences). Net radiation was measured using a net radiometer (CNR4, Kipp & Zonen). Air temperature and relative humidity were measured using an HMP45 sensor (Vaisala). Three components of wind speed were measured using a sonic anemometer (Windmaster, Gill Instruments Ltd.), with the sensor head on the sonic anemometer installed adjacent to the gas analyzers intake and optical path as required by the manufacturer's specifications. High frequency measurements of the wind components, water vapour and CO_2 density, air temperature, and atmospheric pressure were taken at 10 Hz. A value was recorded every 30 minutes at the reclaimed forest and reclaimed grasses sites, and every hour at the bare rock site, and stored on USB flash drives using a LICOR Li-7550 interface unit.

In addition, ground heat flux was measured using a heat flux plate placed at 0.05m depth at the reclaimed forest and grasses sites. The soil moisture content (CS-616, Campbell Scientific), soil temperature (107-L, Campbell Scientific), and soil suction (CS-229, Campbell Scientific) were measured using sensors placed at depths of 5cm, 10cm 20cm, 30cm, 40cm, 50cm, 75cm, and 100cm at all sites.

Data was collected from 23 May until 16 September at the bare rock site, from 23 May until 1 October at the reclaimed forest site, and from 22 May until 1 October at the reclaimed grasses site.

2.2.2 Data Processing

Eddy covariance fluxes were calculated at a half-hourly time step using the EddyPro software (Version 5.1, LICOR Biosciences). The sonic anemometer data was rotated using the Wilczak et al (2001) method and all signals were detrended using 5-minute block averages. Density fluctuations were accounted for by applying the Webb Pearman Leuning Correction (Webb et al., 1980) and time lags between the signals were accounted for using the maximum covariance method. Fluxes were calculated without application of either low- or high-pass spectral corrections so the calculated fluxes likely represent an underestimate of the true flux. Once the half-hourly fluxes were computed the data was cleaned by removing the half hours where the sensor indicated a failure, the occurrence of precipitation, or low friction velocities.

2.3 Biometric Assessment

Leaf area index, which is the ratio of the total leaf area of vegetation to the area of the ground surface, was measured at each site. Measurements were taken at the reclaimed grasses site between 10 September 2013 and 17 September 2013, and were taken at the reclaimed forest site on 12 July 2013 and 13 July 2013. The LAI at both sites was measured using the LI-COR LAI 2200 plant canopy analyzer. The height and diameter of the trees, as well as the species composition of the site were also measured at the reclaimed forest site.

2.3.1 Forest (Reclaimed) site

The LAI 2200 measures the attenuation of diffuse sky radiation; with the amount of foliage being calculated from measurements of how quickly radiation is attenuated as it passes through the canopy (LI-COR, 2010). Measurements were taken along transects through the forest, starting at the same elevation as the eddy covariance tower and ending at the furthest extent of the forest. Overall, sixteen transects were completed. Canopy transmittance was calculated using measurements from above and below the canopy, which was converted to LAI by the LAI 2200. The trees were too tall to be able to place the sensor above them, so above canopy measurements were taken in an unobstructed clearing above the forest. Below canopy measurements were taken every meter along each transect.

As shown in the LAI 2200 manual, measurements taken by the instrument underestimate the true LAI, as a result of foliage clumping. This is especially true for measurements taken from conifer stands where needles tend to concentrate around stems. The clumping factor varies by species, meaning that a correction factor applied to LAI measurements would need to account for the proportion of each species along the transects. The proportion of each species along each measurement transect was not evaluated, and therefore a correction factor was not applied to LAI measurements. Chen (1996) determined that measurements of LAI for conifers may experience an error of 15-40%, of which 3-10% of that error originates from clumping. Other errors in the LAI measurements may be due to the ratio of woody area, the ratio of needle to shoot area, and instrumental errors (Chen, 1996).

2.3.2 Grasses (Reclaimed) site

At the grass covered site LAI measurements were taken at 5 sampling plots, with 40 measurements taken at each plot. At each plot measurements were taken at evenly spaced intervals between 1 and 5 meters from the plot center, along 8 radial transects. Measurements were taken at a height of 2 to 5 centimeters from the ground surface. Measurements were averaged to provide one LAI value for each plot. Above canopy measurements were taken at least 8 times at each plot. To compute the data the FV-2200 software (LI-COR, 2013) was used, with above and below canopy measurements matched by closest record in time and abnormal transmittances clipped. Results were corrected for sunlight scattering using Kobayashi et al.,'s model (1993), and were computed using the FV-2200 software. Measurements and calculations were made by Integral Ecology Group Ltd.

CHAPTER 3: RESULTS

3.1 Climate

3.1.1 Air Temperature

Air temperatures during the 2013 growing season were warmer than normal (Table 3.1.1). At the nearby Sparwood weather station, operated by Environment Canada, the average daily temperature from 1 May to 30 September was 14.3 °C, whereas the average temperature according to the 1981-2010 climate normal is 12.7 °C. In addition, the average temperature was higher at the Sparwood weather station for each month in measurement period compared to the 1981-2010 climate normal for the same month. There is a distinct seasonal trend in daily average air temperature at each study site, with the warmest temperatures recorded at the reclaimed forest site (Figure 3.1.1). The average daily air temperature at the forest site was 12°C, and was 9°C at both the reclaimed grasses and bare rock sites. At each of the sites the highest daily average air temperature was recorded on 2 July, measuring 24°C at the forest site and 22°C at both the grasses and bare rock sites. The coldest average daily air temperature was recorded on 1 May at each site, with a temperature of -1°C at the forest site, and a temperature of -4°C at both the grasses and bare rock sites.

3.1.2 Precipitation

Over the 2013 growing season 526 mm, 512 mm, and 437 mm of precipitation were recorded at the reclaimed forest, reclaimed grasses, and bare rock sites, respectively. A large precipitation event occurred from 18 June to 21 June, with 195 mm of rain measured at the forest site, 112 mm measured at the grasses site, and 106 mm measured at the bare rock site (Figure 3.1.2). A precipitation event from 22 May to 23 May also contributed a large amount of precipitation with 62 mm of rain at the forest site, 82 mm at the grasses site, and 55 mm at the bare rock site. Overall, precipitation was measured on 73 days at the forest site, 75 days at the grasses site, and 71 days at the bare rock site over the course of the season. Precipitation measured at the nearby Sparwood weather station over the growing season showed May, June, August, and September to be wetter than normal, based on Environment Canada's 1981 – 2010 climate normals (Table 3.1.2). Overall, the 2013 growing season was much wetter than average, as 363.7 mm of precipitation was recorded at the nearby Sparwood weather station, compared to the 1981-2010 climate normal of 258.8 mm of precipitation. This large difference in precipitation was the result of a much wetter than normal June (42.1 mm more than normal) and September (51.8 mm more than normal).

3.2 Soil Moisture and Suction

Soil moisture measurements integrated to a depth of 75 cm were analyzed at each site. The start of the growing season at each site was characterized by variable soil moisture conditions, as precipitation events were frequent (Figure 3.2.1). The beginning of June was a drying down period for the soil at each site, as precipitation events were infrequent. The large precipitation event on 18 to 21 June greatly increased the soil moisture content at each site, as over 100 mm of rainfall was measured at each site. This event allowed soil moisture to reach maximums at each site, and was followed by another drying phase. Shortly spaced precipitation events caused a partial recharge of soil moisture from 5 to 8 August. Following this, soil moisture declined to seasonal lows at each of the sites. The soil water content at each site reached a minimum on 28 August. The minimum value at the grasses site was 0.04 $m^3 m^{-3}$, at the forest site was 0.09 $m^3 m^{-3}$, and at the bare rock site was 0.11 $m^3 m^{-3}$. The recharge of soil moisture during precipitation events, and the drying of soil following the events continued through to the end of the measurement period. Overall, soil moisture responded rapidly to precipitation events at each site, causing soil moisture to be quite variable throughout the growing season. Soil moisture content was the highest at the bare rock site throughout the season, except for a short period in mid-May. In contrast, soil moisture content was the lowest at the reclaimed grasses site throughout the season, with the exception of a short period at the beginning of May.

As a result of a sensor error at the reclaimed forest site, there were no measurements of soil suction taken at a depth of 20 cm for the entire growing season. At the reclaimed grasses site sensor error resulted in no measurements at a depth of 5 cm and 30 cm after August 11. To allow for consistency between sites, soil suction measurements at a depth of 10 cm are reported, which is close to the surface and would represent suctions within the rooting zone. As expected, periods of higher soil moisture were associated with lower soil suction values (Figure 3.2.2). Precipitation events caused an increase in soil moisture, resulting in a decrease in the magnitude of soil suction. Large or frequent precipitation events, such as the event in mid-June, resulted in recorded soil suction values of 0 kPa. Periods of minimal precipitation caused a drying of the soil and an increase in the soil suction. The influence of precipitation on soil suction was observed throughout the entire growing season and at all sites. Generally, soil suction values at the bare rock site were larger compared to the reclaimed vegetated sites despite higher volumetric water content, and experienced more variation throughout the season. Soil suction values at the grasses site experienced the least amount of variation throughout the season and were the smallest. At each site the highest soil suction values were seen at the end of August, after an extended dry period. The highest soil suction at the grasses site was -40 kPa, -150 kPa at the reclaimed forest site, and -460 kPa at the bare rock site.

3.3 Energy Balance

3.3.1 Net Radiation (R_n)

Net radiation exhibited the same seasonal trend at each of the sites, with an increase in the radiant flux until a peak in mid-July, and a decline thereafter (Figure 3.3.1). While the trend was the same at each of the sites the magnitude of the net radiation varied among them. Rn was greatest at the reclaimed forest site and was lowest at the bare rock site. For example, from 20-28 July, the forest site received 6632 MJ of radiation compared with 4890 MJ and 4414 MJ at the grasses and bare rock sites, respectively. Throughout the season, differences in the pattern of Rn were evident among the sites for certain periods. Over the period of 17-21 June, Rn at the forest site exhibited a different pattern than the grasses and bare rock sites, which was also observed from 21-31 August. While the grasses and bare rock sites are located only 500 m apart, the forest site is approximately 25 km north and at a lower elevation. Consequently, the forest site would be subject to a different localized climate than the grasses and bare rock sites, which may in part explain the periods when radiation patterns are different.

3.3.2 Latent Heat Flux (LE)

The seasonal pattern of the daytime (9:00 to 18:00) latent heat flux exhibited noticeable variation among the sites (Figure 3.3.2). Across the entire growing season, the value of the latent heat flux was greatest at the reclaimed forest site. At the beginning of the season the difference between values at the reclaimed forest and at the reclaimed grasses sites was larger than at the end of the season. Daytime LE values at the reclaimed forest site peaked earlier in the season compared to values at the reclaimed grasses site. Across the entire growing season the daytime latent heat flux from the bare rock site was noticeably lower than the vegetated sites. A large day-to-day variation in daytime LE was observed at the vegetated site, however at the bare rock site there was much less variation in LE. At the forest and grasses site, LE was the largest consumer of Rn, with seasonal average LE/Rn values of 0.58 and 0.44, respectively, calculated using 24 hour sums (Table 3.3.1). The forest site experienced the largest LE/Rn values at the beginning of the growing season, whereas the grasses site experienced the largest LE/Rn values at the end of the growing season, with a peak in September. The bare rock site was a comparatively small consumer of Rn, with a seasonal average LE/Rn value of 0.26. Similarly to the both the forest and grasses site, the largest LE/Rn value was observed for the month of May and August.

The average daily evapotranspiration rate during the 2013 growing season was 2.3 mm d^{-1} at the reclaimed forest, 2.1 mm d^{-1} at the reclaimed grasses sites, and 1.0 mm d^{-1} at the bare rock site (Table 3.3.2). Daily evapotranspiration rates were largest at the forest site until mid-August, after which the largest rates were measured at the grasses site (Figure 3.3.3). Average daily evapotranspiration rates at the vegetated sites were largest during the month of July, with an average 3.0 mm d^{-1} of ET at the forest site and an average of 2.8 mm d^{-1} at the grasses site. The peak daily evapotranspiration rate at the grasses site was measured on 10 July (4.8 mm d^{-1}), and was followed by the peak ET rate at the forest site on 11 July (6.5 mm d^{-1}). Mean daily evaporation rates were variable across the growing season at the vegetated sites, and were much less variable at the bare rock site. Daily evaporation rates ranged from 0.1 – 6.5 mm d^{-1} at the reclaimed forest site, from 0.01 – 4.8 mm d^{-1} at the reclaimed grasses site, and from 0.2 – 2.6 mm d^{-1} at the bare rock site.

3.3.3 Sensible Heat Flux (H)

There was a greater partitioning of the net radiation into the sensible heat flux at the bare rock site compared to the vegetated sites (Figure 3.3.4, Table 3.3.3). As the growing season progressed, sensible heat flux at the bare rock site increased beyond the values observed at the vegetated sites and remained higher through the end of the growing season. Peak sensible heat flux values were observed at each site at the end of July. The average daily daytime sensible heat flux ranged between -41 and 308 W m^{-2} at the reclaimed forest site, between -27 and 304 W m^{-2} at the reclaimed grasses site, and between -17 and 314 W m^{-2} at the bare rock site. At the bare rock site the largest portion of the daily net radiation was consumed as sensible heat. The seasonal average of H/Rn at the site was 0.63, with monthly averages ranging from 0.44 to 0.68, calculated using 24 hour sums. Comparatively, at the vegetated sites a smaller portion of the net radiation was partitioned into sensible heat. At the forest site, monthly averages ranged from 0.23 to 0.43, and at the grasses site monthly averages ranged from 0.19 to 0.33. At both vegetated sites the largest monthly average of the daily 24-hour sums of H/Rn occurred during the month of July, however the maximum at the bare rock site occurred during the month of August.

Overall, the daytime (9:00- 18:00) Bowen ratio (β) at the bare rock site was greater than the Bowen ratio at the vegetated sites. The average β for the entire growing season at the bare rock site was 3.50, compared with 1.00 and 0.97 at the forest and grasses sites, respectively (Table 3.3.4). Furthermore, the average β at the forest and grasses site varied around 1 (Figure 3.3.5), whereas the β values at the bare rock site were much larger (Figure 3.3.6). At all of the sites the largest average β occurred in July, corresponding with the peaks in the sensible heat flux at the sites. The highest β values were associated with periods of low precipitation and low soil water contents. After a precipitation event, as the soil would dry, each of the sites showed a general increase in the β. The occurrence of precipitation events caused a decrease in β at each of the sites, as LE increased. While β at the forest and grasses site generally exhibited similar patterns, there were periods where differences were noticeable. These differences occurred from 27 May – June 04, 24 -29 July, and from 25 – 30 August.

3.3.4 Ground Heat Flux (G)

The seasonal trend of the ground heat flux during the 2013 growing season can only be fully reported for the reclaimed grasses site, as no data was available at the bare rock site and limited data was available at the reclaimed forest site. Throughout the growing season G consumed the smallest amount of daily Rn compared to H and LE, as calculated using 24 hour sums. The seasonal trend exhibited a peak in average G values in mid-July (Figure 3.3.7). There was large fluctuation in G values from day to day through the first half of the growing season, with less variation seen from August through to the end of September. The ground heat flux at the grasses site was predominantly positive throughout the growing season, with negative values of the average daily daytime G occurring on at the very beginning and ending of the season, as well as on 20 June. Values of G at the grasses site ranged from -31 to 169 W m^{-2}. The limited data for G at the forest site shows values that are generally much smaller than those measured at the grasses site. G values at the forest site show much less day-to-day variation, with the exception of the relatively large negative values measured from 20 to 22 June. Without a complete data set the seasonal trend of G at the forest site cannot properly be examined and compared with the grasses site.

To confirm the G values at the reclaimed grasses and reclaimed forest, the measured values obtained from the heat flux plate were compared to calculated values. G values were calculated using measurements of the change in soil temperature, as well as soil property information. To determine the amount of energy partitioned into the ground heat flux, the total weekly daytime net radiation at each site was calculated, as well as the total weekly daytime soil heat flux. To account for only daytime hours net radiation values below 50 W m^{-2} were excluded from the calculations, along with the corresponding ground heat flux value. The percentage of soil heat flux to net radiation was calculated at each site, allowing for comparison with measured values. At the reclaimed grasses site the calculation approach determined that the daytime soil heat flux was approximately 25% of the net radiation throughout the growing season. Using the soil heat flux plate measurements it was calculated that the daytime ground heat flux accounted for 24% of the daytime net radiation. This result compares well to that of the calculation approach, and confirms that values obtained from the heat flux plate were acceptable and similar to values expected given a theoretical approach. At the reclaimed forest site the calculation method resulted in daytime ground heat flux measurements that were approximately 10% of the daytime net radiation. When values obtained from the heat flux plate were used to calculate the percentage of soil heat flux to net radiation it was determined that daytime G was approximately 6% of the daytime Rn. This result is similar to the percentage obtained using the theoretical calculation approach, and confirms that values measured by the heat flux plate are acceptable and similar to calculated values. Given the confirmation that calculated values and measured values of G were similar, the calculated G values were used at the reclaimed forest site for the purpose of estimating the energy balance closure, as large gaps were present in the measured data.

3.4 Diurnal Energy Partitioning

The mean diurnal variation of energy partitioning in May at each of the sites showed H to be slightly greater than LE at the grasses and bare rock sites throughout the day (Figure 3.4.1). At the forest site H was greater than LE until the middle of the day, after which LE became greater than H. The least amount of energy was partitioned into G at the forest site, with G measurements below that of H and LE during daytime hours (9:00 to 18:00). However, at the grasses site G exceeded LE between 9:30 and 11:30 and had values similar to H and LE throughout the day. At the bare rock site G exceeded H between 8:00 and 11:00 and exceeded LE between 7:00 and 14:00.

In June there was a general continuation of the patterns observed during May at the forest and grasses site, however there was a noticeable difference at the bare rock site. At the bare rock site, H greatly exceeded LE during the daytime hours, with very low values of LE measured at the site. From 8:00 to 14:00 G exceeded LE, however at no point did the average G exceed the average H during the day. At the grasses site very similar values of H and LE were seen throughout the day, with H slightly exceeding LE between 8:00 and 17:00. Throughout the day the least amount of energy was partitioned into G. At the forest site LE was seen to be greater than H at the beginning of the day (0:00 to 6:00) and from 12:00 through to the end of the day.

In July an even greater proportion of energy was recorded as sensible heat compared to latent heat at the bare rock site. Both H and LE peaked later in the day (15:00 and 16:00

respectively), whereas G peaked at 11:00. While H exceeded LE during the entirety of the daytime hours, G exceeded LE from 9:00 to 14:00. At the grasses site LE exceeded H throughout most of the day, except from 10:00 until 14:00 when Rn was at its peak. The least amount of energy at this site was partitioned into G throughout the day, although values of G similar to those of H and LE were seen from 10:00 to 12:00. At the forest site H exceeded LE from 7:00 to 14:00 with LE being greater the rest of the day. H peaked earlier in the day than LE, with H peaking at 12:00 and LE peaking at 14:00.

At the bare rock site in August most of the net radiation was partitioned into H during the majority of the daytime hours. From 8:00 to 11:00 more energy was partitioned into G than both of H and LE. After this peak in G there was a steady decline, and G remained the smallest energy flux after 14:00. H was seen to peak later in the day at this site compared to G and LE, which both peaked at 12:00. At the grasses site LE was greater than both H and G throughout the entire day. During daytime hours, H was greater than G. Both H and LE peaked at 13:00 (after the peak of net radiation) whereas G peaked at 11:00 (before the peak of net radiation). At the forest site H was greater than LE between 7:00 and 13:00 and both were greater than G for the entire day. While LE peaked at 13:00 along with net radiation, H peaked much earlier at 10:00, and G peaked later at 14:00.

At the bare rock site the month of September exhibited the same trends as August. During the daytime hours H was greater than LE, and G was greater than H from 8:00 to

10:00, and greater than LE from 8:00 to 12:00. H peaked at 15:00, while LE peaked at 14:00, and G peaked at 11:00. At the grasses site LE was once again greater than H and G across the entire day. The least amount of energy was proportioned into G throughout the day. Both H and LE peaked at 12:00 and peak values of G were seen at 11:00. At the forest site LE exceeded H except between 8:00 and 13:00. Values of G were less than values of LE and H during daytime hours. H experienced peak values at 10:00, peak LE values were seen at 12:00, and peak G values occurred at 14:00.

3.5 Water Balance

The cumulative patterns of precipitation (P) and evapotranspiration (ET) across the 2013 growing season show that P exceed ET across the entire study at the forest (Figure 3.5.1), grasses (Figure 3.5.2), and bare rock (Figure 3.5.3) sites. ET was approaching P during the middle of June, but the large precipitation event on 18-21 increased the cumulative P from 84 to 270 mm at the forest site, from 95 to 202 mm at the grasses site, and from 181 to 183 mm at the bare rock site. After this event cumulative ET was greatly exceeded by cumulative P, resulting in a precipitation surplus of 172 mm at the forest site, of 161 mm at the grasses site, and of 251 mm at the bare rock site at the end of the measurement period.

The cumulative soil moisture storage followed a similar pattern to the precipitation. When there was a precipitation event the soil moisture storage in the top 1 m of cover increased. Between precipitation events the occurrence of evapotranspiration resulted in a decrease in the amount of moisture in the soil. A decreasing trend of the soil moisture storage occurred until there was another precipitation event, and an influx of water into the soil. Over the growing season there was a net change in soil moisture of 1 mm at the forest site, of 12 mm at the grasses site, and of 4 mm at the bare rock site. Given that no runoff was observed at the sites, any water that was not evaporated or held in storage in the soil would have been lost of deep percolation.

The large precipitation event in June caused an increase in the amount of soil moisture by 36 mm at the forest site, by 39 mm at the grasses site, and by 30 mm at the bare rock site. At the forest site 195 mm of precipitation was recorded during this event, and 44 mm of evapotranspiration occurred prior to the start of the next precipitation event. The net change in soil moisture storage from the start of this event to the start of the next precipitation event was 2 mm, so 149 mm of water was unaccounted for in the top 75 cm and presumed lost to deep percolation. At the grasses site 112 mm of precipitation occurred during the event and 37 mm of this was evaporated prior to the next rain event. The net change in the soil moisture storage during this period was 4 mm, so 71 mm of water was unaccounted for and lost to deep percolation. At the bare rock site there was 106 mm of precipitation that fell during this event and 17 mm of evapotranspiration occurred. The net change in the soil moisture storage at this site was 4 mm, so 85 mm of water was unaccounted for and lost to deep percolation.

Over the measurement period (23 May to 30 September) at the reclaimed forest site there was 478 mm of precipitation, 305 mm of evapotranspiration, and a change in soil storage of 1 mm. Therefore, during the measurement period there was 172 mm of water that was lost to deep percolation. At the reclaimed grasses site there was 433 mm of precipitation, 272 mm of evapotranspiration, and a change in soil storage of 12 mm. This resulted in 148 mm of water being lost to deep percolation at this site during the measurement period. At the bare rock site 391 mm of precipitation reached the surface, of which 140 mm was evaporated and 4 mm was stored in the soil. Therefore, at the bare rock site 246 mm of water was lost to deep percolation. Overall, at the bare rock site, there was 143% more water lost to deep percolation when compared to the forest site, and 166% more lost when compared to the grasses site. At the bare rock site only 36% of the total precipitation was evaporated, whereas 64% and 63% of the precipitation was evaporated at the forest and grasses sites, respectively.

CHAPTER 4: DISCUSSION

4.1 Energy Balance and Evapotranspiration

Throughout the entire growing season at the forest and grasses sites more net radiation was partitioned into the latent heat flux than the sensible heat flux. At the forest site this difference in partitioning was greatest at the beginning of the measurement period, during the months of May and June. At the grasses site the greatest difference in partitioning was at the end of the measurement period, during August and September. Conversely, at the bare rock site a greater amount of energy was partitioned into the sensible heat flux, and the greatest disparity was observed in June and July. The absence of vegetation and rapid drainage of water to depth at the bare rock site limited the partitioning of net radiation into the latent heat flux and sensible heat dominated.

The partitioning of the net radiation at the reclaimed forest and grasses sites had distinct characteristics from the patterns observed at other boreal forest sites, which is the source of most literature available for comparison. Unlike results here, where latent heat dominates energy partitioning, at natural boreal forest sites more net radiation is typically partitioned into sensible heat (Baldocchi and Vogel, 1996; Baldocchi et al., 1997; Blanken et al., 2001). It has been observed that conifer forests partition between one third and one half of net radiation into evaporation (Baldocchi et al. (2000). Baldocchi and Vogel (1996) observed that over a boreal jack pine forest LE consumed only 36% of the available energy, and much more energy was used for the sensible heat flux. Arain et al. (2003) observed at a northern Saskatchewan black spruce forest that during the spring most of the available energy was partitioned into the sensible heat flux, with almost equal proportions of the latent and sensible heat fluxes in the summer months. While results from this study have a greater latent heat flux in the early season compared with the findings of Arain et al. (2003), similar proportions of the sensible and latent heat fluxes were observed at the reclaimed forest site for July and August. Lindroth (1985) observed that LE/Rn had a value of about 0.40 in May, increasing to 0.60 in August and September at a Scots pine forest in Sweden. At the reclaimed forest site, a greater value of LE/Rn was observed in May, however similar LE/Rn values were observed in August and September compared to Lindroth (1985). McCaughey et al., (1997) observed that the sensible heat flux was the dominant convective flux over the entire study period at a Manitoba jack pine forest. As well, at a jack pine forest examined by Baldocchi et al. (1997) H/Rn was greater than LE/Rn from the end of May until the beginning of July. At a black spruce forest in Saskatchewan a seasonal trend in LE/Rn was observed, with values increasing from May-June to July-August, and decreasing in September (Pattey et al., 1997; Jarvis et al., 1997). At a tamarack and black spruce forest examined by Lafleur (1992) in northern Manitoba LE /Rn was 0.48 and H/Rn was 0.44. At the beginning of the season H was dominant, however in the latter part of the study LE was greater. The values measured by Lafleur (1992) display more even partitioning of Rn into H and LE over the season than observed at the reclaimed forest site. The seasonal pattern of partitioning found by Lafleur (1992) agrees with other studies of the boreal forest, and further suggests that the reclaimed forest was partitioning energy differently than natural forests during the 2013 growing season. This is especially true in the spring, when a greater proportion of the net radiation was partitioned into the latent heat flux compared to observations at natural boreal forests

A grassland near Lethbridge, Alberta, examined during an above average precipitation year by Wever et al. (2002), exhibited an energy balance that was dominated by the latent heat flux, which was similar to what was observed at the reclaimed grass site. In addition, at grassland in Kansas that generally received ample rainfall examined by Kim and Verma (1990) most of the net radiation was partitioned into the latent heat throughout the season, with the sensible heat flux only being dominant during a dry spell and during plant senescence. A mountain grassland in Austria examined by Hammerle et al. (2008) exhibited dominant partitioning of the net radiation into the latent heat flux over the vegetated period of the year, with an average LE/Rn value of 0.54. Overall, results obtained from the reclaimed grasses site agree with the results of studies completed on natural grasslands, and suggests that the reclaimed has a similar energy balance regime to natural sites.

While evapotranspiration rates were variable across the growing season and between the sites, the seasonal average ET rate at the vegetated sites were similar, with the forest site only recording 33 mm more ET than the grasses site over the period of observation. To further evaluate the performance of the reclaimed sites, ET rates were compared with natural boreal sites (Table 4.1.1). ET rates at the reclaimed forest site exhibited slightly more variability than observed at a natural jack pine forest in Saskatchewan, where daily

ET rates ranged from 0.5 to 2.5 mm d^{-1} (Baldocchi et al., 1997). At a Churchill, Manitoba, site dominated by black spruce the average daily ET rate was 2.24 and 2.33 mm d^{-1} in 1989 and 1990, respectively (Lafleur 1992). While air temperatures at the Churchill site are colder than in the Elk Valley, the climate normal June to August precipitation of 52 mm m^{-1} is similar to that of the Sparwood weather station. The year 1989 was considered to be a dry year, and 1990 was considered to be a wet year at the Churchill site (Lafleur 1992). The seasonal average ET rate at the reclaimed forest site agrees well with ET rate measured at the Churchill site during the wet year of 1990. At a black spruce forest in Saskatchewan, Canada, Arain et al. (2003) observed a maximum daily evaporation rate during the summer of 3.5 mm d^{-1}. Ewers et al. (2005) observed peak transpiration per unit leaf area of 3.5 mm d^{-1} at a jack pine forest, 2.5 mm d^{-1} at an aspen forest, and 1.3 mm d^{-1} at a black spruce forest in Manitoba. A black spruce forest in Saskatchewan examined by Jarvis et al. (1997) had an average evaporation rate of 2 mm d^{-1}. Moore et al. (2000) observed an average daily ET rate 1.1 mm d^{-1} at a jack pine forest in Manitoba. Average daily ET rates at a black spruce forest in Manitoba was 1.3 mm d^{-1}, while in Saskatchewan a jack pine forest had an ET rate of 1.6 mm d^{-1} and a black spruce forest had an ET rate of 1.9 mm d^{-1} (Moore et al., 2000). The average daily ET rate at the reclaimed forest site over the 2013 growing season was greater than at each of these natural sites. While the reclaimed forest contains both pine and spruce trees, the LAI is much greater than at these comparable natural studies. As well, these natural sites in Manitoba and Saskatchewan are located within a continental climate regime, and would receive less precipitation than the Elk Valley. These physiological and climatological differences would result in different ET rates compared to the reclaimed Elk Valley sites. A coastal douglas-fir site in British Columbia examined by Humphreys et al. (2003) had an average evaporation rate of 2.7 mm d^{-1} and maximum summer evaporation rates of 3.7 mm d^{-1} and 3.4 mm d^{-1}, in 1998 and 1999, respectively. Ponton et al. (2006) observed a maximum evapotranspiration rate of 2.7 mm d^{-1} at a Vancouver Island douglas-fir forest and 3.2 mm d^{-1} at a Saskatchewan aspen forest. The douglas-fir sites are characterized by dry and cool summers, making them climatologically different than the Elk Valley sites. Therefore, differences in ET rates may partly be related to different climate between the sites. At the reclaimed forest site the maximum evapotranspiration rate during the 2013 growing season exceeded the maximum rates observed at other boreal forest sites. The maximum rate observed was 6.5 mm d^{-1} on 11 July and on 14 days the average evapotranspiration rate exceeded 3.5 mm d^{-1}, the maximum rate reported at several natural sites. The seasonal average evapotranspiration rate at the reclaimed forest site, 2.3 mm d^{-1}, however, agrees well with other seasonal average evapotranspiration rates. While one season of data is not definitive, the evapotranspiration rates observed at the reclaimed forest site are similar to rates at other boreal forest sites in Canada.

A study completed by Wever et al. (2002) at a grassland site near Lethbridge, Canada, resulted in a peak evapotranspiration rate of 4.5 mm d^{-1} observed during a year where more precipitation occurred than normal. Over years where there was precipitation near or less than the average, the peak measured ET rate was 3 mm d^{-1} (Wever et al., 2002).

At a Southern Alberta grassland site studied by Ponton et al. (2006) the maximum evapotranspiration rate was 5 mm d^{-1} during a lower than average precipitation year. These results are similar to the peak evapotranspiration rate observed at the reclaimed grasses site (4.8 mm d^{-1}) during the 2013 growing season. At a grassland site in Japan examined by Li et al. (2005) the daily mean evapotranspiration rate ranged from 0.3 to 4.2 mm d^{-1}, however this site was characterized by much greater amounts of rainfall than measured in the Elk Valley. Overall, the evapotranspiration rates from the reclaimed grasses site are similar to rates observed at natural grassland sites. The similarity of evaporation rates between the reclaimed forest and reclaimed grasses site agrees with the finding of Kelliher et al. (1993) that under well watered conditions maximum evapotranspiration rates at a forest and grassland site are similar.

While absolute ET rates vary compared with sites in the literature, as would be expected, similar seasonal patterns were observed. At a jack pine forest in central Saskatchewan, evapotranspiration rates peaked during the months of June and July (Baldocchi et al., 1997). Arain et al. (2003) observed maximum evaporation rates in late June and July at a black spruce forest, agreeing with an earlier study that observed maximum evaporation at a black spruce forest in July (Jarvis et al., 1997). At jack pine and black spruce forests in Saskatchewan, as well as a grassland site near Lethbridge, Alberta, Zha et al. (2010) observed peak evaporation in July. At the reclaimed forest site examined in this study, peak evapotranspiration rates were also observed during June and July. While evapotranspiration at the forest site peaked in June and July, the grasses and bare rock sites experienced a peak later in the season, during July and August. Lower evapotranspiration rates during the spring, as well as during September, were observed at each of the study sites. This result agrees with the findings from other studies, which linked the presence of frost to the reduction in evapotranspiration rates during these periods (Baldocchi and Vogel, 1996; Betts et al., 1999). Other factors, such as the lower air temperatures and net radiation values observed at each of the sites in the spring and fall, would also result in lower ET rates in the spring and fall (Baldocchi et al., 2000). As well, lower ET rates in the spring and fall have been linked to the emergence and senescence of leaves and the seasonal pattern of LAI (Wever et al., 2002). These factors would affect the lower ET rates in the spring and fall at the reclaimed grasses site. The seasonal pattern of evapotranspiration at the reclaimed forest and reclaimed grasses sites agree with patterns observed at natural forest and grassland sites.

Over the course of the day, peak evapotranspiration was observed at different times at the sites. At the reclaimed forest and bare rock sites, the peak in evapotranspiration was observed during the mid-afternoon, while at the reclaimed grasses site the peak occurred during the middle of the day. This agrees with a study done by Kelliher et al. (1993) that found that peak evapotranspiration from forests generally occurs during mid to late afternoon and the peak rate at a grassland occurs at mid day when net radiation is at a maximum. Arain et al. (2003) observed that at a black spruce forest a peak in the latent heat flux was observed in the early afternoon during the spring, and around noon during summer months. However, at the reclaimed forest site peak evapotranspiration rates occurred in the early afternoon during both spring and summer months.

4.2 Energy Balance Closure

The closure of the energy balance at each of the Elk Valley sites was slightly poorer than other boreal eddy covariance studies, however estimated closure at the reclaimed forest site did fall within the range common to other studies, as stated by Twine et al. (2000). Twine et al. (2000) reported that a residual error between 10 and 30% is common in eddy covariance studies, while Foken et al. (2006) stated that most land experiments achieve an energy balance closure of 80%. Amiro et al. (2006) achieved energy balance closure fractions of 0.89 at a black spruce forest and 0.85 at a jack pine forest, while Barr et al. (2006) determined that surface available energy was underestimated by 15% in black spruce and 14% in jack pine forests. Arain et al. (2003) achieved energy balance closure fractions of 0.86 and 0.81 during 1999 to 2000 and 2000 to 2001 at a black spruce forest. The estimated closure of the energy balance at the reclaimed forest site, assuming G was 10% of Rn, was 73% ($r^2 = 0.71$) (Figure 4.1.1). Closure at the reclaimed grasses site was 55% ($r^2 = 0.87$) (Figure 4.1.2), using all measured values. The estimated energy balance closure at the bare rock site, assuming G was 30% of Rn, was 66% ($r^2 = 0.64$) (Figure 4.1.3). During daytime (9:00 – 18:00) hours closure increased to 77% ($r^2 = 0.57$) at the forest site and 57% ($r^2 = 0.81$) at the grasses site, while closure decreased to 52% ($r^2 = 0.39$) at the bare rock site during the daytime. This suggests that nighttime errors had a small negative impact on the overall energy balance closure at the vegetated sites. Barr et al. (2006) also observed a decrease in closure during the night, as closure decreased from between 0.85 and 0.90 to 0.65 to 0.79.

While the underestimation of turbulent fluxes (H, LE) by the eddy covariance method is often the main source of error preventing full closure, at the Elk Valley sites much of the error is attributed to the lack of ground heat flux data. The reclaimed grasses site was the only Elk Valley study site that had a full set of ground heat flux plates working throughout the summer. At the reclaimed forest site there were large gaps in the ground heat flux data, so an estimation of the energy balance was calculated using the assumption that G was 10% of Rn. This assumption creates error in the available energy calculations at this site, and increases the overall error in the energy balance closure calculation. At the bare rock site no ground heat flux data was available for the 2013 growing season and G was estimated as 30% of the net radiation. This assumption resulted in error in the available energy calculations at this site, and would limit the ability to close the estimated energy balance. While 30% was determined to be a proper estimation of the percentage of the net radiation that went into the ground heat flux, it is possible that given surface properties this is an underestimation. While the lack of G data is suspected to have contributed the greatest error, other factors may also have contributed. Anthoni et al. (1999) observed that large gaps in the canopy resulted in spatial variation in the soil heat flux and upwelling radiation. The reclaimed forest site did have sporadic gaps in the canopy, which may have negatively impacted the energy balance closure. Other potential sources of error include insufficient fetch, missing of the covariance of low frequency fluctuations, or sensor errors due to sensor separation, frequency response, or alignment problems (Twine et al., 2000).

4.3 Controls on Evapotranspiration

To determine controls on evapotranspiration at each site, the relationship between the latent heat flux and air temperature (Figure 4.3.1), vapour pressure deficit (Figure 4.3.2), net radiation (Figure 4.3.3), wind speed (Figure 4.3.4), surface resistance (Figure 4.3.5), and aerodynamic resistance (Figure 4.3.6) were examined. There was a distinct difference between the controls on evapotranspiration at the vegetated sites and the bare rock site. At the vegetated sites, ET was more strongly related to meteorological variables: net radiation, air temperature, and the vapour pressure deficit, whereas surface resistance as calculated from the Penman-Monteith method controlled evapotranspiration at the bare rock site.

The greatest control on evapotranspiration at the reclaimed forest site was net radiation. Net radiation exhibited a strong positive relationship with the latent heat flux during daytime hours. While many studies have shown a relationship between net radiation and evapotranspiration at forest sites, net radiation is often not the strongest control. Betts et al. (2001) observed that evapotranspiration decreased at high values of net radiation at a boreal forest in Manitoba, a result not observed at the reclaimed forest site. Baldocchi and Vogel (1996) found that net radiation was not the main control on evapotranspiration at jack pine forest in Saskatchewan, and that the influence of the vapour pressure deficit was significant. They observed that an increase in the vapour pressure deficit resulted in a decrease in evapotranspiration, a relationship also observed by Humphreys et al., (2003) at a coastal British Columbia Douglas-fir forest. At the reclaimed forest site, there was positive a relationship between the latent heat flux and the vapour pressure deficit, which was inconsistent with the results of other studies. Bernier et al. (2006) observed that at a jack pine forest in Saskatchewan radiation was responsible for 30% of the variation in transpiration, with the vapour pressure deficit and relative water content responsible for 13% and 10% of the variation, respectively. This agrees with the findings at the reclaimed forest site, where net radiation and the vapour pressure deficit exhibited a control on evapotranspiration. However, at the reclaimed forest site, soil moisture did not exhibit control on ET (Figure 4.3.7). The reclaimed forest site is made up of coarse grained soils that did not allow for much variation in soil moisture content throughout the measurement period. As a result, soil moisture was not a limiting factor on ET, as ample water was available to be evaporated throughout the season. It is therefore suggested that the soil texture was a reason why the soil moisture did not exhibit control on ET. Betts et al. (2001) observed that at a jack pine site, stomatal control played an important role in the transpiration of conifers. Baldocchi et al. (2000) found that canopy resistance in boreal conifers tends to be inherently large and observed a relationship between stomatal resistance and the vapour pressure deficit. Humphreys et al. (2003) observed a relationship between stomatal conductance and the vapour pressure deficit, a result that further suggests that a relationship between the vapour pressure deficit and surface resistance may be expected at the reclaimed forest site. At the reclaimed forest site there was no relationship observed between surface resistance and the latent heat flux in contrast to the results of studies at natural conifer sites. In addition, there was no relationship observed between the vapour pressure deficit and the surface resistance. This may be the result of sufficient soil moisture to allow for evaporation during periods of greater vapour pressure deficits, reducing the need for stomatal closure at high pressure deficits. Surface resistance can also be influenced by soil moisture, however at the reclaimed forest site there was no relationship found between these two variables. While there was not a strong relationship observed between aerodynamic resistance and the latent heat flux, there was less aerodynamic resistance at this site compared to the reclaimed grasses and bare rock sites, which is expected as rougher vegetation generates a greater amount of turbulence and less aerodynamic resistance to the transfer of water vapour (Baldocchi et al., 2000).

Similar to the reclaimed forest site, the greatest control on evapotranspiration at the reclaimed grasses site was net radiation. There was a strong positive relationship between the net radiation and the latent heat flux. This result agrees with Calder (1998), who determined that evapotranspiration rates from a moorland in the United Kingdom are closely related to the supply of radiant energy, and are radiation limited. Li et al. (2005) also observed that at a grassland, in Japan variations in the latent heat flux were primarily dependent on changes in the available energy. While net radiation exhibited the strongest control on the latent heat flux at the reclaimed grasses site, air temperature (R^2 = 0.57) and vapour pressure deficit (R^2 = 0.51) were also observed to exert a strong control on evapotranspiration. There was also a weakly negative relationship between the latent heat flux and the soil moisture observed at this site (Figure 4.3.7.). This relationship differed from the one observed at the reclaimed forest site and was an unexpected result, as a greater availability of water was expected to allow for increased evapotranspiration. When this relationship was examined on a monthly basis negative relationships were observed during May, June, and September, whereas positive relationships were observed during July and August. The largest average daily daytime (9:00 to 18:00) net radiation occurred during July and August. This suggests that during months where negative relationships were observed between LE and soil moisture less energy was available for ET. As well, the large June rainfall event occurred during a period with low net radiation, and lower ET potential. As a result, the larger soil moisture content of the soil occurred during a period of low ET, driving this negative relationship. Zha et al. (2010) observed that summer soil moisture was a dominant climatic control on evapotranspiration at a grassland site near Lethbridge, Alberta, however at the reclaimed grasses site soil moisture only exhibited a weak control. The lesser control on soil moisture on ET compared to the natural grassland site may be the result of different soil textures. The natural grassland was underlain by loamy soils that experienced high variability in soil moisture (Zha et al., 2010), while the reclaimed grasses site is underlain by coarse soils that experienced less variation in soil moisture. A positive relationship between water content and evapotranspiration was observed in July at the reclaimed grasses site, agreeing with Zha et al., who observed the same relationship. Wever et al. (2002) observed that evapotranspiration from a grassland near Lethbridge, Alberta, was strongly controlled by stomatal conductance, however no relationship was observed at the reclaimed grasses site. As well, there was no relationship observed between the surface conductance and precipitation, as suggested by Wever et al. (2002). Compared to the reclaimed forest site there was greater aerodynamic resistance observed, however there was no relationship between the latent heat flux and aerodynamic resistance at the reclaimed grasses site. The reclaimed grasses site was the only study site where there was a relationship between wind speed and the latent heat flux, however the relationship was not strong compared with other factors.

The largest control on evapotranspiration at the bare rock site out of the variables tested was the surface resistance. This negative relationship also represented the strongest control that surface resistance had on evapotranspiration at any of the study sites. The average daily daytime surface resistance at this site was greater than at both of the vegetated sites, with an average value of 205 s m^{-1}. This far exceeded the average values observed at the reclaimed forest and grasses sites, 29 s m^{-1} and 28 s m^{-1}, respectively. The greatest control on the surface resistance at the bare rock site was relative humidity (R^2 = 0.23), with a negative relationship observed. Surface resistance exhibited positive relationships with vapour pressure deficit, as well as air temperature. Given the larger volumetric water content and moisture availability in the near-surface zone of the bare rock site, it was expected that there would be a relationship between soil moisture and the latent heat flux. There was, however, no relationship found between the latent heat flux and soil moisture at this site (Figure 4.3.7). While there was no relationship between LE and soil moisture over a seasonal time frame, there was a noticeable difference in ET rates after rainfall. On days where rainfall occurred the average daily ET rate was 1.4 mm d^{-1}, however the day after a precipitation event the average daily ET rate increased to 1.6 mm d^{-1}. This relationship was most prominent during the month of July when the average ET rate on days where precipitation occurred was 1.07 mm d^{-1}, and the day after a precipitation event the average ET rate increased to 1.8 mm d^{-1}. The month of July experienced the largest average daily radiation at this site, and therefore had the most energy available for ET. Combined with increased available moisture this allowed for July to experience the largest ET rates after rainfall at the bare rock site. It can therefore be concluded that soil moisture does exhibit control on ET, however this occurs during periods after rainfall. Given that the bare rock surface can be considered more aerodynamically smooth than the forest or grasses sites, it was expected that higher values of aerodynamic resistance would be observed. The largest amount of aerodynamic resistance did occur at this site, however there was no relationship between the aerodynamic resistance and the latent heat flux. While surface resistance appeared to exhibit the strongest control on the latent heat flux, given the type of surface this likely is not true. Rather, the dominant control on the latent heat flux would be the near surface soil moisture content.

4.4 Water Balance

At each of the study sites the net change in soil storage over the growing season was negligible, meaning that most of the water at the surface was lost to either evapotranspiration or to deep percolation within the spoil piles. At the reclaimed forest site the soil moisture storage only increased by 1 mm over the growing season. There was an increase of only 12 mm of soil moisture storage at the reclaimed grasses site, and at the bare rock site the soil moisture storage only increased by 4 mm over the 2013 growing season. At the bare rock site 36% of the precipitation was evaporated, meaning that 63% of the precipitation that reached the surface was unaccounted for and lost to deep percolation. At the reclaimed grasses site 63% of the precipitation was evaporated, which resulted in the loss of 34% of the precipitation to deep percolation. At the reclaimed forest site 36% of the precipitation was lost to deep percolation, as 64% was lost to evapotranspiration.

The water balance at natural boreal sites generally resulted in a large proportion of the precipitation being returned to the atmosphere as evaporation, and in some studies evaporation exceeded precipitation. Arain et al. (2003) observed that 85% of the precipitation was evaporated over a black spruce forest in Saskatchewan, 26% greater than at the reclaimed forest site. Barr et al. (2007) observed that over most growing seasons at a trembling aspen forest in Saskatchewan evaporation exceeded precipitation, resulting in a net loss of water. The average amount of yearly precipitation that occurred was 422±103 mm, while the yearly average amount of evaporation was 418±64 mm, resulting in an average of 99% evaporation of precipitation. At a douglas-fir forest examined by Humphreys et al. (2003) evapotranspiration exceeded precipitation in both the summer of 1998 and the summer of 1999, by 102 mm and 29 mm respectively. At sites examined by Amiro et al. (2006) a jack pine forest evaporated 86% and 59% of the total precipitation in 2001 and 2002, respectively. The black spruce forest studied evaporated 94% and 76% during the same years. The reclaimed forest site, however, evaporated a smaller percentage of the total precipitation compared to these natural sites. The boreal region is characterized by low precipitation, with the Canadian boreal region generally receiving between 200 and 600 mm of precipitation yearly (Baldocchi et al., 2000). However, the Elk Valley study sites received between 391 mm and 478 mm of precipitation over just the 5 month growing season. Therefore, while it was observed that ET rates over the reclaimed forest site are similar to rates at natural boreal sites, the larger amounts of precipitation in the Elk Valley resulted in a smaller proportion of the precipitation being evaporated. This difference was also further enhanced by the large precipitation event in June 2013, as it greatly increased the difference between cumulate P and cumulative ET. Zha et al. (2010) observed that the largest difference between precipitation and evapotranspiration occurred at a jack pine forest, followed by a black spruce forest. At the jack pine forest the mean annual difference was 177 mm, and at the black spruce forest the mean annual difference was 123 mm (Zha et al., 2010). The reclaimed forest site in the Elk Valley is made up of spruce and pine trees and the difference between precipitation and evapotranspiration was 198 mm for the 2013 growing season. The Saskatchewan sites studied by Zha et al. (2010) received less precipitation than the Elk Valley study sites, and therefore a smaller P-ET value was expected for these locations compared to the reclaimed forest site. The mean annual growing season ET at the natural jack pine site was 300 ± 20 mm, very similar to the amount of ET that occurred at the reclaimed forest site. The disparity between the P-ET values at the natural jack pine site compared to the reclaimed forest site is therefore largely due to more precipitation in the Elk Valley compared to Saskatchewan.

Zha et al. (2010) observed that the least annual difference between precipitation and evapotranspiration occurred at a grassland site, with values fluctuating around 0 mm and a mean annual difference of -3 mm. This finding is quite different than what was observed at the reclaimed grasses site, where there was a 157 mm difference between precipitation and evapotranspiration. Wever et al. (2002) also observed that almost all of the precipitation that fell during the study was evaporated, and that there was similarity between cumulative precipitation and cumulative evapotranspiration. The Elk Valley, however, receives a greater amount of precipitation than Lethbridge, the location of the Zha et al. (2010) and Wever et al. (2002) studies. It is therefore suggested that the differences in the water balance between the natural grassland and the reclaimed grasses site is largely a result of this climatic difference. It is also important to note that a large precipitation event occurred in June 2013, and was anomalous to the region. A precipitation event of similar magnitude did not occur during the study periods of Zha et al. (2010) and Wever et al. (2002). This large precipitation event caused a large influx of water into the ground over a short time period, resulting in a large amount of deep percolation. This event was a main reason for the large difference between P and ET at the reclaimed grasses site. Overall, the different amount of precipitation received at the reclaimed forest site compared to natural boreal grasslands makes it difficult to compare water balances.

The results of this study indicate a clear distinction between the bare rock surface and the reclaimed vegetated surfaces. The placement of a vegetated cover atop the waste rock causes a noticeable decrease in the amount of net percolation that occurs. The growing of a forest on the waste rock resulted in a 30% decrease in the amount of net percolation that occurred and the presence of a grass surface resulted in a 40% decrease in the amount of percolation that occurred when compared to the bare rock surface. Overall, the reclaimed forest evaporated 64% (305 mm) of the precipitation and the reclaimed grasses site evaporated 63% (272 mm) of the precipitation. Compared to the bare rock site, where 36% (140 mm) of the precipitation was evaporated, it is clear that there is a large difference in the amount of water available for deep percolation between the vegetated and non-vegetated sites. With a goal of decreasing the amount of water lost to deep percolation the placement of a vegetated cover atop the waste rock has a positive impact, with the grass cover allowing for the greatest decrease in net percolation over the 2013 growing season. The reclaimed forest site experienced the largest amount of ET over the season and evaporated a greater portion of the precipitation, however due to the increased amount of precipitation at this site compared to the reclaimed grasses site there was still more water available for net percolation.

The results of this study represent a high precipitation year, so results were scaled to provide insight into what might occur over a normal year. At the Sparwood weather station there was 29% more precipitation measured during the 2013 growing season compared to the 1981-2010 climate normal. At the bare rock site during a normal year, assuming a 29% decrease in precipitation from 2013 and the same weighting of water balance components, it would be expected that 175 mm would have been lost to deep percolation. Comparatively, 122 mm and 105 mm of precipitation would have been expected to be lost to deep percolation at the reclaimed forest and reclaimed grasses sites, respectively.

Given these results the reclaimed grasses surface resulted in the least amount of deep percolation. However, the largest amount of evapotranspiration occurred at the reclaimed forest site, resulting in a slightly greater percentage of the precipitation being returned to the atmosphere at this site compared to the reclaimed grasses site. The potential for the forest cover to reduce net percolation may increase over time and surpass the abilities of the grass cover. The grasses located at the reclaimed grasses site have likely matured and reached their maximum evapotranspiration capabilities. However, the reclaimed forest may not have yet reached maturation and has the potential to evaporate greater volumes of water over time compared to the grasses site.

CHAPTER 5: CONCLUSION

In order to evaluate different management strategies, and determine the most effective method to reduce selenium loading to the surrounding ecosystem, the potential influence of vegetation on evapotranspiration from coal spoils needed to be examined. This aspect of the research program, established by Teck Coal Ltd., aimed to examine the potential benefits of placing vegetation on coal spoils as a method to limit deep percolation through the waste rock, as a mechanism to reduce contaminant loading. The comparison method used by this study allowed for the examination of two different vegetated covers, as well as the bare waste rock surface, to determine the best method of reducing percolation through an increase in evapotranspiration. While many studies have examined evapotranspiration rates and controls from forests and grasslands, the implementation of this study on Teck Coal Ltd. mine sites allows for more accurate insight into beneficial reclamation methods. The results from this study can aid in the implementation of effective management and reclamation strategies across all Teck Coal Ltd. mine sites.

The results of this study indicate many differences between sites with a re-vegetated forest cover, a re-vegetated grasses cover, and a bare waste rock surface. Soil moisture was generally greatest at the bare rock site and lowest at the reclaimed grasses site throughout the 2013 growing season. At each of the sites soil moisture responded rapidly to precipitation events, recharging during the events and drying following the events. This resulted in soil moisture being quite variable throughout the season at each of the sites.

The partitioning of net radiation into the latent heat flux was greatest at the reclaimed forest site, and least at the bare rock site. The seasonal average LE/Rn at the reclaimed forest site was 0.58, whereas the seasonal averages at the reclaimed grasses and bare rock sites were 0.44 and 0.26, respectively. LE/Rn was largest during May at the reclaimed forest site, during September at the reclaimed grasses site, and during August at the bare rock site. The large partitioning of energy into the latent heat flux resulted in seasonal ET rates of 2.3 mm d^{-1} and 2.1 mm d^{-1} at the reclaimed forest and reclaimed grasses sites, respectively. The bare rock site experienced the lowest seasonal ET rate, 1.2 mm d^{-1}. ET rates were highest during the month of July at the both the reclaimed forest and reclaimed grasses sites, while ET rates at the bare rock site were highest during the month of August. At both of the vegetated sites ET was dominantly controlled by net radiation, with lesser controls being air temperature and the vapour pressure deficit. At the bare rock site ET was dominantly controlled by the near surface soil moisture.

The seasonal average partitioning of the net radiation into the sensible heat flux was greatest at the bare rock site (0.63), followed by the reclaimed forest site (0.34), and the reclaimed grasses site (0.29). The seasonal average mean daily daytime Bowen ratio at the bare rock site was thus dominated by the sensible heat flux (3.50). The seasonal average daily daytime Bowen ratios at the reclaimed forest and reclaimed grasses sites were 1.00 and 0.97, respectively, and showed a near-even ratio of sensible to latent heat fluxes.

The presence of a vegetated cover atop the spoil pile resulted in an increased amount of evapotranspiration and a decreased amount of net percolation compared to the bare rock site. At the bare rock site 140 mm of ET was measured over the 2013 growing season, returning only 36% of the precipitation to the atmosphere. As a result, 246 mm of precipitation was lost to deep percolation. At the reclaimed grasses site 272 mm of ET occurred, accounting for 63% of the precipitation. The increased ET resulted in a lesser 148 mm of precipitation being lost to deep percolation. At the reclaimed forest site 305 mm of ET was measured, meaning 64% of the precipitation at this site was evaporated and 172 mm was lost to deep percolation. Overall, the forest cover resulted in the largest amount of ET over the growing season, and also resulted in the evapotranspiration of the greatest percentage of precipitation. The least amount of deep percolation occurred at the reclaimed grasses site, however there was also less precipitation at this site.

Results from this study represent a growing season that experienced above normal precipitation, during which a large anomalous rainfall event occurred. Given the greater amount of precipitation more water was available for ET, soil storage, and deep percolation compared with a normal year. It therefore is expected that during a growing season with normal precipitation, and without such a large rainfall event, less water would be subject to deep percolation at each of the sites. While the amount of ET, soil storage, and deep percolation may change, it is expected that ET will always be greater at the vegetated sites, and that the largest amount of deep percolation will occur at the bare rock site. In order to obtain a more accurate estimate of the amount of ET and percolation that occurs during a normal growing season at each of the sites this study needs to be continued on a long-term basis.

This study suggests that in order to maximize the amount of ET a forest cover should be placed on top of the waste rock. Given equal amounts of precipitation at each of the sites a forest cover would also result in the least amount of deep percolation of water and constituents. However, it is recommended that future reclamation projects use both grasses and forest to cover the waste rock at Teck Coal Ltd. sites. The use of both forest and grasses is a practical reclamation technique that would allow for the reduction of infiltration and percolation through the waste rock piles.

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Appendix A

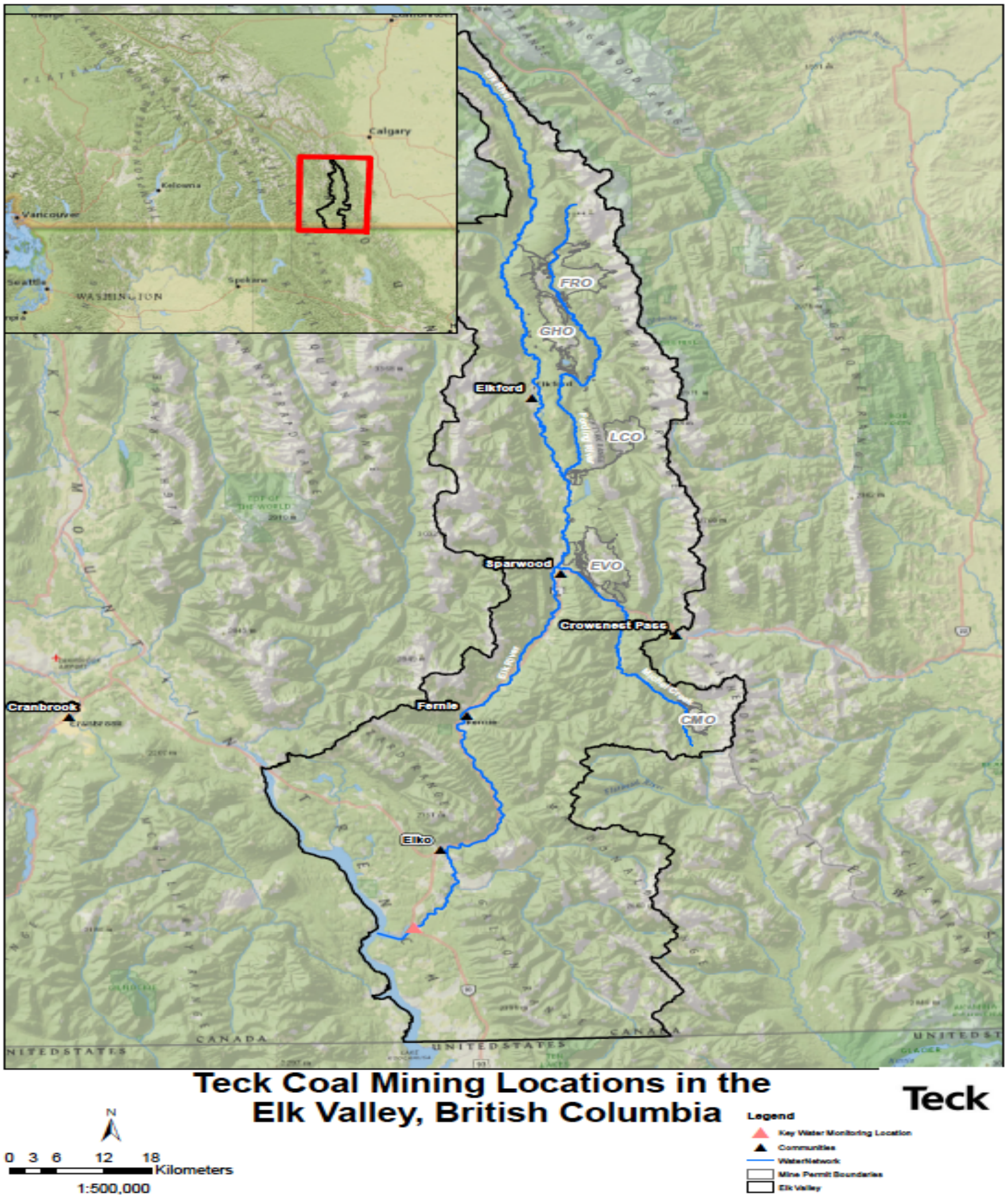

Figure 2.1.1: Map of the Elk Valley, British Columbia, including Teck Coal Ltd. operated mines.

Figure 2.1.2: Photograph of tower, showing eddy covariance system and forested vegetation. The tower is set in a small clearing.

Figure 2.1.3: Photograph of tower and reclaimed grasses site.

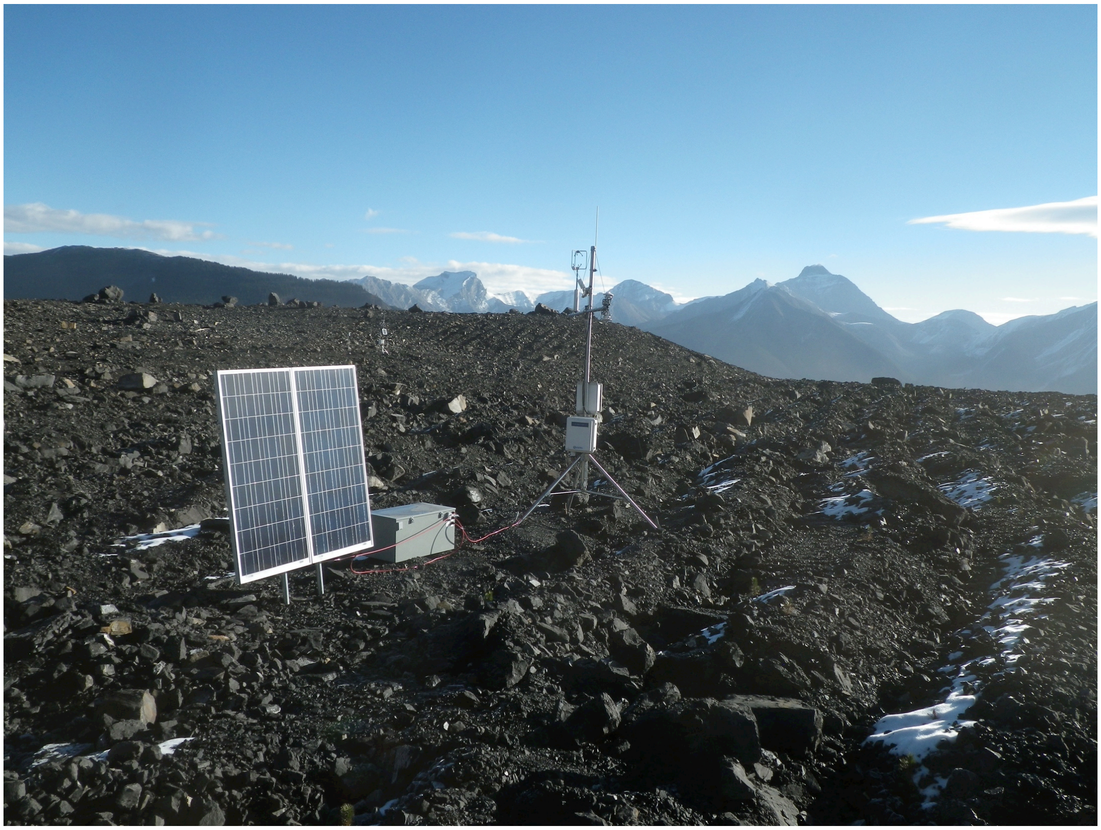

Figure 2.1.4: Photograph of site, showing the eddy covariance tower and the bare spoil surface.

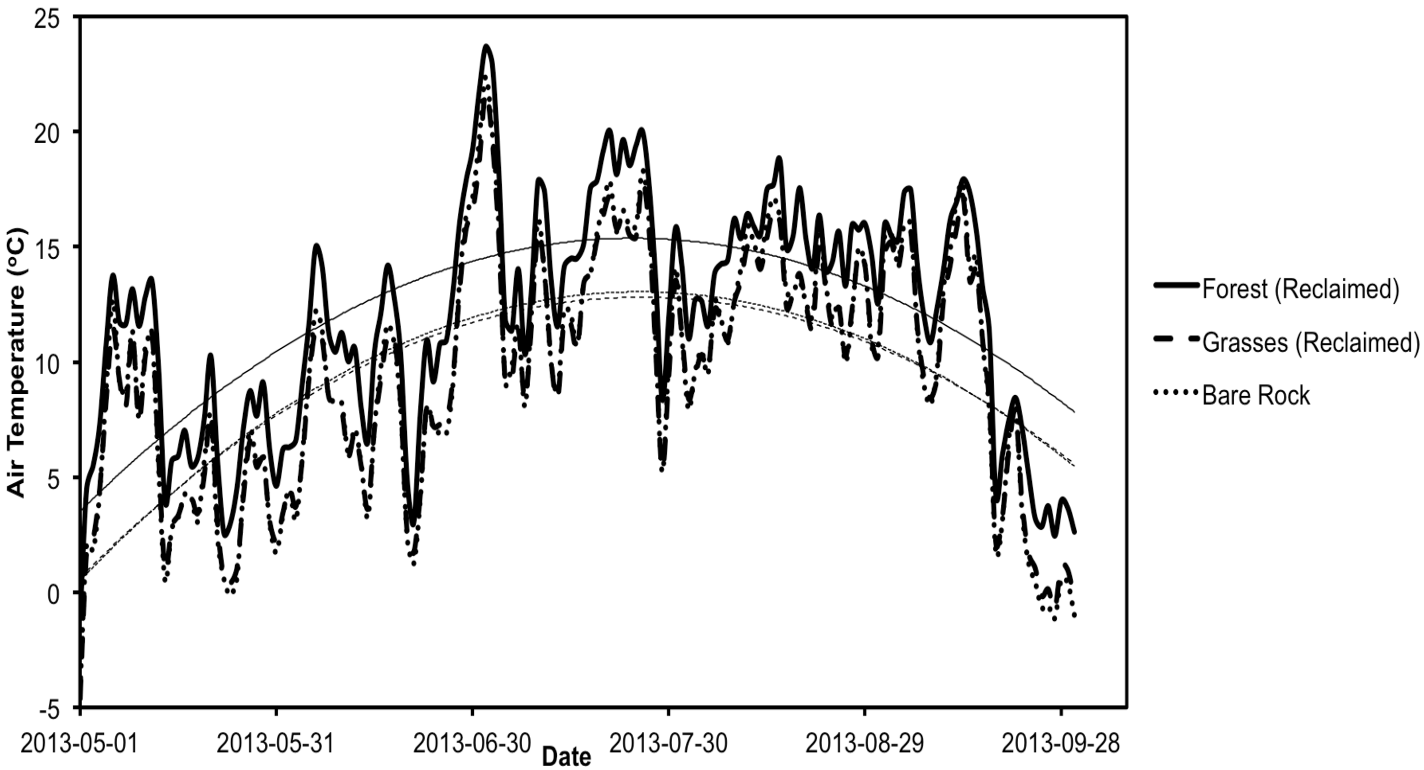

Figure 3.1.1: The average daily air temperature at each study site over the 2013 growing season. The seasonal trend at each site is depicted with a second-order polynomial.

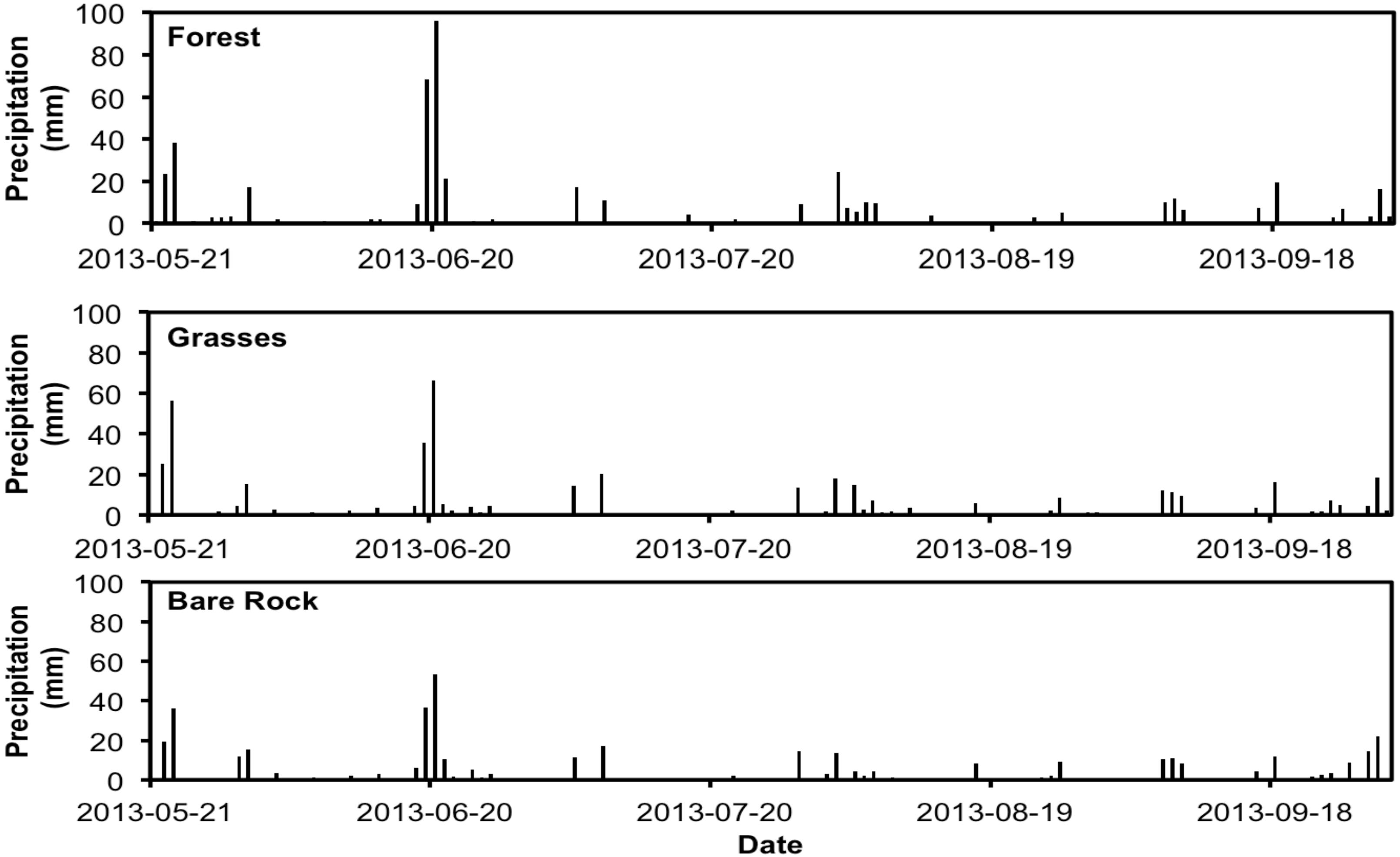

Figure 3.1.2: Total daily precipitation at each of the sites for the 2013 growing season.

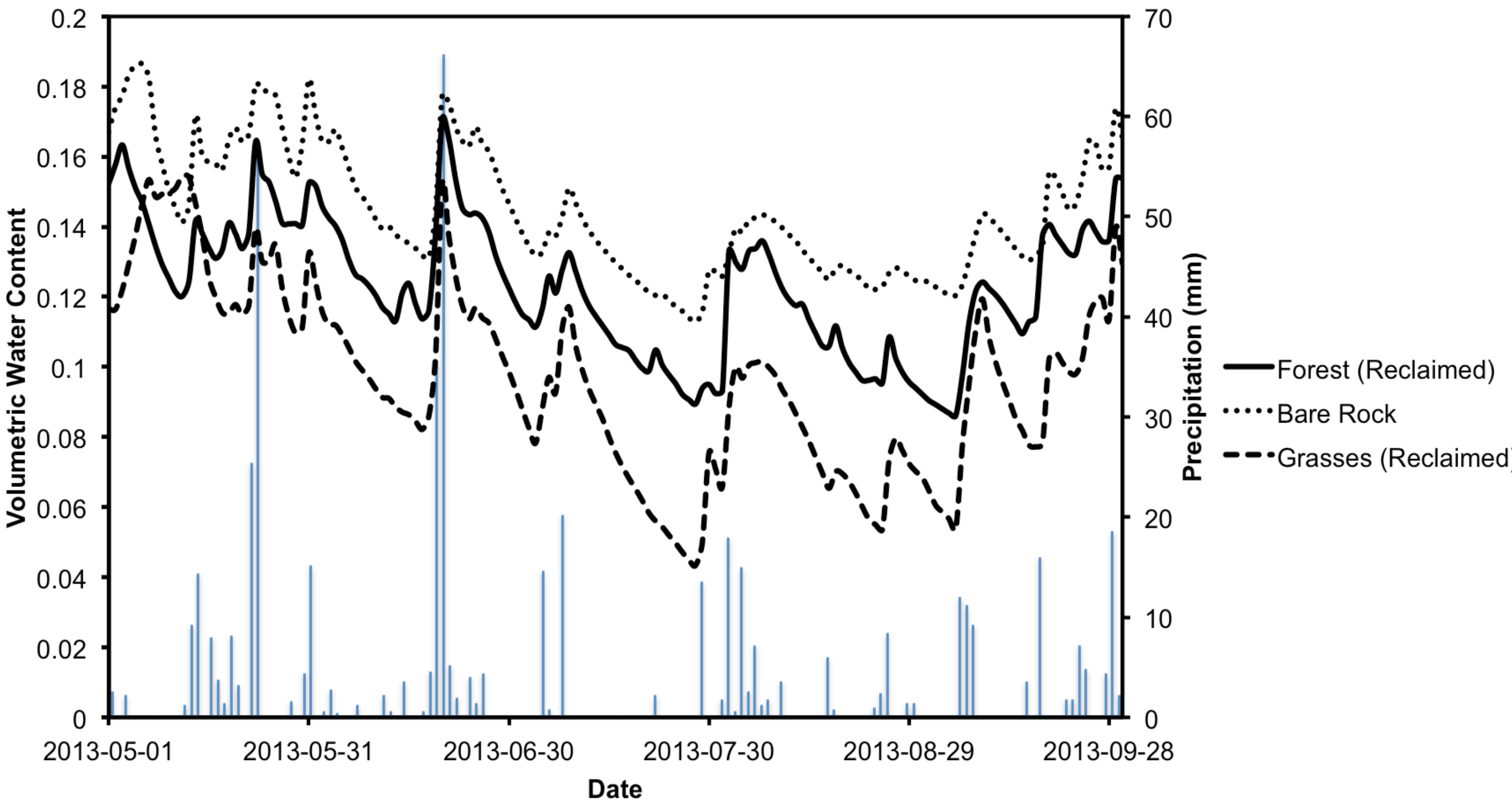

Figure 3.2.1: The mean daily volumetric water content in the top 75 cm of soil at each site over the 2013 growing season. The daily precipitation at the reclaimed grasses site is displayed to illustrate the response to precipitation events.

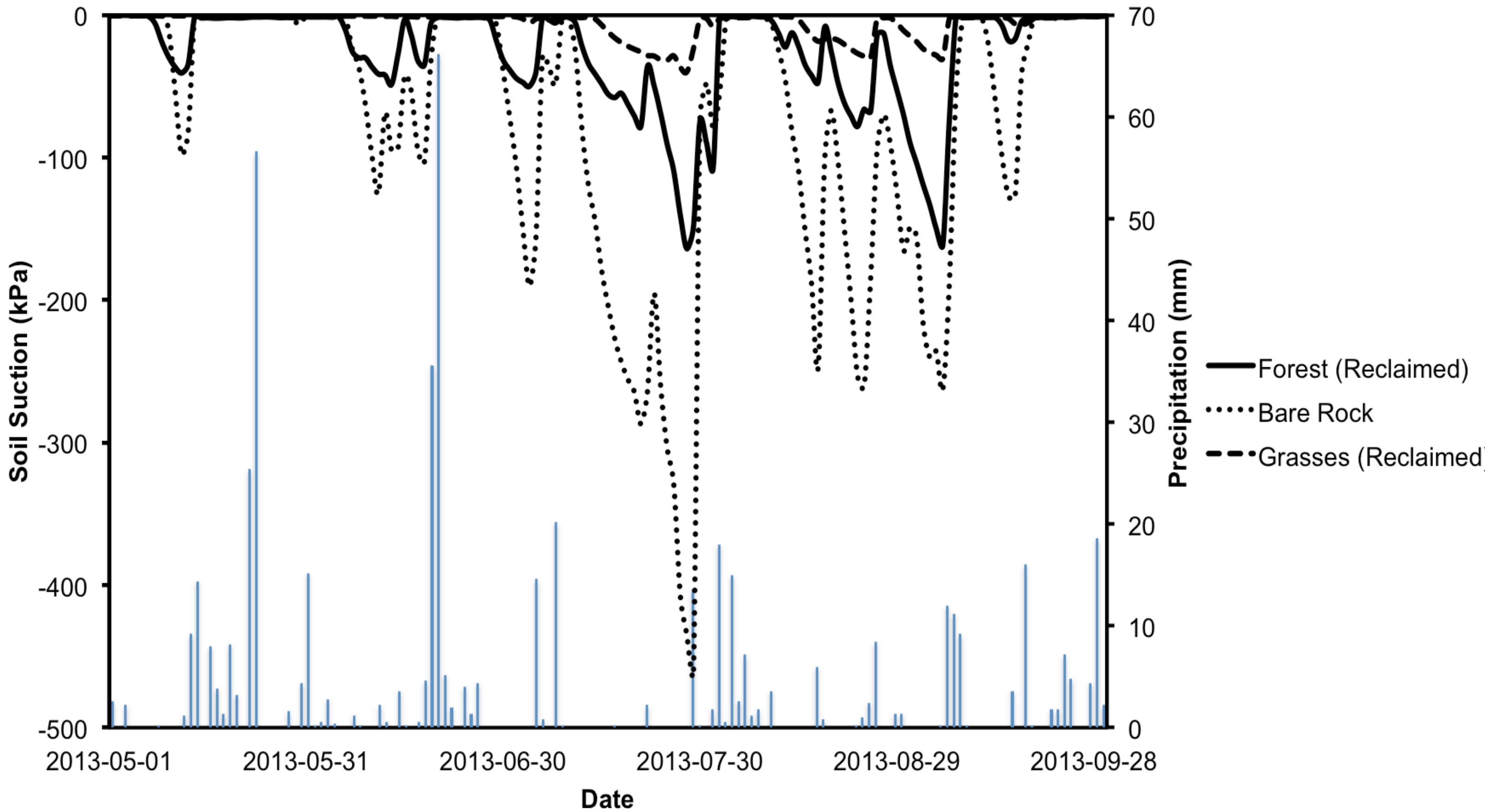

Figure 3.2.2: Mean daily soil suction (at a depth of 10cm) at each of the sites over the 2013 growing season. The daily precipitation at the reclaimed grasses site is presented to illustrate the response of each cover to rain.

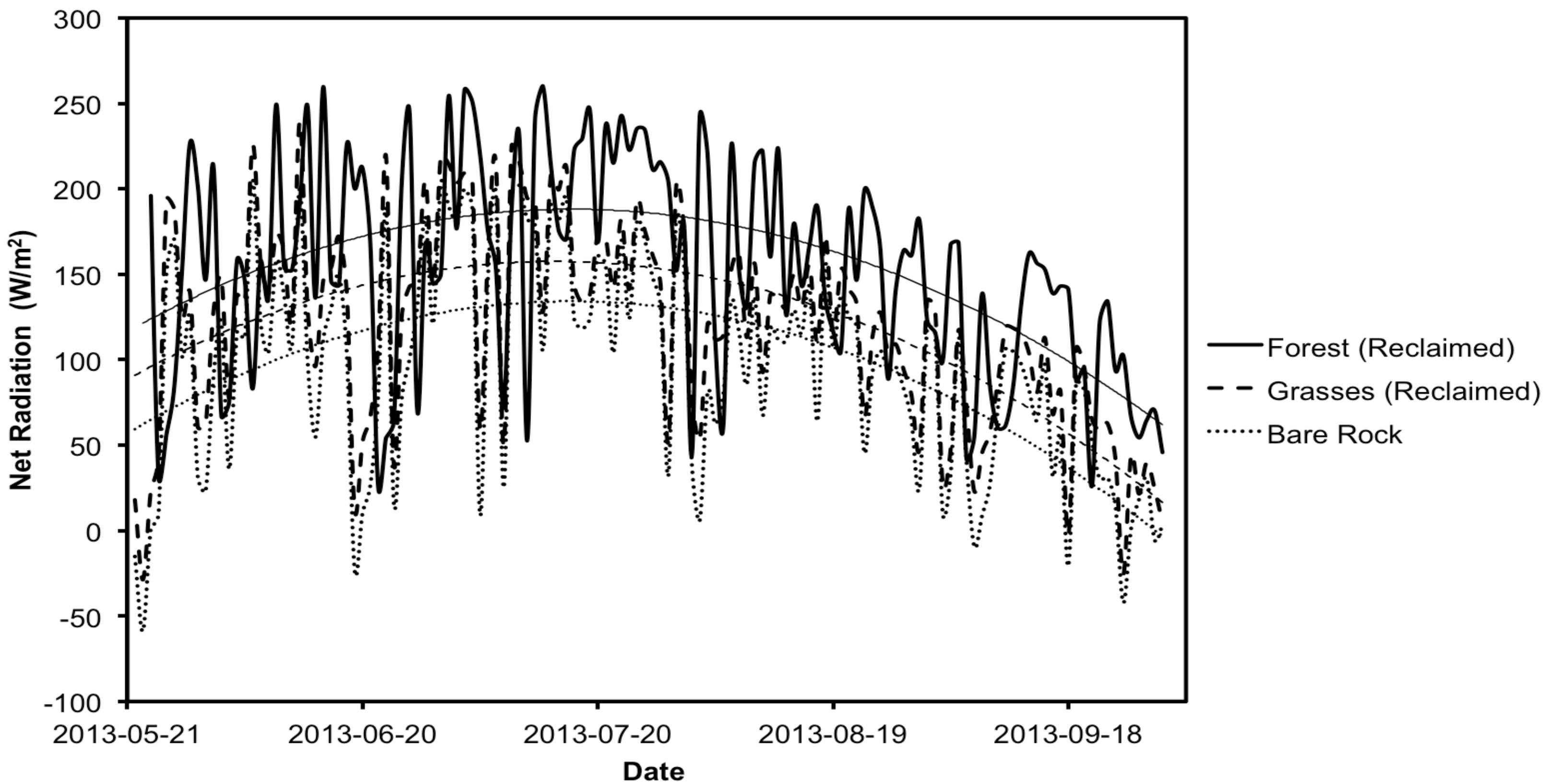

Figure 3.3.1: The mean daily average net radiation (Rn) for each site during the 2013 growing season. Positive values of Rn represent a flux towards the surface. The seasonal trend at each site is represented as a second-order polynomial.

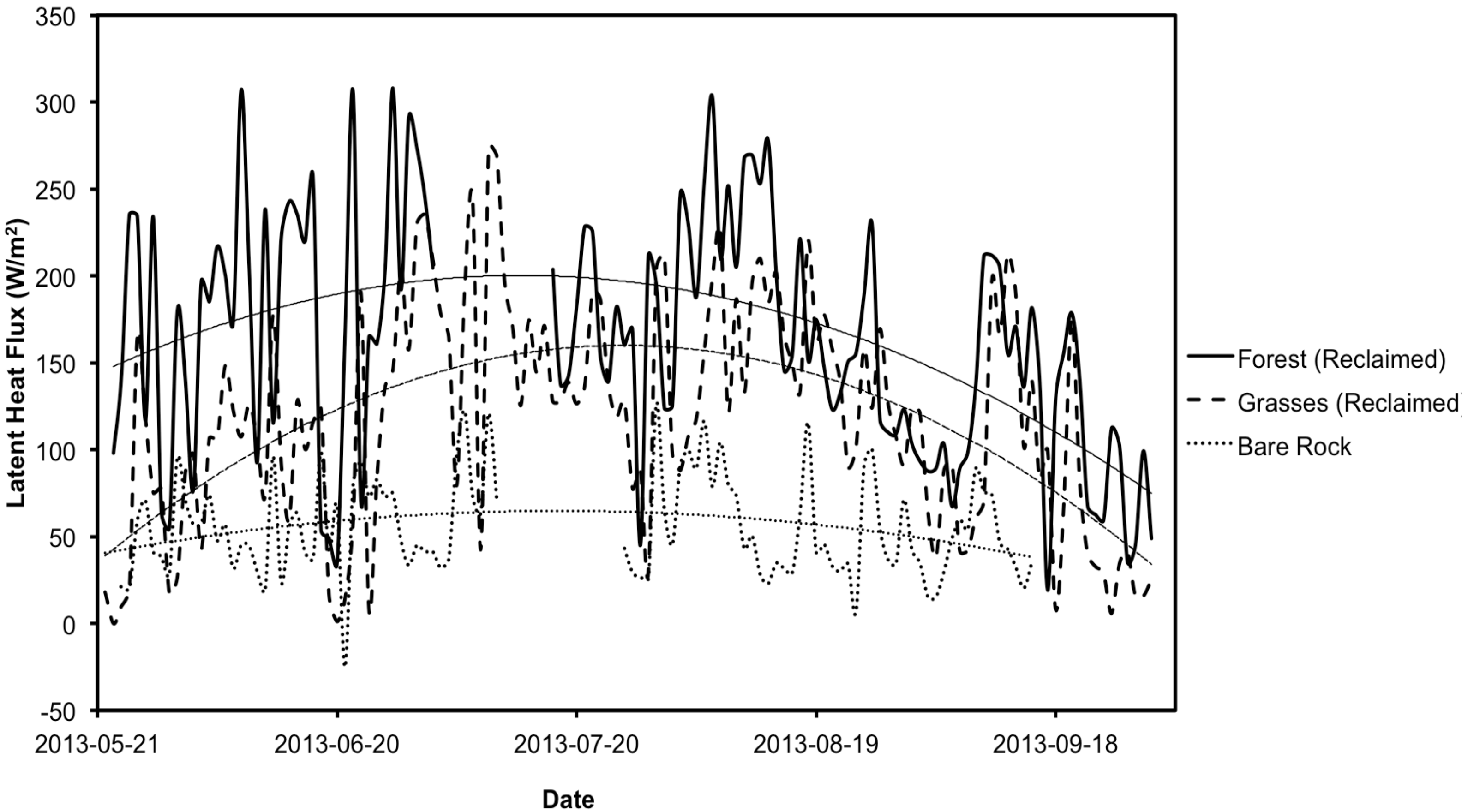

Figure 3.3.2: Daily daytime latent heat flux (LE) at each of the sites during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Positive LE values represent a flux directed away from the surface. The seasonal trend at each site is identified using a second-order polynomial.

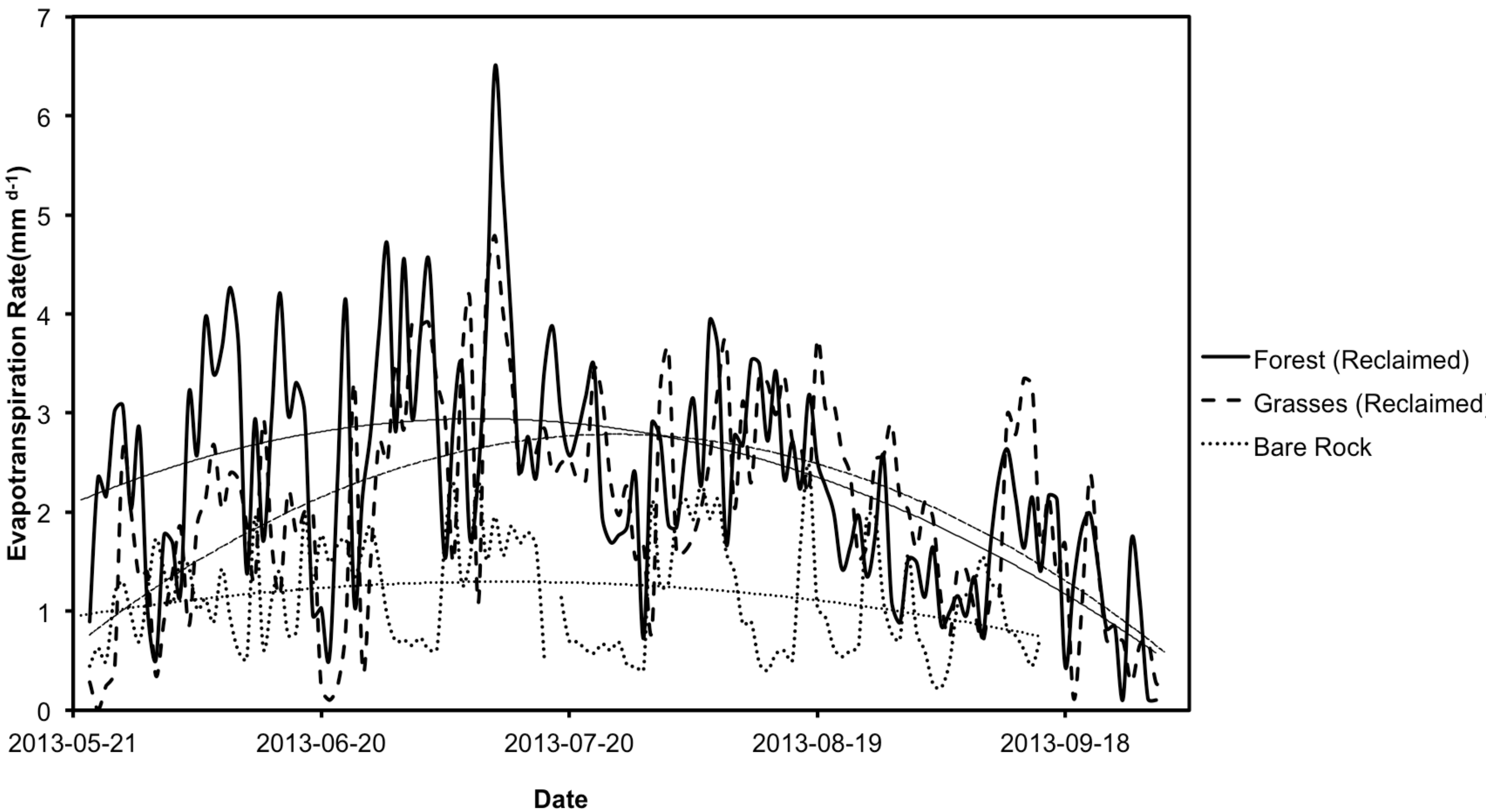

Figure 3.3.3: Mean daily evapotranspiration (ET) rate (mm d^{-1}) at each of the sites during the 2013 growing season. The seasonal trend at each site is identified using a second-order polynomial.

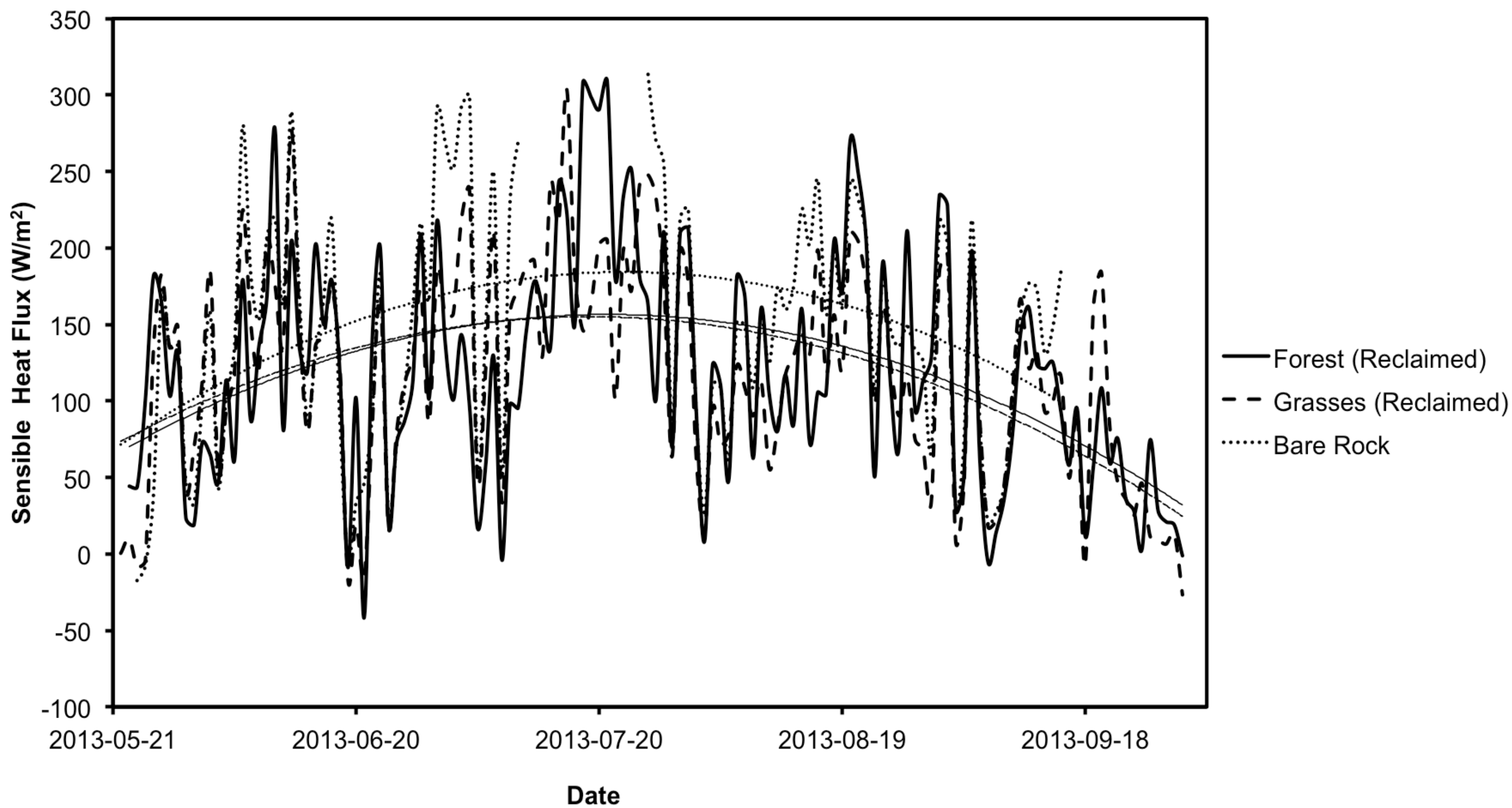

Figure 3.3.4: Mean daily daytime sensible heat flux (H) at each of the sites during the 2013 growing season. Positive H values represent fluxes away from the surface. Daytime hours are from 9:00 to 18:00. The seasonal trend is identified using a second-order polynomial.

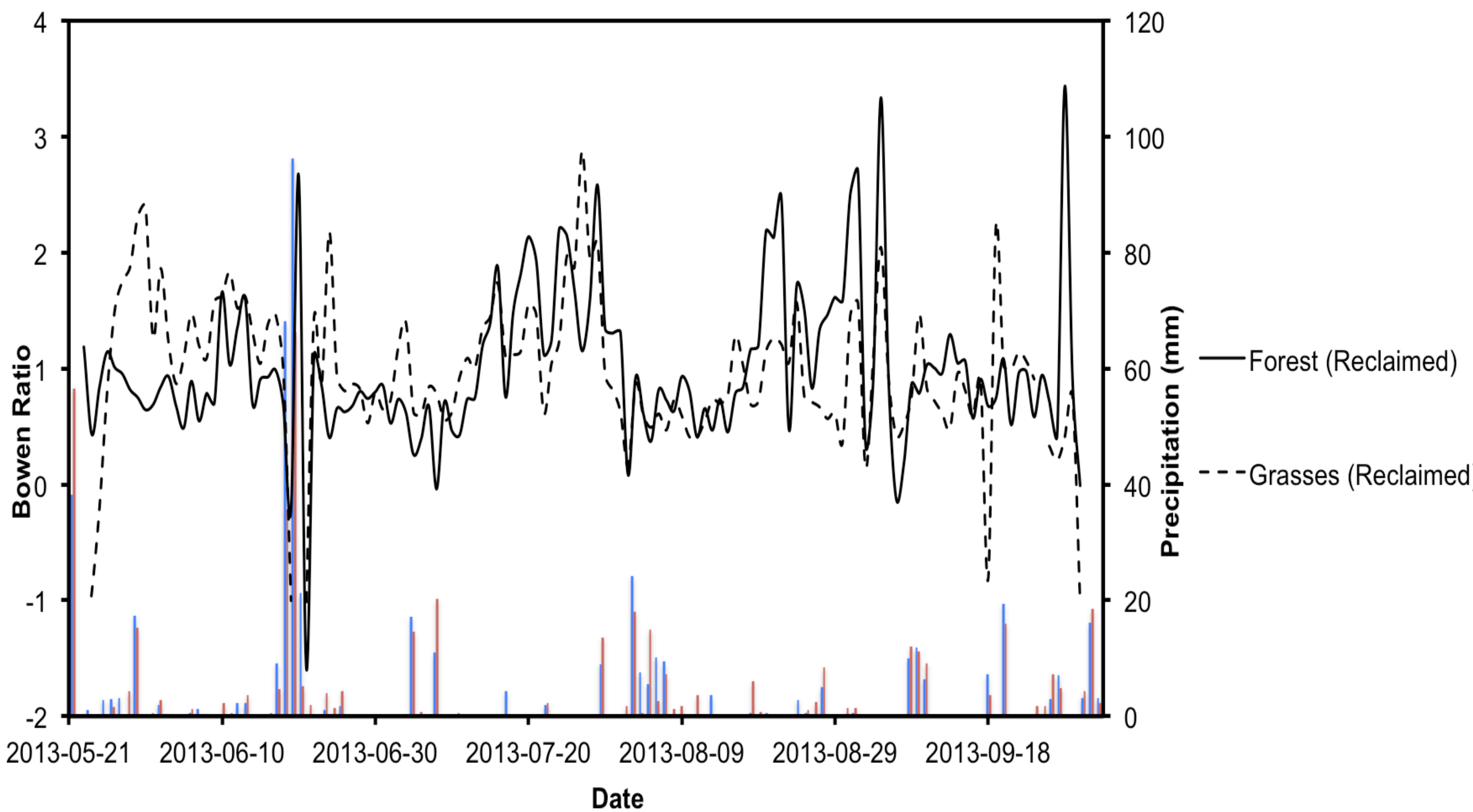

Figure 3.3.5: Bowen ratio (β) at the forest and grasses sites during the 2013 growing season. β was calculated as the ratio of the daily daytime (9:00 – 18:00) sums of H and LE.

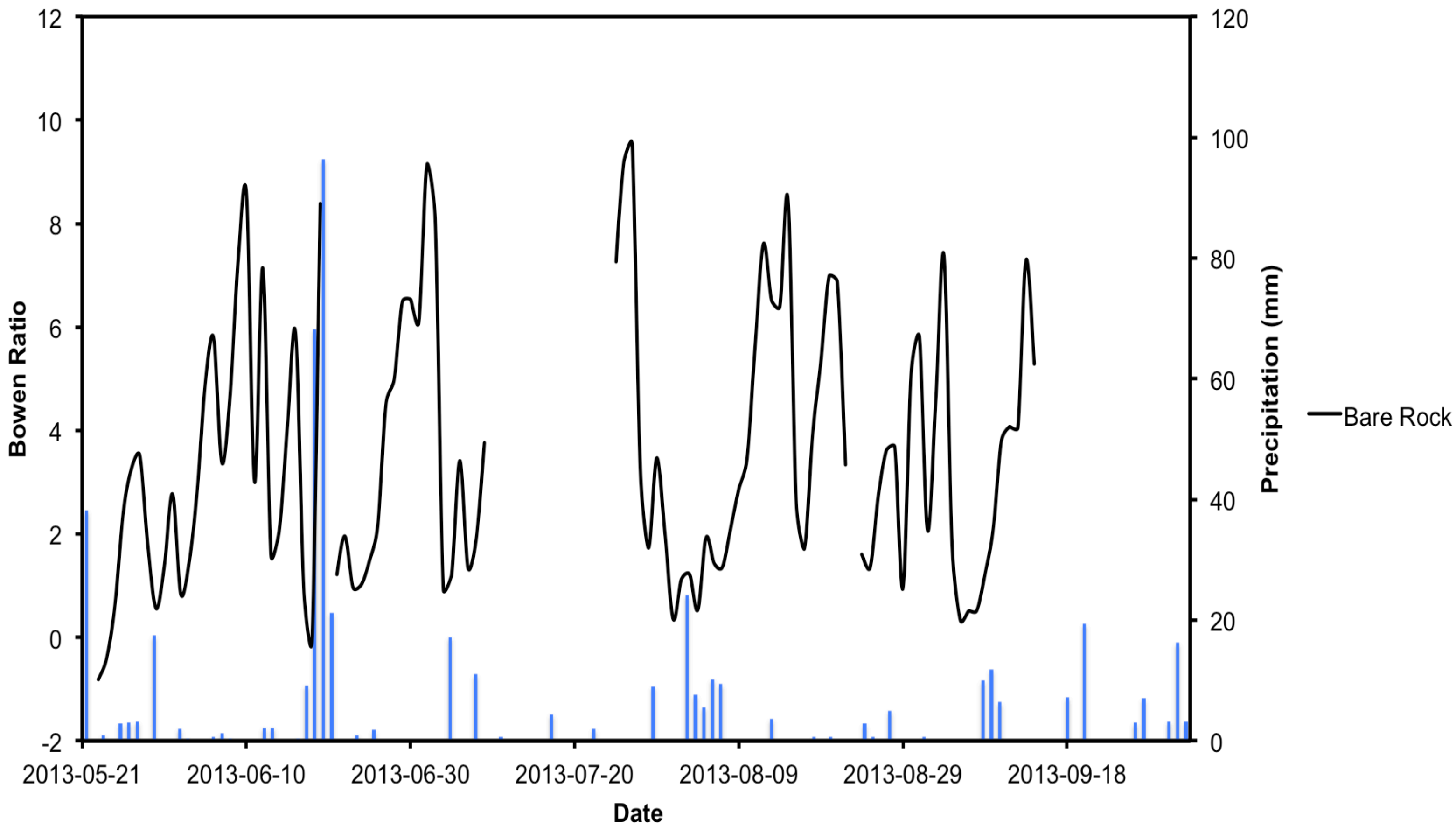

Figure 3.3.6: Bowen ratio (β) at the bare rock site during the 2013 growing season. β was calculated as the ratio of the daily daytime (9:00 – 18:00) sums of H and LE.

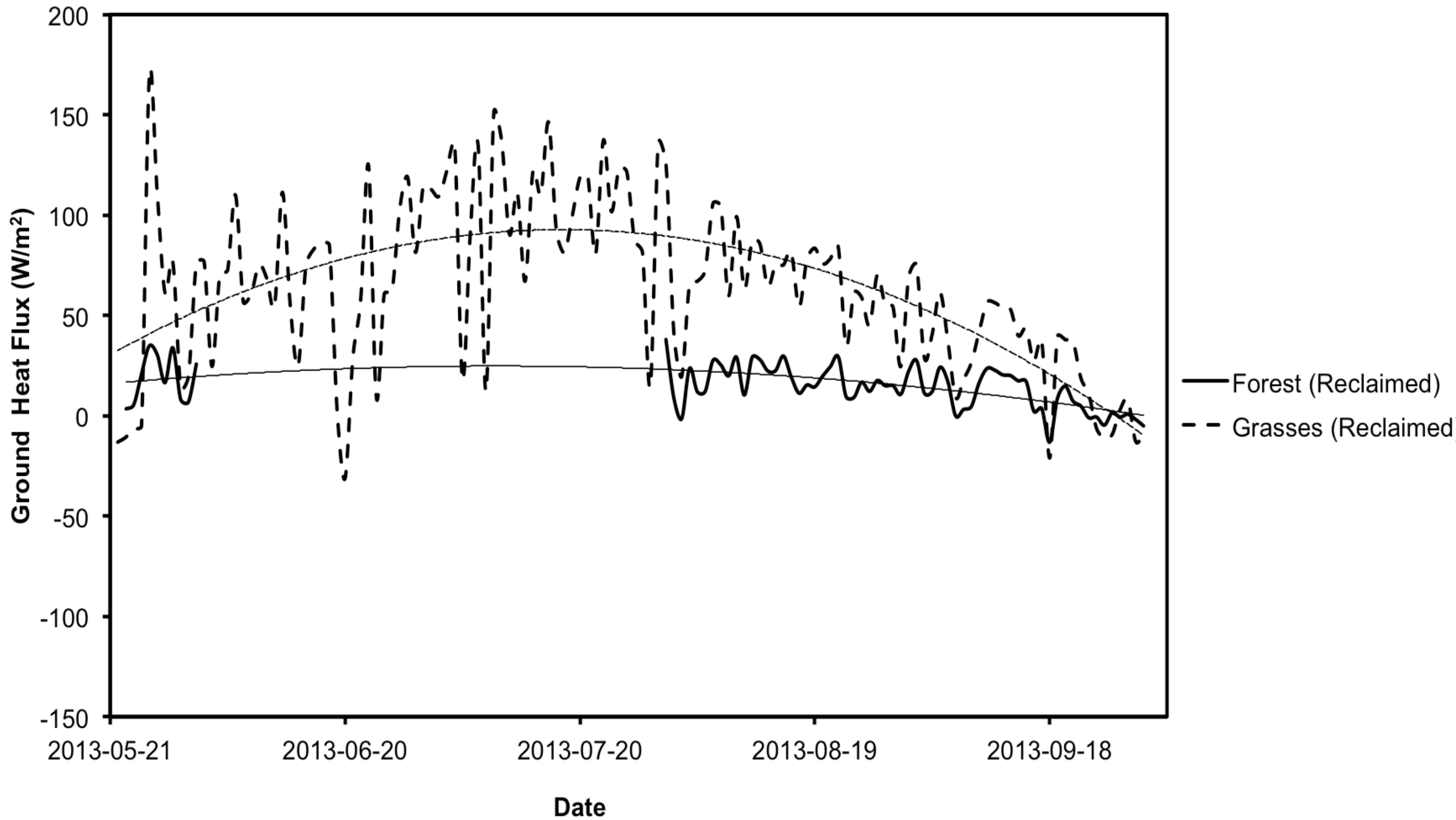

Figure 3.3.7: Daily daytime ground heat flux (G) from the forest and grasses sites during the 2013 growing season. Daytime hours are from 9:00 to 18:00. The seasonal trend is identified using a second-order polynomial.

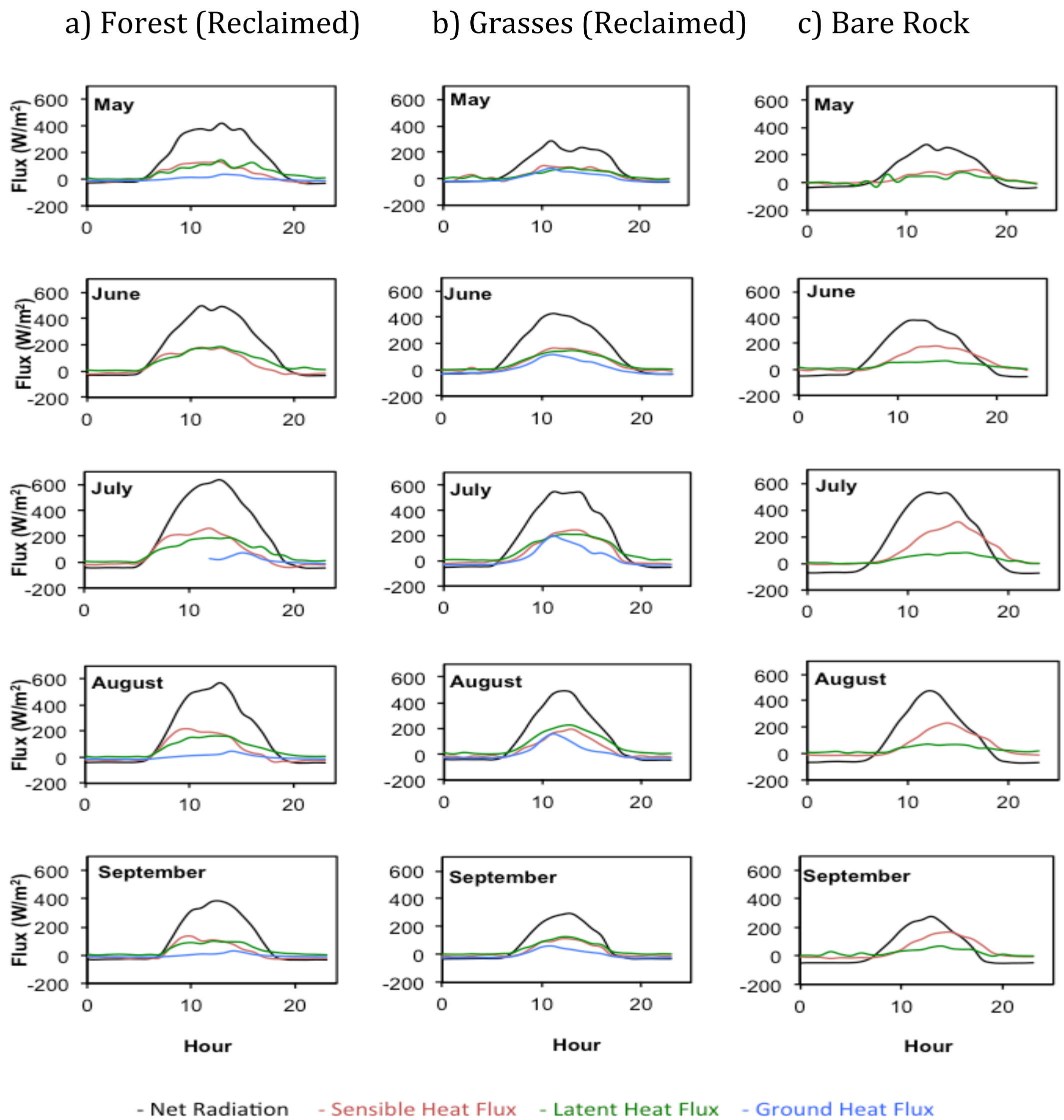

Figure 3.4.1: Mean daily variation of net radiation, the sensible heat flux, the latent heat flux, and the ground heat flux across the 2013 growing season at each site.

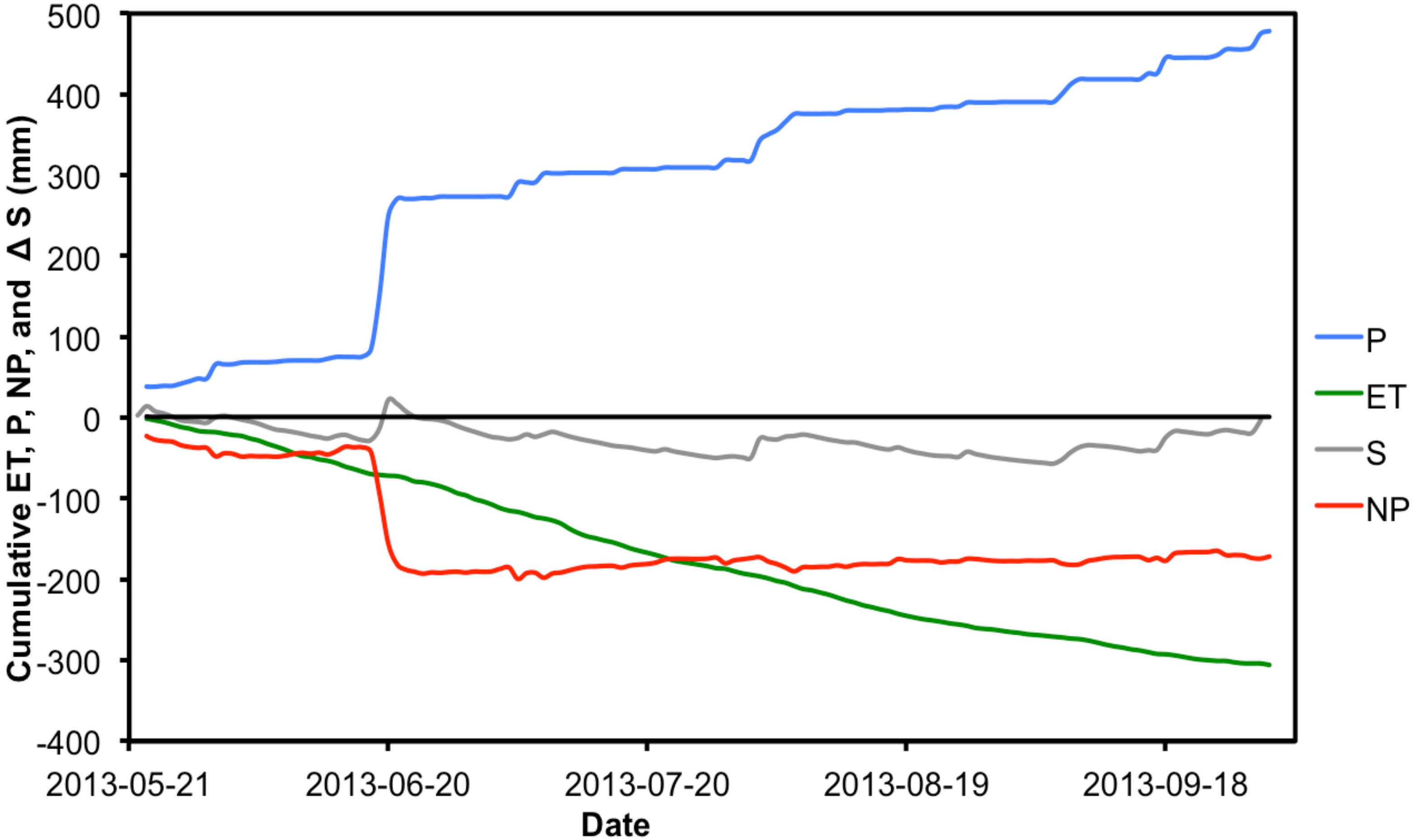

Figure 3.5.1: The cumulative sum of evapotranspiration (ET), precipitation (P), net percolation (NP) and the change in soil moisture storage (of the top 1m) (ΔS) at the reclaimed forest site over 2013 growing season.

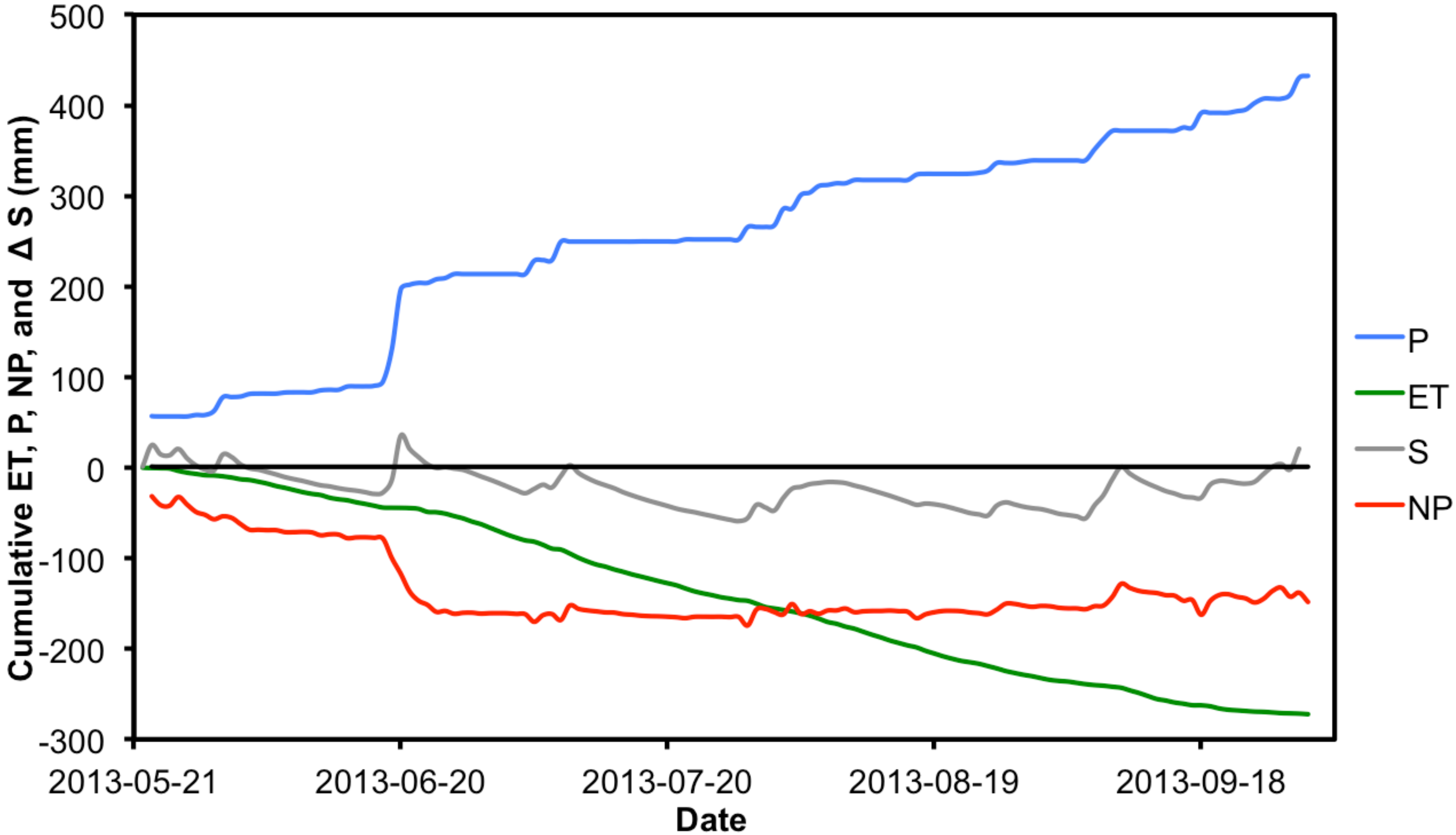

Figure 3.5.2: The cumulative sum of evapotranspiration (ET), precipitation (P), net percolation (NP) and the change in soil moisture storage (of the top 1m) (ΔS) at the reclaimed grasses site over 2013 growing season.

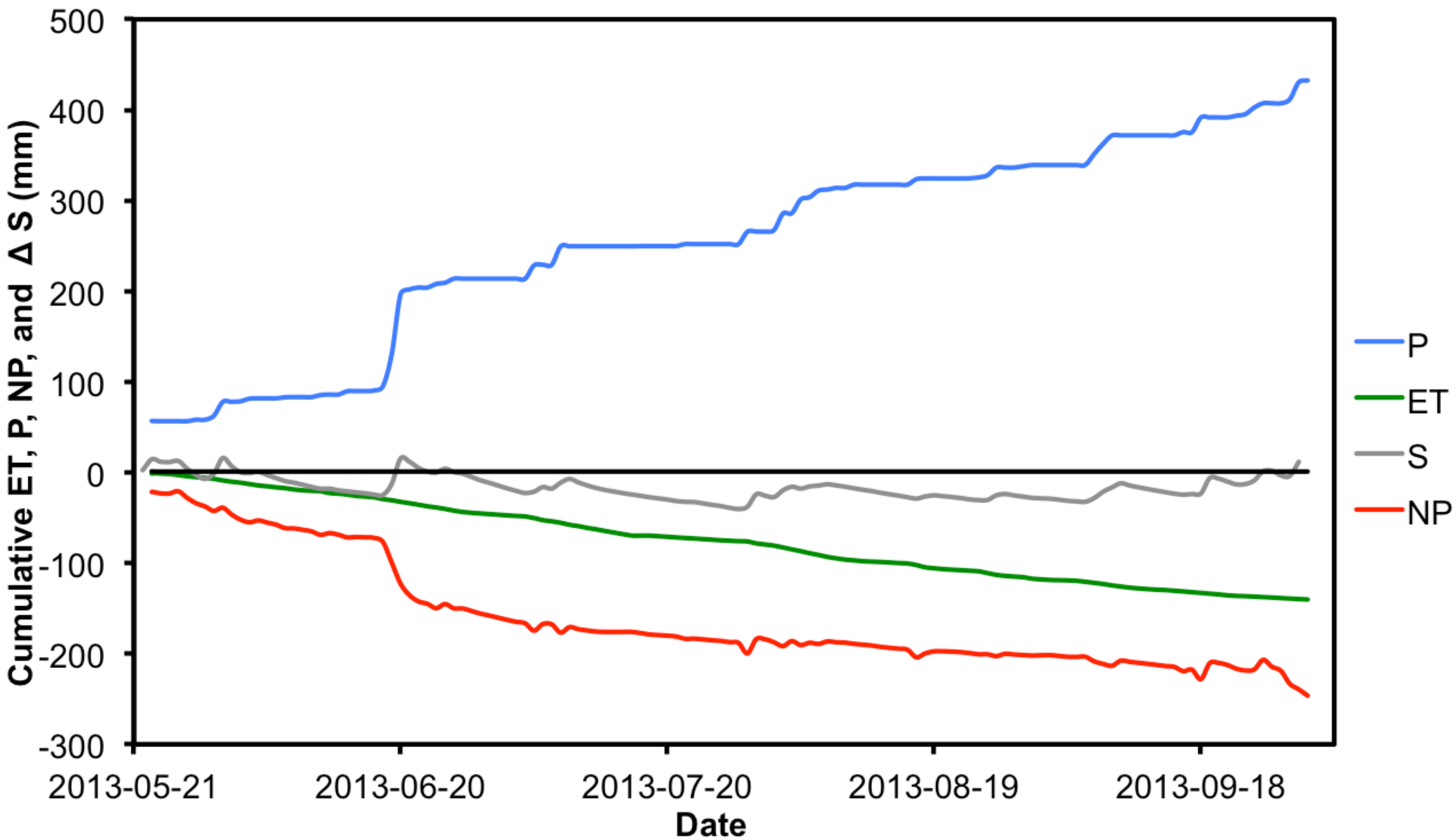

Figure 3.5.3: The cumulative sum of evapotranspiration (ET), precipitation (P), net percolation (NP) and the change in soil moisture storage (of the top 1m) (ΔS) at the bare rock site over 2013 growing season.

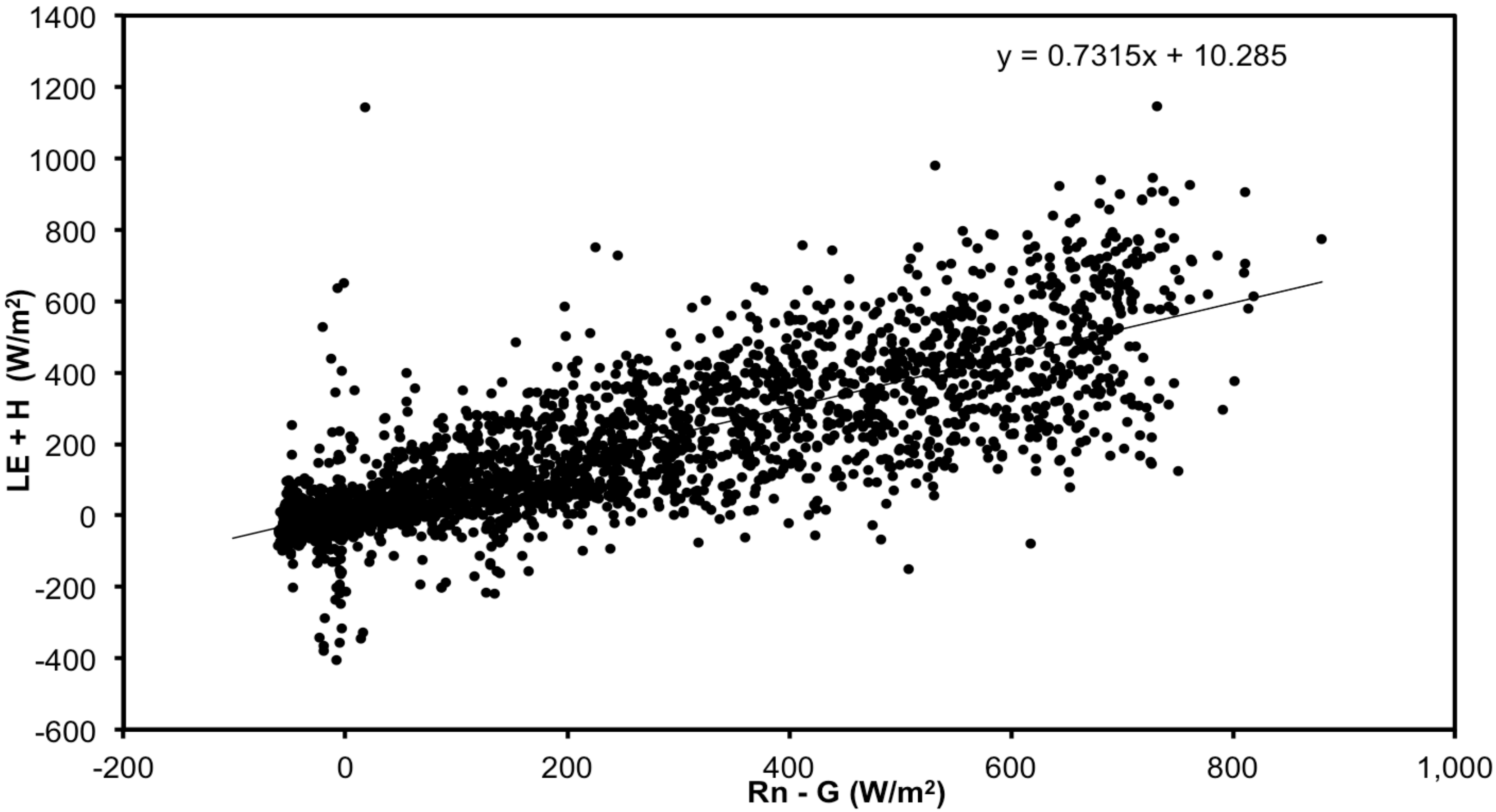

Figure 4.1.1: The estimated closure of the energy balance at the reclaimed forest site over the 2013 growing season. G values were estimated as 10% of the net radiation. The coefficient of determination (R^2) is 0.71.

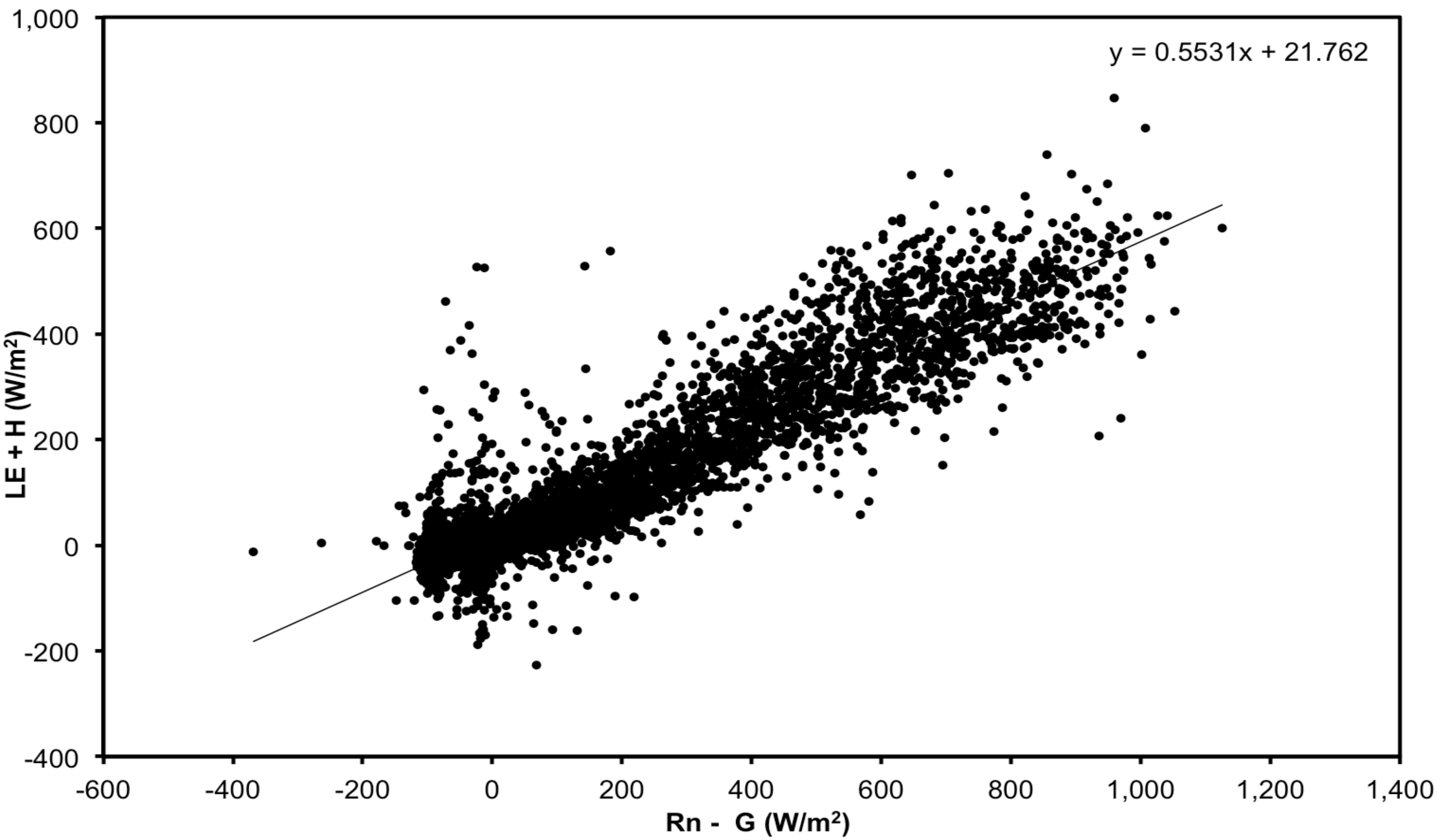

Figure 4.1.2: The closure of the energy balance at the reclaimed grasses site over the 2013 growing season. The coefficient of determination (R^2) is 0.87.

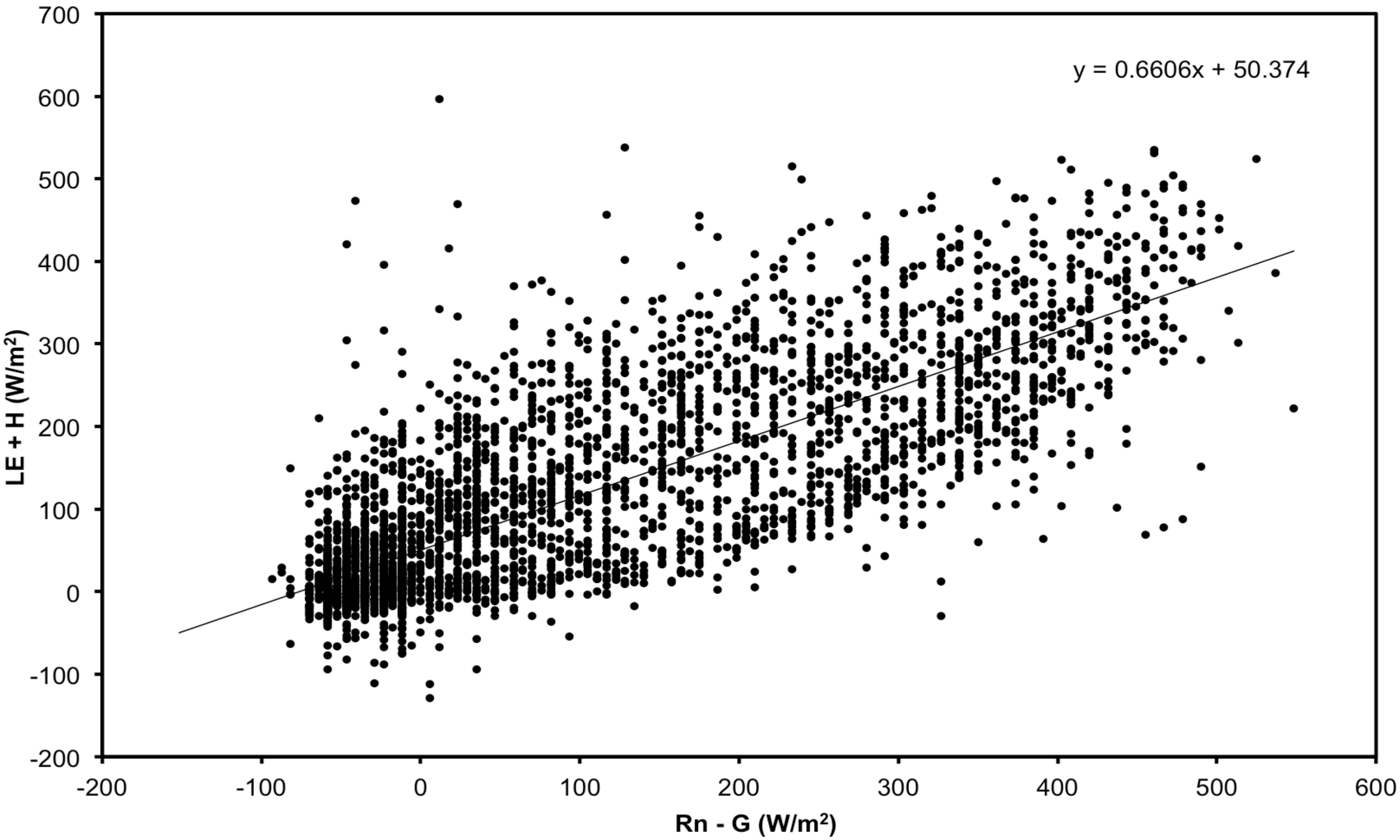

Figure 4.1.3: The estimated closure of the energy balance at the bare rock site over the 2013 growing season. G values were estimated as 30% of the net radiation. The coefficient of determination (R^2) is 0.64.

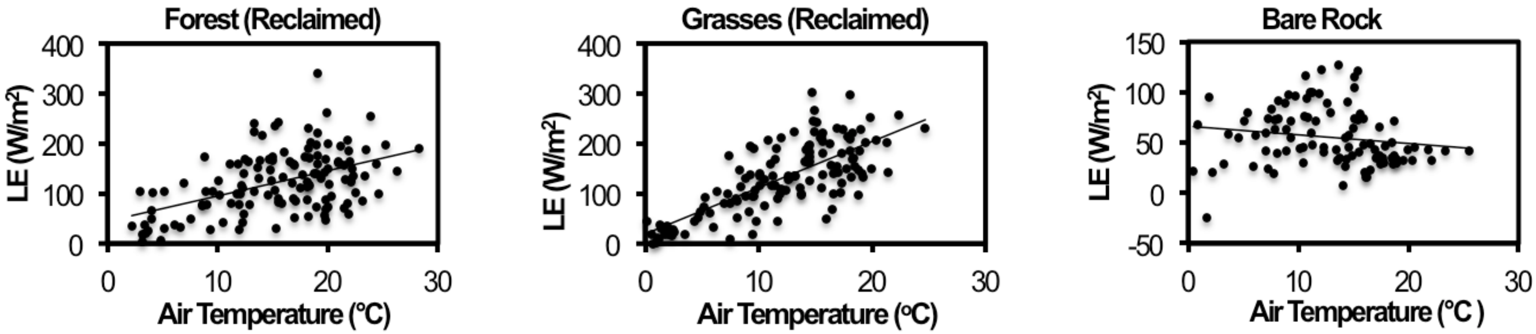

Figure 4.3.1: The relationship between the average daily daytime latent heat flux (LE) and the average daily daytime air temperature at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.24, 0.57, and 0.03 at the forest, grasses, and bare rock sites.

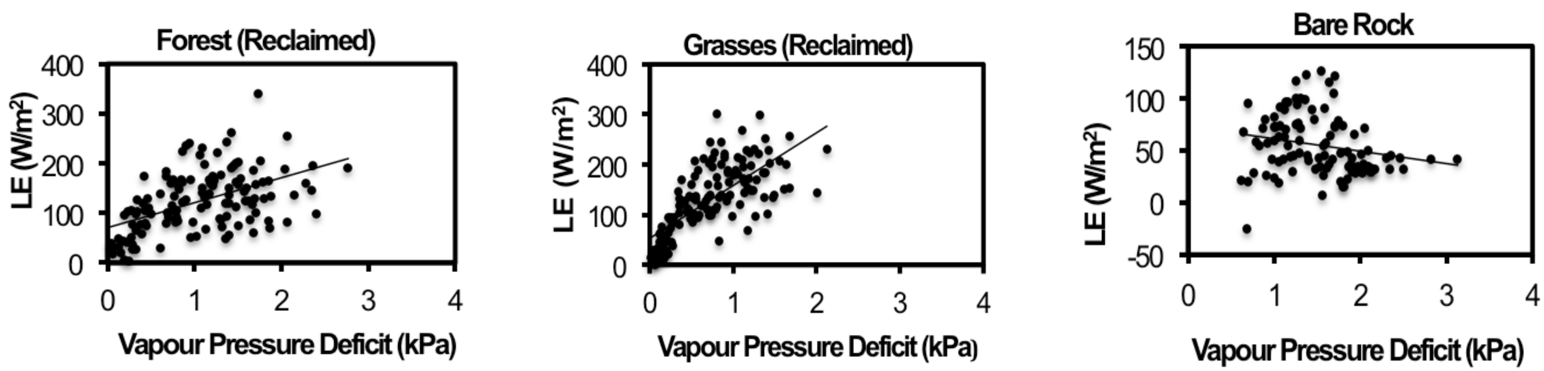

Figure 4.3.2: The relationship between the average daily daytime latent heat flux (LE) and the average daily daytime vapour pressure deficit (VPD) at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.25, 0.51, and 0.04 at the forest, grasses, and bare rock sites.

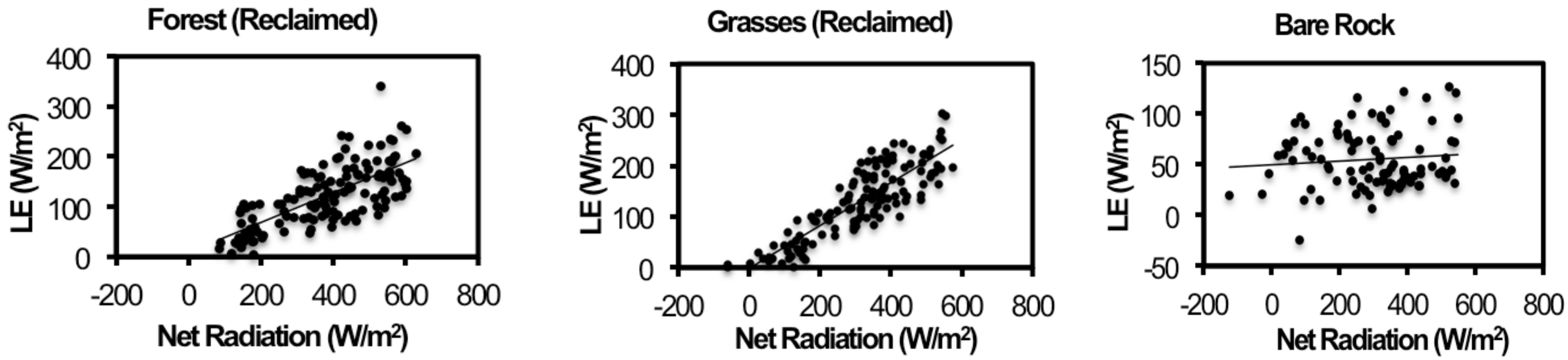

Figure 4.3.3: The relationship between the average daily daytime latent heat flux (LE) and the average daily daytime net radiation (R_n) at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.51, 0.74, and 0.01 at the forest, grasses, and bare rock sites.

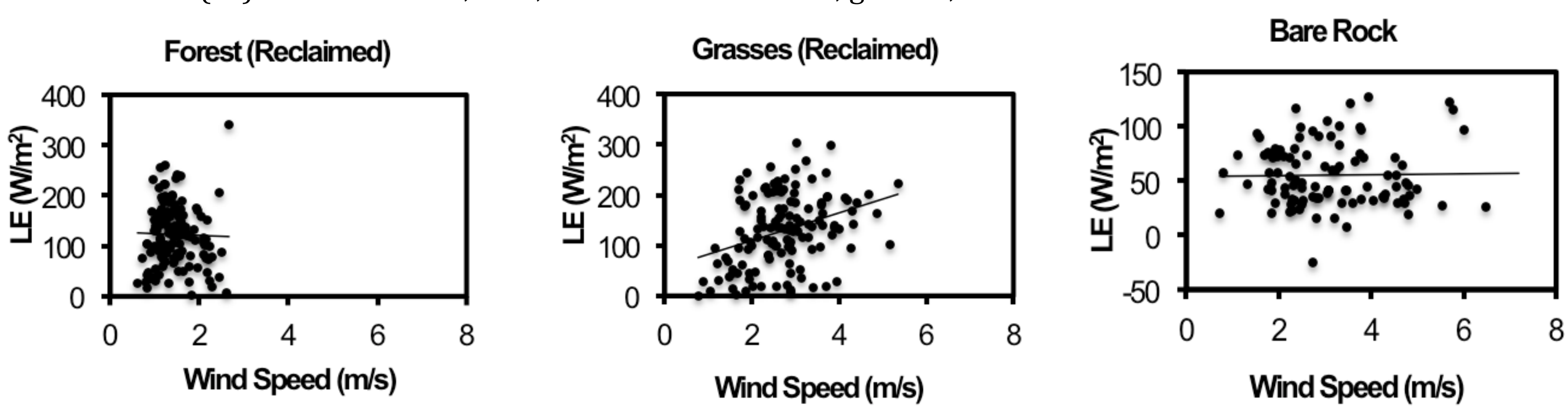

Figure 4.3.4: The relationship between the average daily daytime latent heat flux (LE) and the average daily daytime wind speed at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.00, 0.11, and 0.00 at the forest, grasses, and bare rock sites.

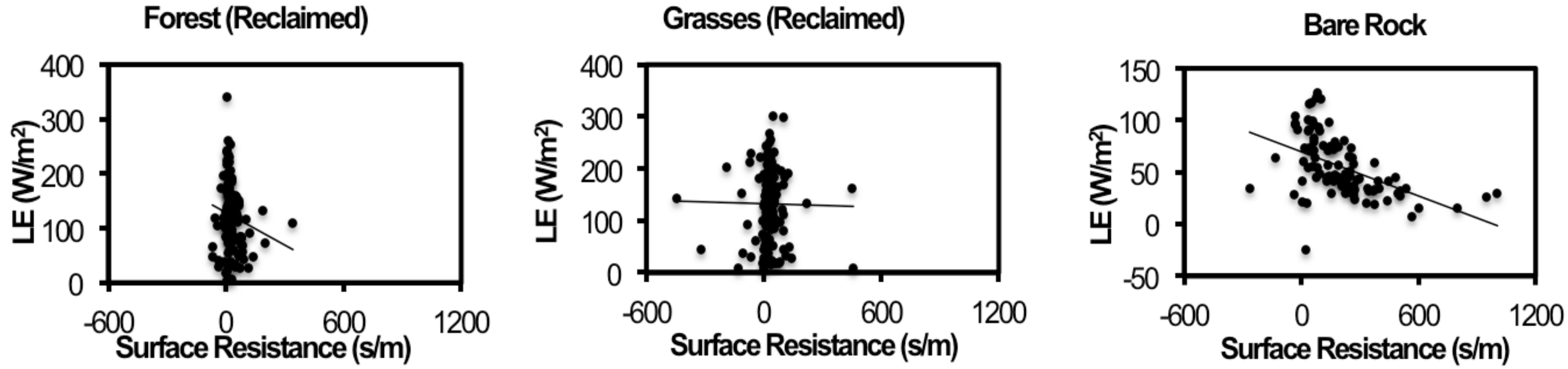

Figure 4.3.5: The relationship between the average daily daytime latent heat flux (LE) and the average daily daytime surface resistance at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.02, 0.00, and 0.25 at the forest, grasses, and bare rock sites.

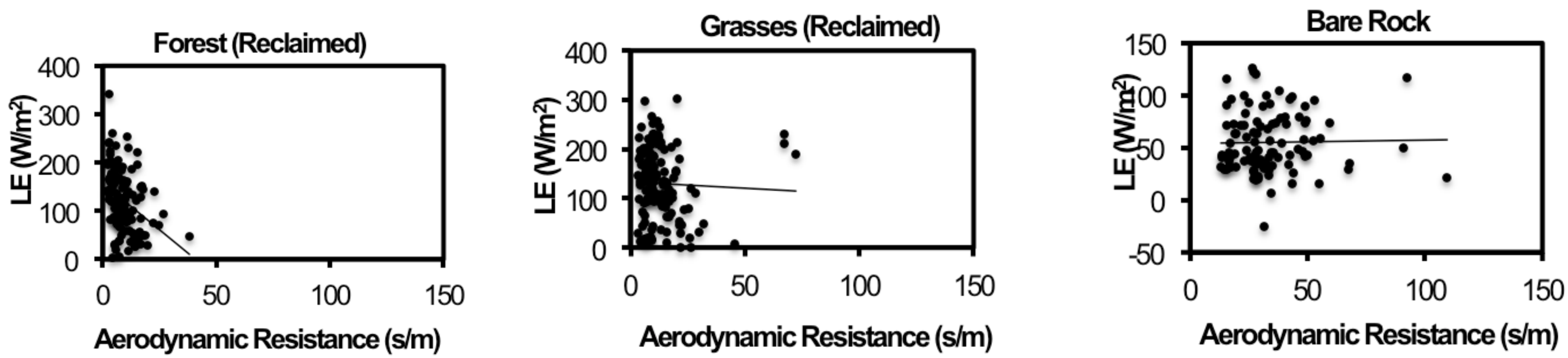

Figure 4.3.6: The relationship between the average daily daytime latent heat flux (LE) and the average daily daytime aerodynamic resistance at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.12, 0.00, and 0.00 at the forest, grasses, and bare rock sites.

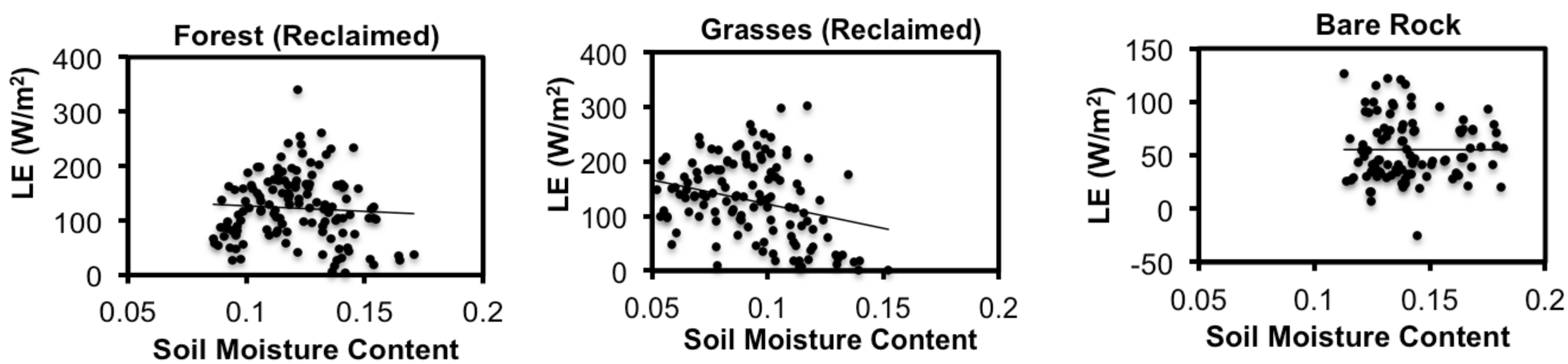

Figure 4.3.7: The relationship between the average daily daytime latent heat flux (LE) and the average soil moisture content at each site during the 2013 growing season. Daytime hours are from 9:00 to 18:00. Coefficient of determination (R^2) values are 0.00, 0.09, and 0.00 at the forest, grasses, and bare rock sites.

Table 3.1.1: Average monthly air temperatures (°C) during the 2013 growing season at the Sparwood weather station and the 1981-2010 climate normal.

	May	**June**	**July**	**August**	**September**	**Seasonal Average**
Sparwood	10.5	13.1	18.0	17.0	12.8	14.3
Climate Normal	9.1	12.7	15.8	15.5	10.5	12.7

Table 3.1.2: Total monthly precipitation (mm) during the 2013 growing season at the Sparwood weather station and the 1981-2010 climate normal.

	May	**June**	**July**	**August**	**September**	**Seasonal Total**
Sparwood	68.2	111.4	32.3	52.6	99.2	363.7
Climate Normal	60.4	69.3	46.8	34.9	47.4	258.8

Table 3.3.1: The average ratio of 24 hour sums of LE/Rn for every month of the 2013 growing season at each site.

	May	**June**	**July**	**August**	**September**	**Seasonal Average**
Forest (Reclaimed)	0.66	0.62	0.46	0.57	0.59	0.58
Grasses (Reclaimed)	0.23	0.33	0.43	0.43	0.51	0.44
Bare Rock	0.36	0.24	0.20	0.36	0.14	0.26

Table 3.3.2: The average daily evapotranspiration (ET) rate (mm d^{-1}) for every month of the 2013 growing season at each site.

	May	**June**	**July**	**August**	**September**	**Seasonal Average**
Forest (Reclaimed)	2.0	2.8	3.0	2.3	1.4	2.3
Grasses (Reclaimed)	1.0	1.9	2.8	2.5	1.3	2.1
Bare Rock	1.0	1.2	1.2	1.2	1.2	1.2

Table 3.3.3: The average ratio of 24 hour sums of H/Rn for every month of the 2013 growing season at each site.

	May	**June**	**July**	**August**	**September**	**Seasonal Average**
Forest (Reclaimed)	0.34	0.35	0.43	0.34	0.23	0.34
Grasses (Reclaimed)	0.19	0.25	0.33	0.30	0.32	0.29
Bare Rock	0.44	0.64	0.66	0.68	0.58	0.63

Table 3.3.4: The average daytime (9:00 – 18:00) Bowen Ratio (β) for every month of the 2013 growing season at each site.

	May	**June**	**July**	**August**	**September**	**Seasonal Average**
Forest (Reclaimed)	0.86	0.80	1.16	1.09	0.99	1.00
Grasses (Reclaimed)	1.18	1.09	1.20	0.77	0.76	0.97
Bare Rock	1.37	3.68	4.51	3.55	3.21	3.50

Table 4.1.1: Maximum and daily average evapotranspiration (ET) rates at various boreal sites.

Vegetation Type	Max. ET Rate (mm d^{-1})	Avg. ET Rate (mm d^{-1})	Author(s)
Black Spruce	3.5		Arain et al. (2003)
Black Spruce		2.0	Jarvis et al. (1997)
Black Spruce		2.24 (1989), 2.33 (1990)	Lafleur 1992
Black Spruce	1.3		Ewers et al. (2005)
Black Spruce		1.9	Moore et al. (2000)
Black Spruce		1.3	Moore et al. (2000)
Aspen	2.5		Ewers et al. (2005)
Aspen	3.2		Ponton et al. (2006)
Douglas-fir	3.7 (1998), 3.4 (1999)	2.7	Humphreys et al. (2003)
Douglas-fir	2.7		Ponton et al. (2006)
Jack Pine	3.5	0.5 to 2.5	Baldocchi et al. (1997)
Jack Pine	3.5		Ewers et al. (2005)
Jack Pine		1.6	Moore et al. (2000)
Jack Pine		1.1	Moore et al. (2000)
Reclaimed forest	6.5	2.3	Fraser et al. (2014)
Grassland	4.5 (1998), 3 (1999, 2000)		Wever et al. (2002)
Grassland	5.0		Ponton et al. (2006)
Reclaimed grasses	4.8	2.1	Fraser et al. (2014)