SOIL RESPIRATION DYNAMICS IN RESPONSE TO CLIMATE OSCILLATIONS AND SHELTERWOOD HARVESTING IN A TEMPERATE PINE FOREST

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LAY ABSTRACT

Coniferous forest plantations in eastern North America are undergoing silvicultural management to enhance their carbon sequestration capabilities and native-tree species diversity. This study investigated the combined influence of climate oscillations and shelterwood harvesting on soil carbon dynamics of a planted pine forest in southern Ontario, Canada. Between pre- and post-harvesting, soil temperature and soil moisture did not show any significant differences. However, soil CO_2 effluxes in post-harvesting years were lower than pre-harvesting years. A Gaussian-Gamma specification model determined that heterotrophic (autotrophic) respiration decreased (increased) post-harvesting and mineral-soil respiration was similar between pre- and post-harvesting. An increase in ecosystem respiration post-harvesting, despite soil CO_2 efflux decreasing and being the largest component, was primarily caused by the increase in autotrophic respiration due to enhancement in forest growth. This study improved the understanding of forest carbon dynamics by highlighting the importance of accounting for all components, which may contribute to ecosystem respiration. Results can be useful for forest management practitioners, specifically those focused on carbon sequestration and forest conservation.

ABSTRACT

Understanding forest carbon uptake and associated growth response is important for carbon sequestration and water management practices given the large quantities of carbon stored in forest ecosystems. Climate variability and forest management practices influence the magnitude and rate of soil CO_2 efflux; however, their combined effects are complex and not well understood. This study investigated the response of soil CO_2 efflux to the combined effects of climate variability, including those caused by climate oscillations, and shelterwood harvesting in a mature temperate white pine (*Pinus strobes L.*) forest, located near Lake Erie in southern Ontario, Canada. Analyses indicated that local winter temperatures and precipitation were influenced by climate oscillations, which affected forest carbon dynamics. After the shelterwood harvest removed approximately a third of the overstory canopy, no significant differences were found for soil temperature and soil moisture between the pre-harvesting (2008 to 2011) and post-harvesting (2012 to 2014) periods. Despite similar climate conditions pre- and post-harvesting, soil CO_2 effluxes post-harvesting were lower. A Gaussian-Gamma specification model determined that heterotrophic (autotrophic) respiration decreased (increased) between pre- and post-harvesting, respectively. Mineral-soil respiration were similar pre- and post-harvesting. Soil CO_2 efflux accounted for 78±9% of the annual ecosystem respiration (R_E), derived using eddy-covariance fluxes. However, the overall net ecosystem productivity showed no significant difference between pre- and post-harvesting.

This was attributed to an increase in the gross ecosystem productivity post-harvesting, compensating for the increased losses (i.e. increased R_E). This study highlights the complexities of measuring various components of ecosystem respiration after a disturbance, such as a harvest. The knowledge gained from this study provides a better understanding of climate variability and shelterwood harvesting influences on ecosystem respiration and can be useful for forest managers focused on carbon sequestration and forest conservation.

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CHAPTER 1

INTRODUCTION

Forest ecosystems account for approximately 80% of the world's biomass and are considered an important sink for atmospheric carbon dioxide (CO_2) (Kindermann et al., 2008; Pan et al., 2011). They sequester large amounts of atmospheric CO_2 through photosynthesis, referred as gross ecosystem productivity (GEP), and return a substantial portion of fixed carbon back to the atmosphere in the form of ecosystem respiration (R_E), comprising of autotrophic and heterotrophic respiration components (Lorenz, 2010). Only a small fraction of this assimilated carbon is stored in above-ground and below-ground biomass, litter and soil, which combined are referred to as net ecosystem productivity (NEP) (Chapin et al., 2009; Lorenz, 2010).

Past studies show that the forest carbon cycle is strongly influenced by climate variability (Allen et al., 2010; Bastos et al., 2013; Heimann and Reichstein, 2008; Hogg et al., 2017; Noormets et al., 2008; Piao et al., 2008; Wharton et al., 2012). Predicted changes in future climate, and the frequency, and/or severity of extreme weather events, such as droughts, flooding, heatwaves, duration and timing of the winter, and freeze/thaw periods, can have an influence on carbon fluxes from forest ecosystems (Amiro et al., 2010; Davis et al., 1997; Granier et al., 2007; Hogg et al., 2017; Intergovernmental Panel for Climate Change (IPCC), 2013; Reichstein et al., 2013; Vose et al., 2012; Wu et al., 2017;

Wullschleger and Hanson, 2006; Zscheischler et al., 2014). A warming-induced increase in the length of the growing season is expected to increase carbon assimilation (Black et al., 2000; Keenan et al., 2014; Piao et al., 2008; Reay et al., 2007). With increasing atmospheric CO_2 levels and climate warming, some studies suggest that forest ecosystems may sequester more carbon (Drake et al., 2011; Lal, 2005; Peters et al., 2013; Thomas et al., 1996), whereas others predict increasing CO_2 levels and climate warming may either increase below-ground allocation of carbon, leading to an increase in carbon release, or soil fertility, and associated feedbacks, may restrain the response of carbon sequestration to increasing atmospheric CO_2 levels (Chapin et al., 2009; Frelich et al., 2010; Hungate et al., 1997; Luo et al., 1996; Oren et al., 2001). The quantitative understanding of these effects is challenging due to feedback processes, in particular those associated with the availability and amount of below-ground plant biomass and soil carbon pools, which play a major role in regulating the overall forest carbon cycle (Clemmensen et al., 2013; Noormets et al., 2008). Therefore, a better understanding of soil carbon processes is crucial for future predictions of the response of forest soils, and their overall carbon budgets, to climate variability and forest management practices (Chapin et al., 2009; Clemmensen et al., 2013; Reichstein et al., 2013).

Soil CO_2 efflux, also referred to as soil respiration, is the largest component of forest ecosystem respiration (Khomik et al., 2010). It is comprised of autotrophic respiration from roots and heterotrophic respiration from mycorrhizal fungi, microorganisms in the rhizophere and decomposition of litter and soil organic matter in the organic and mineral-soil horizons (Lorenz, 2010). Removal of the litter-layer may modify soil water content at the near-surface soil layer, which may affect heterotrophic activity and disturb the root-soil interface (Hanson et al., 2000).

Previous studies have measured total soil respiration in forested ecosystems; however, only a few have attempted to separate soil respiration into its autotrophic and heterotrophic components (Khomik et al., 2010; Scott-Denton et al., 2006; Wang et al., 2013). It is crucial to study these two major components of soil respiration since they may respond differently to changes in environmental factors (Hanson et al., 2000; Scott-Denton et al., 2006). One method to separate total soil respiration into its autotrophic and heterotrophic components is called trenching; a root exclusion method by means of inserting a barrier around the periphery of a sampling plot, such that the soil inside the plot is detached from the roots and its surroundings (Comstedt et al., 2011). A problem associated with this method includes the decomposition of remaining roots within the sampling plot that can lead to an overestimation of heterotrophic respiration. However, some studies have shown that this overestimation tends to only last a few months after trenching and initial measurements can be omitted from analysis (Bowden et al., 1993; Comstedt et al., 2011). Bowden et al. (1993) found that fine root decomposition had little impact on soil respiration measurements, but suggested it was best to avoid large roots by placing trenched plots away from tree stems.

Also, trenching techniques terminate plant uptake of water, which may result in higher soil water content in trenched plots compared to control plots. The microbial respiration portion of heterotrophic respiration may be stimulated due to the availability of additional water because of lower vegetation uptake or suppressed due to water logging or soil saturation (Comstedt et al., 2011).

In cold regions, snow can also influence carbon cycling by creating an insulating layer that can decouple the warmer soil matrix from the colder air temperatures, generating a microclimate in the near-surface soil that affects the activity of decomposing soil organisms, thus affecting the rate of soil respiration (Monson et al., 2006). Therefore, it is important to analyze the variability of soil respiration in the winter, given that part of the carbon assimilated during the preceding growing season can be released the following winter (Monson et al., 2006; Schindlbacher et al., 2014). Changes in climates, through increased warming or increased variability in the timing and amount of winter precipitation, can lead to changes in snow cover thickness and extent, directly impacting soil respiration during the winter season (Schindlbacher et al., 2014).

Several studies have focused on understanding the linkages between local climatic variables, such as soil temperature and soil moisture, and forest carbon fluxes. However, local conditions are often the result of larger regional climate oscillations. In North America, low frequency climate oscillations, such as the El Niño-Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), the

Arctic Oscillation (AO), the Pacific Decadal Oscillation (PDO), the Eastern Pacific Oscillation (EPO) and the Pacific-North American (PNA) teleconnection pattern, are important atmospheric phenomenon that influence seasonal and annual climate variations (Bonsal et al., 2001; Shabbar et al., 1997; Shabbar and Bonsal, 2004; Zhang et al., 2011). Time scales of these oscillations range from interannual to interdecadal, and their influences are typically stronger during the winter, but can persist into the spring and summer months. Changes in atmospheric circulation associated with these climate oscillations can lead to changes in seasonal temperatures and precipitation patterns, including snow. With carbon flux data records now reaching greater than decadal in lengths, it is now possible to assess meaningful impacts of some of these climatic oscillations on local forest carbon dynamics in North America.

In addition to environmental and edaphic (soil influenced) factors mentioned above, silvicultural practices, such as shelterwood harvesting, can also influence forest carbon dynamics by altering soil temperature, soil moisture, soil organic matter, root biomass and microbial activity (D'Amato et al., 2011; Gustafsson et al., 2012; Magruder et al., 2013; Peng et al., 2008; Powers et al., 2009; Vesala et al., 2005). These silvicultural practices are being adopted in North America and across the world, not only to manage forests, particularly planted stands, but also to enhance their carbon sequestration capabilities and/or to restore them to their previous native species (Gustafsson et al., 2012). Shelterwood harvesting encourages natural regeneration and development of seedlings in partial shade (Ontario Ministry of Natural Resources and Forestry (OMNRF), 1998; OMNRF, 2000; Trant, 2014). Shelterwood harvesting, or selective thinning, which changes tree spacing, number and size distribution, can affect the dynamics of carbon uptake through above-ground productivity and carbon losses through soil respiration processes (Marchi et al., 2018; Vesala et al., 2005). Greater open canopy alters the radiation dynamics, interception of precipitation, wind flow patterns, water vapour deficit, soil temperature and soil moisture (Stoffel et al., 2010; Vesala et al., 2005; Woods et al., 2006).

Peng et al. (2008) found that amongst the studies that examined the impacts of forest harvesting on soil respiration, only a few have examined silviculture practices, such as selective harvesting or thinning effects, with little information on spatial and temporal patterns of soil CO_2 efflux response in these forests, particularly in conifer stands. Additional information is needed to better understand the effects of forest selective thinning on soil processes and soil carbon sequestration (Jandl et al., 2007). This is important because an increasing number of temperate conifer forests in eastern North America are being managed to enhance their carbon sequestration capabilities to restore them to their native species composition, enhance biodiversity and conserve water resources. Therefore, there is an urgent need to enhance our understanding of the effects of silvicultural practices, in particular shelterwood harvesting, on soil processes and carbon sequestration (Jandl et al., 2007).

1.1 OBJECTIVES

The main objectives of this dissertation are:

1. Investigate the occurrence and trends of low frequency climate oscillations in the Great Lakes region in Canada over the past seven decades (1950 to 2014) and quantify their influence on carbon fluxes, including the gross ecosystem productivity (GEP), ecosystem respiration (R_E) and net ecosystem productivity (NEP), in a temperate pine forest in the region, using long-term eddy-covariance flux data from 2003 to 2014.

2. Examine the response of soil CO_2 efflux and its components to changes in forest microclimate, soil temperature and soil moisture, after a shelterwood harvesting event by comparing pre- (2008 to 2011) and post-harvesting (2012 to 2014) measurements.

3. Develop and utilize a Gaussian-Gamma specification model to explore how major components of soil CO_2 efflux, such as autotrophic and heterotrophic respiration, and soil CO_2 efflux from the mineral-soil horizon, respond to soil temperature, soil moisture, shelterwood harvesting and climate oscillations from 2008 to 2014.

The knowledge gained from this research will help in developing polices for better managements of forest ecosystems in eastern Canada, in addition to assisting in planning and developing realistic strategies to offset fossil fuel CO_2 emissions.

1.2 STUDY AREA

Carbon flux measurements and supporting environmental measurements, used in this study, were collected from a mature white pine forest near Turkey Point Provincial Park on the northern shore of Lake Erie in southern Ontario, Canada. This site is part of the Turkey Point Flux station (TPFS), (also known as the Turkey Point Observatory), of the Global Water Futures (GWF) and global Fluxnet initiatives. The forest is dominated (>82%) by eastern white pine (*Pinus strobus L.*). Other tree species include 11% balsam fir (*Abies balsamea L. Mill*) and native Carolinian species, including 4% Oak (*Quercus velutina L.*, *Q. alba L.*), 2% Red Maple (*Acer rubrum L.*) and some wild black cherry trees (*Prunus serotina Ehrh*). The understory consists of ferns (*Pteridium aquilinum L.*), mosses (*Polytrichum spp.*), poison ivy (*Rhus radicans L. ssp.*) and Rubus species. The topography at the site is fairly flat with well drained sandy soil (Brunisolic Gray Brown Luvisol, following the Canadian Soil Classification system), which is composed of ~98% sand, 1% silt, and 1% clay (Peichl et al., 2010a; b).

This forested site is managed by the Ontario Ministry of Natural Resources and Forestry (OMNRF) under the shelterwood silvicultural system (OMNRF, 1998; OMNRF, 2000; Trant, 2014). This system is characterized by two or more partial thinning treatments over a few decades to allow for regeneration and development of seedlings in partial shade. The forest was first thinned in 1983, where approximately one-third of trees were randomly removed. In March of 2012, a second partial thinning was conducted and approximately 30% of the dominant overstory trees were selectively removed to improve light and water availability and stimulate growth of the remaining trees. The long-term aim of the management is to convert the pine plantation back to the native stand with mixed-wood ecosystem characteristics (OMNRF, 1998). Average basal area was reduced by 13% (Kula, 2014), while peak leaf area index was reduced by 35% (Trant, 2014) following shelterwood harvesting. It was estimated that the thinning added 43 t of foliage, 200 t of live branches and 129 t of dead branches to the forest floor, while an additional 515 t of roots were made inactive from growth and production (Trant, 2014).

Water, carbon and energy fluxes were measured using a closed-path eddy-covariance system. Flux measurements were made at 20 Hz above the canopy at on top of a scaffolding tower at a height of 28 m. Half-hourly net ecosystem productivity (NEP) was calculated by adding CO_2 flux and the CO_2 storage, which was calculated as change in CO_2 concentration over time in the air column below the eddy-covariance sensors (Brodeur, 2014). Ecosystem respiration (R_E) was estimated using a non-linear logistic relationship between nocturnal half-hourly CO_2 fluxes during high turbulence conditions and soil temperature at 2 cm depth (Brodeur, 2014; Peichl et al., 2010a; b). Gross ecosystem productivity (GEP) was determined by adding measured NEP to modelled daytime R_E. Soil temperature and soil moisture were measured at 5, 10, 20, 50 and 100 cm depths at two locations, near the chamber system location. For the analysis, soil temperature at a depth of 5 cm and soil moisture content, averaged between 0 and 30 cm depths were used. All flux and meteorological data were quality controlled and averaged at half-hourly intervals (Brodeur, 2014). For more details on eddy-covariance tower instrumentation refer to Arain and Restrepo-Coupe (2005) and Peichl et al. (2010a; b).

Soil CO_2 efflux was measured using automated non-steady state chambers developed by the University of British Columbia (Drewitt et al., 2002; Gaumont-Guay et al., 2008; Jassal et al., 2005; 2012). The soil CO_2 efflux chamber system (chamber system) occupied an area of 50 m × 50 m and was installed approximately 100 m north from the eddy-covariance flux tower at the TPFS. The advantage of automated chamber systems to manual measurements is their ability to record continuous long-term measurements and thousands of measurements annually of soil CO_2 efflux which is important when assessing short- and long-term responses of soil CO_2 efflux (Jassal et al., 2012). The experimental area was chosen to complement the ecosystem carbon flux measurements by the flux tower. The individual chambers were placed in locations within the footprint of the flux tower measurement, with the aim to provide data for the soil CO_2 efflux contribution to the ecosystem respiration. Chambers were placed away from tree stems, thick understory, and placed on level ground, in order to avoid large roots, particularly for those chambers measuring heterotrophic respiration. As soil respiration is comprised of contributions from the mineral-soil, the organic-litter horizon and roots with their associated mycorrhizae, some collars for the chambers were manipulated to help elucidate the contribution of these different components to the total soil respiration.

The chamber system was established to measure total soil CO_2 efflux, including fluxes from root exclusion and forest floor removal plots. It was installed in June 2008, with the initial installation of four chambers: one chamber to measure heterotrophic respiration (F_{RT1}) in the trenched plot whereby live tree roots were severed in the area surrounding the chamber (Nicholas, 2011); and three chambers used as control plots to measured total soil CO_2 efflux (F_S) (F_{S1}, F_{S2} and F_{S22}). One of the control chambers (F_{S22}) was altered in May 2009 by removing the Litter-Fermenting-Humified (LFH) layer and used to measure the contributions from the mineral-soil horizons (F_{LR}) to total soil CO_2 efflux. Additional litter, which landed inside this chamber, was periodically removed, including any material from the harvesting event. In May 2009, two more control chambers were added (F_{S3} and F_{S4}) to the chamber system to measure F_S. In May 2010, two additional chambers were installed in the chamber system, one used as a control (F_{S5}), and the other to measure heterotrophic respiration (F_{RT2}). At this point, eight chambers comprised the chamber system.

1.3 THESIS STRUCTURE

The objectives of this dissertation are addressed within the second, third and fourth chapters as described in the following sections. These chapters are formatted for publication in scientific journals. The paper presented in Chapter 2 is already published in the journal *Forests* (Thorne and Arain, 2015). The manuscript presented in Chapter 3 has been submitted to the journal *Forests*, while the manuscript presented in Chapter 4 has been submitted to the journal *Plant, Soil and Environment*. Study conclusions are provided in Chapter 5. Each chapter is briefly summarized below.

Chapter 2 – Influence of Low Frequency Variability on Climate and Carbon Fluxes in a Temperate Pine Forest in eastern Canada

Carbon, water and energy exchanges between forests and the atmosphere depend on seasonal dynamics of both temperature and precipitation, which are influenced by low frequency climate oscillations such as: El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), Arctic Oscillation (AO), Eastern Pacific Oscillation (EPO) and the Pacific-North American (PNA). This study investigated the influence of climate oscillations on the local climate and carbon fluxes in a mature temperate white pine (*Pinus strobus L.*) forest, near Lake Erie in southern Ontario, Canada. Analyses indicated mean winter temperatures were correlated to NAO, AO and EPO, total winter precipitation was correlated to PNA and AO, while total snowfall was correlated to PNA and ENSO. These climate oscillations influenced carbon dynamics of the forest during the winter and spring seasons. The EPO had a significant inverse correlation with winter and spring carbon fluxes, while the PDO was significantly correlated with winter respiration.

In 2012, a warm event linked to climate oscillations raised temperatures and resulted in a large release of carbon from the forest due to higher ecosystem respiration. As low frequency climate oscillations are important drivers of extreme weather events, affecting their intensity, frequency and spatial patterns, as predicted by climate change models, can cause large changes in carbon exchanges in forest ecosystems in northeastern parts of North America.

Chapter 3 – Response of Soil CO_2 Efflux to Shelterwood Harvesting in a Mature Temperate Pine Forest

In forest ecosystems, soil CO_2 efflux is an important component of ecosystem respiration (R_E), which is generally driven by variability in soil temperature and soil moisture. Tree harvesting in forests can alter the soil variables, and consequently impact soil CO_2 efflux. This study investigated the response of total soil CO_2 efflux, and its components, to a shelterwood harvesting event of a mature temperate white pine (*Pinus strobus L.*) forest, located in southern Ontario, Canada. The objective was to explore the response of soil CO_2 efflux dynamics to changes in the forest microclimate, soil temperature and soil moisture, after the harvesting removed approximately one-third of the overstory canopy. No significant differences were found for soil temperature and soil moisture between pre- (2008 to 2011) and post-harvesting (2012 to 2014) periods. Despite similar soil microclimate, total soil CO_2 effluxes were significantly reduced by up to 37%. Soil CO_2 effluxes from heterotrophic sources were significantly reduced by about 27%, post-harvesting while no significant difference in the soil CO_2 effluxes from mineral-soil horizon was measured between pre- and post-harvesting. An analysis of R_E, measured with an eddy-covariance tower over the study area, showed an increased post-harvesting. However, the overall net ecosystem productivity showed no significant difference between pre- and post-harvesting. This was attributed to an increase in the gross ecosystem productivity (GEP) post-harvesting, compensating for the increased losses (i.e. increased R_E). This study highlights the complexities of assessing soil CO_2 effluxes after a disturbance, such as harvesting. The knowledge gained from this study provides a better understanding of the response of ecosystem carbon exchange to shelterwood harvesting and will be useful for forest managers focused on carbon sequestration and forest conservation.

Chapter 4 – Evaluating the Response of Seasonal and Annual Dynamics of Soil CO_2 Efflux to Environmental Factors, Including Low Frequency Climatic Oscillations and Forest Management, in a Temperate Pine Forest

Climate variability and forest management practices influence the rate of soil CO_2 efflux; however, their combined effects are complex and not well understood. This study investigated the response of soil CO_2 efflux to seasonal and interannual climate variability and shelterwood harvesting in a mature wine pine (*Pinus strobus L.*) forest, near Lake Erie in southern Ontario, Canada. By developing a Gaussian-Gamma specification model, the particular focus was to explore the response of the major components of soil CO_2 efflux, autotrophic and heterotrophic respiration, and soil CO_2 efflux from mineral-soil horizon, to soil temperature, soil moisture, shelterwood harvesting and climate oscillations. Soil CO_2 effluxes were higher when soil temperatures were between 16 and 23°C, and while the soil moisture was between 0.08 and 0.15 m^3/m^3. Soil CO_2 efflux was found to be the major contributor to the total forest carbon emissions, particularly during the winter months, contributing, on average, 78±9% to the annual total ecosystem respiration (R_E) of the forest. It was estimated that the shelterwood harvesting treatment reduced heterotrophic respiration contribution to total soil CO_2 efflux from 80±6% (2008 to 2011) to 53±21% (2012 to 2014), while increasing autotrophic respiration contribution from 20±5% to 47±21% following shelterwood harvesting. The mineral-soil horizon contribution was similar pre- and post-harvesting. By incorporating the climate oscillations that prevail in the region, and their active years into the model, the relationship between soil CO_2 efflux and environmental factors were significantly improved. Results from this study will aid to better understand the influence of shelterwood harvesting and climate variability, affected by local and large-scale climate oscillations on estimates of soil CO_2 effluxes and develop polices for better management focused on carbon sequestration and forest conversation.

Chapter 5 – Conclusions

In all chapters that constitute either a published paper or manuscripts submitted to peer-reviewed journals: I have reviewed the literature; conducted the field-work for soil CO_2 efflux data, meteorological and soil environmental data collection using the suggested methodology, performed data analysis, interpreted results, and wrote the first draft of the manuscripts. Co-authors provided insightful comments and feedback throughout the course of the study and during the write-up of these chapters. Acknowledgements of specific contributions and to collaborators are provided at the end of each chapter.

Due to the nature of this dissertation, comprising either a published paper or manuscripts submitted to peer-reviewed journals, there is some overlap in sections of the dissertation related to study introduction, study site and methodology.

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CHAPTER 2

INFLUENCE OF LOW FREQUENCY VARIABILITY ON CLIMATE AND CARBON FLUXES IN A TEMPERATE PINE FOREST IN EASTERN CANADA

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Abstract: Carbon, water, and energy exchanges between forests and the atmosphere depend on seasonal dynamics of both temperature and precipitation, which are influenced by low frequency climate oscillations such as: El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), Arctic Oscillation (AO), Eastern Pacific Oscillation (EPO) and the Pacific-North American (PNA). This study investigated the influence of climate oscillations on the local climate and carbon fluxes in a 75-year old temperate pine (*Pinus strobus L.*) forest, near Lake Erie in southern Ontario, Canada. Analyses indicated mean winter temperatures were correlated to NAO, AO and EPO, total winter precipitation was influenced by PNA and AO, while total snowfall was correlated with PNA and ENSO. These climate oscillations influenced carbon dynamics of the forest during the winter and spring seasons. The EPO had a significant inverse correlation with winter and spring carbon fluxes, while the PDO was significantly correlated with winter respiration. In 2012, an extreme warm event associated with climate oscillations raised air temperatures and resulted in a large release of carbon from the forest due to higher ecosystem respiration. As low frequency climate oscillations are important drivers of extreme weather events, affecting their intensity, frequency and spatial patterns, can cause large changes in carbon exchanges in forest ecosystems in northeastern parts of North America.

2.1 INTRODUCTION

Forests are considered an important sink for atmospheric carbon dioxide (CO_2) and have the potential for temporarily storing atmospheric CO_2 in terrestrial ecosystems to offset anthropogenic greenhouse gas emissions (Kindermann et al., 2008; Pan et al., 2011). Previous studies have shown forest carbon (C) cycle responds to climate variations (Heimann and Reichstein, 2008; Wharton et al., 2012; Bastos et al., 2013). Therefore, predicted future climate change may have a severe impact on forests ecosystems. While a warming-induced increase in the growing season length (Piao et al.,2008), CO_2 fertilization and higher nitrogen deposition effects are expected to increase C assimilation in the future (Black et al., 2000; Reay et al., 2007), the short-term effects through extreme climatic events such as droughts, flooding, heatwaves, duration and timing of the winter and freeze/thaw periods, as well as disturbance regimes are expected to severely affect forests (Amiro et al., 2010; Davis et al., 1997; Granier et al., 2007; Reichstein et al., 2013; Vose et al., 2012; Wullschleger and Hanson, 2006).

In North America, low frequency climate oscillations, such as the El Niño-Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), the Arctic Oscillation (AO), the Pacific Decadal Oscillation (PDO), the Eastern Pacific Oscillation (EPO) and the Pacific-North American (PNA) teleconnection pattern, are important atmospheric phenomenon that influence seasonal and annual climate variations (Bonsal et al., 2001; Shabbar et al., 1997; Shabbar and Bonsal, 2004; Zhang et al., 2011). Time scale of these oscillations range from interannual to interdecadal, and their influences are typically stronger during the winter months, but can persist into the spring and summer months. Changes in atmospheric circulation associated with these climate oscillations can alter seasonal temperatures and precipitation patterns, and hence modify the forest carbon exchange.

In the Great Lakes region, both the western (ENSO, PDO, EPO and PNA) and eastern (NAO and AO) low frequency oscillations may affect the magnitude and type of winter precipitation (Bai et al., 2012; Wang et al., 2010). Additionally, during the winter months, these climate oscillations can influence the presence or absence of ice cover in Great lakes (Bai et al., 2014; Brown and Duguay, 2010), which affects both regional climate and weather events (Brown and Duguay, 2010). These changes in regional climate may have an impact on the growth and carbon uptake in vegetation ecosystems, in particular forest ecosystems, which are re-establishing in this region as re-growth on abandoned agricultural lands or plantations. These re-emerging forests of North America are a large sink of carbon (Goodale et al., 2002; Woodbury et al., 2007). A summary of impacts of each low frequency climate oscillations in the Great Lakes region, and how they may influence winter temperature and precipitation are summarized in Table 2.1.

ENSO is a coupled ocean-atmosphere interaction that occurs across the equatorial Pacific Ocean (Fleming et al., 2007). It consists of a warm (El Niño) and a cold episode (La Niña), lasting from 6 to 18 months, with neutral years interspersed between the episodes. El Niño (La Niña) episodes often brings less (more) precipitation and warmer (cooler) conditions to a majority of North America (Shabbar et al., 1997; Shabbar and Khandekar, 1996). Higgins et al. (2002) suggested that although La Niña events generally bring cooler temperatures to most of Canada, since the Pacific jet stream is displaced northward; there is a decrease in the frequency of Arctic air intrusions into the Great Lakes region which affects lake ice cover (duration, thickness and extent). In this region, a non-linear relationship exists between ENSO and Great Lakes ice cover as a majority of the maximum ice cover occurs during weak or neutral ENSO episodes, whereas most of minimal ice cover occurs during strong El Niño or La Niña events (Bai et al., 2012). Due to this non-linear response, strong ENSO events (both strong El Niño and La Niña episodes) have generally produced warmer winter temperatures and greater winter precipitation (Hoerling et al., 1997; Rodionov and Assel, 2003; Wu et al., 2005). Assel et al. (2000) observed that lower winter precipitation due to El Niño events generally occurred in the upper Great Lakes region, whereas the lower lakes, such as Lake Erie, observed an opposite effect. LaValle et al. (2000) found that Lake Erie water levels were significantly correlated with ENSO, as El Niño (La Niña) episodes were often associated with higher (lower) water levels, although the response of

Table 0.1. Average climate oscillations influence and associated trends on winter temperature and precipitation in the Great Lakes region. NAO = North Atlantic Oscillation, AO = Arctic Oscillation, SO = Southern Oscillation, PDO = Pacific Decadal Oscillation, PNA = Pacific-North American and EPO = Eastern Pacific Oscillation.

Typical Climate Oscillation Influence	**Predicted Temperature Trend**	**Predicted Precipitation Trend**
NAO Positive	Warmer	Wetter
NAO Negative	Cooler	Drier
NAO Neutral	Weak Influence	Weak Influence
AO Positive	Warmer	Wetter
AO Negative	Cooler	Drier
AO Neutral	Weak Influence	Weak Influence
Strong El Niño (SO)	Warmer	Drier
Strong La Niña (SO)	Warmer	Wetter
Weak-Moderate Events	Weak Influence	Weak Influence
PNA Positive	Cooler	Drier
PNA Negative	Warmer	Wetter
PNA Neutral	Weak Influence	Weak Influence
PDO Positive	Cooler	Drier
PDO Negative	Warmer	Wetter
PDO Neutral	Weak Influence	Weak Influence
EPO Positive	Cooler	n/a
EPO Negative	Warmer	n/a

lake levels to precipitation can take anywhere from three months to two years (Meadows et al., 1997).

The PDO is a low frequency climate oscillation which characterizes the interannual variability in the average North Pacific sea-surface temperature (Mantua et al., 1997). Warm (cold) phases exist with the PDO, often associated with below- (above-) average precipitation and warmer (colder) temperatures in northwestern North America, but without neutral phases (Mantua et al., 1997). The PDO has a much longer cycle than ENSO, persisting for 2 to 3 decades, while a typical ENSO event can persist for 6 to 18 months (Mantua and Hare, 2002), but can influence ENSO events by enhancing or suppressing the strength of their signals. Gershunov and Barnett (1998) found that El Niño (La Niña) signals were stronger and stable during positive (negative) PDO events. Over the Great Lakes region, positive (negative) phases on the PDO usually produce cooler (warmer) and drier (wetter) conditions.

The PNA pattern is characterized by a trough in the east-central North Pacific, a ridge over the Rocky Mountains, and a trough over eastern North America (Leathers et al., 1991). It is a dominant mode of atmospheric variability over North America and strongly influenced by the ENSO phase (positive phase associated with an El Niño episode), and has been observed at time scales ranging from days to decades (Abeysirigunawardena et al., 2009). Angel and Isard (1998) indicated that during the majority of the winter season, lower cyclone frequency is associated with negative PNA events, while higher cyclone frequency is associated with positive PNA events. As with the PDO, positive (negative) phases of the PNA produce cooler (warmer) and drier (wetter) conditions in the Great Lakes region.

The EPO has centers of action at 500 hPa height fields (Barnston and Livezey, 1987; Quiring and Goodrich, 2008). The positive phase is represented by a ridge east of Hawaii and a trough in the Gulf of Alaska, with a secondary ridge near Hudson Bay (Mioduszewski et al., 2014). During a positive EPO phase, the Great Lakes region tends to have cooler conditions, while the negative phase is associated with warmer temperatures (National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center).

The NAO is associated with changes in the surface westerly winds across the Atlantic and into Europe. It is a meridional oscillation in atmospheric pressure with centers of action near the Icelandic Low and the Azores High (Hurrell and Van Loon, 1997). The AO is the leading empirical orthogonal function of wintertime sea-level pressure anomalies over the extratropical Northern Hemisphere. It resembles the NAO in many respects, but covers more of the Arctic (Thompson and Wallace, 1998). A positive phase of the NAO and AO generates warmer and wetter conditions, while a negative phase is expected to make the Great Lakes region generally cooler and drier.

Previous studies have shown that low frequency climate oscillations can have a strong influence on extreme climatic events (Kenyon and Hegerl, 2008; Kenyon and Hegerl, 2010; Ning and Bradley, 2014a;b). Summer and winter climate extremes have been linked to climate oscillations. Loikith and Broccoli (2014) found that extreme warm and cold temperatures were strongest between the PNA and AO in regions most influenced by these teleconnections, with a weak association with ENSO events. For example, during negative AO events, there is an increase in the probability of extreme cold minimum temperatures over the northeastern part of North America. Summer associations were weaker than wintertime associations due to influences on the local surface energy budget and circulation patterns are smaller in scale and weaker in magnitude than in the winter. Ning and Bradley (2014a; b) found similar results for winter temperature and precipitation extremes; however, they noted maximum and minimum values were less sensitive to climate variability compared to mean values. Current sea-ice conditions also influence the NAO and AO, which may lead to changes in the weather patterns around the Great Lakes region, and increased probability of extreme climatic events (Francis et al., 2009; Francis and Vavrus, 2012; Tang et al., 2013).

Several studies have attempted to identify links between ENSO and NAO (Brown, 2010; Giannini et al., 2001; Huang et al., 1998; Rogers, 1984). Both the ENSO and NAO create a combination of influences by increasing or decreasing temperatures and/or the amount of precipitation, depending on the region.

Interference of these low frequency climate oscillations may have profound influence on the climate and forest ecosystems in northeastern North America and Great Lakes regions, particularly during the winter. The simultaneous occurrence of two or more climate oscillations can exaggerate impacts of seasonal and annual climate variability on carbon, water and energy fluxes in the forest ecosystems in the region. Extreme events, such as heatwaves, droughts or flooding, can partially offset carbon sinks or even cause net losses of carbon in forest ecosystems. Forest ecosystems, although potentially susceptible to all types of extreme events, are particularly vulnerable to drought (Reichstein et al., 2013), where the effects on the carbon balance can be both immediate and lagged, and persistent. Since large-scale modes of climate variability are important drivers of changes in climate extremes (Kenyon and Hegerl, 2008), variations to these modes will affect the intensity, frequency, and spatial patterns of extreme climatic events, leading to changes in carbon fluxes in forest ecosystems (Zcheischler et al., 2014).

In this study, the occurrence and impacts of several low frequency climate oscillations in the Great Lakes region in Canada over the past seven decades (1950 to2014) was investigated. The objectives are to: (i) determine correlation, if any, between each low frequency climate oscillation and temperature, precipitation and snowfall in the Great Lakes region from 1950 to 2014; and (ii) quantify changes in gross ecosystem productivity (GEP), ecosystem respiration (R_E) and net ecosystem productivity (NEP) due to low frequency climate oscillations in a temperate pine forest, using measured eddy-covariance flux data from 2003 to 2014.

2.2 STUDY AREA

The study area is a 75-year old eastern white pine forest near Turkey Point Provincial Park on the northern shore of Lake Erie in southern Ontario, Canada (Figure 2.1). This site is part of the Turkey Point Flux station (TPFS), (also known as the Turkey Point Observatory), of the Global Water Futures (GWF) and global Fluxnet initiatives. The topography at the site is fairly flat with well drained sandy soil (Brunisolic Gray Brown Luvisol, following the Canadian Soil Classification system), which is composed of ~98% sand, 1% silt and 1% clay (Peichl et al. (2010a; b).

The forest was planted in 1939 on cleared oak-savanna lands to stabilize soils, which has led to a nearly homogeneous canopy height and structure (Arain and Restrepo-Coupe, 2005). Tree composition is 82% white pine (*Pinus strobus L.*), 11% balsam fir (*Abies balsamea L. Mill*) and native Carolinian species including 4% Oak (*Quercus velutina L.*, *Q. alba L.*) and 2% Red Maple (*Acer rubrum L.*). The average tree height is 21.8±1.7 m, and stand density is about 421±166 stems ha^{-1}. In the winter of 2012, the forest was selectively thinned and approximately 30% of the trees were removed to improve light and water availability, and stimulate growth of the remaining trees. Further site details are described in Arain and Restrepo-Coupe (2005) and Peichl et al. (2010a; b). The climate in the

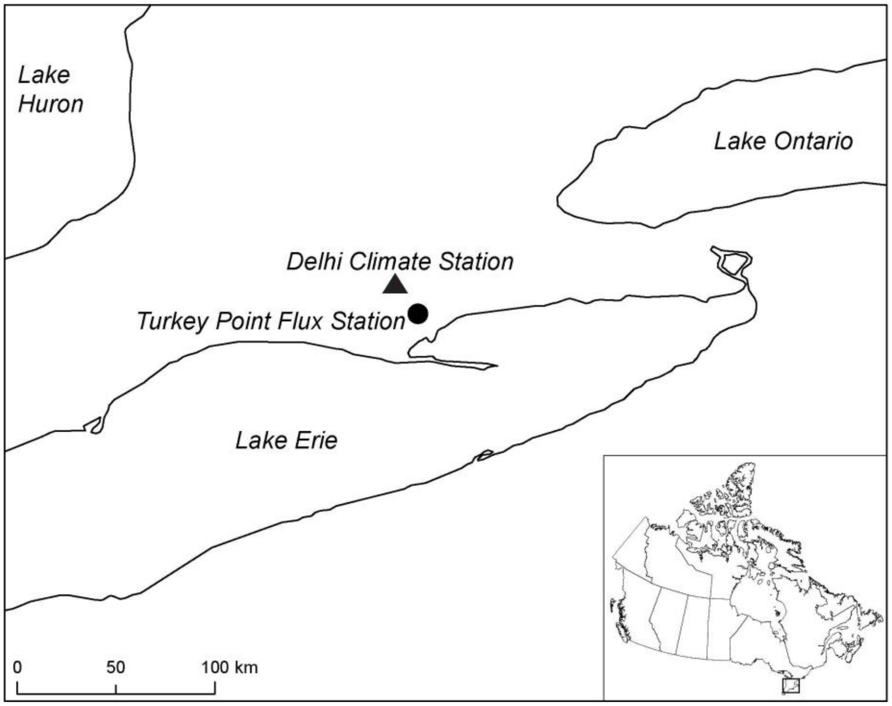

Figure 1. Location of the Turkey Point Flux station on the north shore of Lake Erie in southern Ontario, Canada and the climate station at Delhi (climate ID: 6131983), maintained by Environment Canada.

region is cool-temperate with a 30-year mean annual temperature of 8.0°C and mean annual precipitation of 1036 mm with 83% falling as rain (based on 1935 to 2013 Environment Canada weather data for Delhi, Ontario (climate ID: 6131983), located about 25 km northwest from forest).

2.3 DATA AND METHODS

Water, carbon and energy fluxes were measured using a closed-path eddy-covariance system, comprising of a sonic anemometer (CSAT3, Campbell Scientific Inc. (CSI)) and infrared gas analyzer (Li-7000, LI-COR Inc.). Flux measurements were made at 20 Hz above the canopy at approximately on top of a scaffolding tower at a height of approximately 28 m. Precipitation was measured in an open area about 2 km northeast from the site, using a weighed accumulation rain gauge (T200B, Geonor Inc.) and a complimentary tipping bucket rain gauge (TE525, Texas Inst.). All flux and meteorological data were quality controlled and averaged at half-hourly intervals. Further details of closed-path eddy-covariance systems and meteorological and soil measurements are described in Arain and Restrepo-Coupe (2005) and Peichl et al. (2010a; b).

To investigate the possible linkages of regional climate with large-scale atmospheric variability, Pearson correlations were conducted between the various climate oscillation indices described previously and environmental variables (mean seasonal and annual temperature, seasonal and annual total precipitation and total winter snowfall) from the climate station in Delhi, Ontario (climate ID:

6131983) from 1950 to 2014. These climate oscillation indices are created based on the differences between the relative controlling variables for each climate oscillation (e.g. sea-surface temperature, air pressure) and were obtained from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (NAO index and PNA index), the Bureau of Meteorology (ENSO, as the Southern Oscillation Index (SOI)), and the Joint Institute for the Study of the Atmosphere and Ocean (PDO index and EPO index). For all analyses, only winter index values (averaged between November to March) were used as these climate signals are notably stronger during the winter. A linear correlation was still applied when using the SOI to maintain consistent methodology with the other teleconnections. Carbon fluxes, local climate and climate oscillations from 2003 to 2014 were used in this analysis. Seasons were classified as winter (November to March), spring (April and May), summer (June to August) and autumn (September and October).

For this study, the strength of a climate oscillation were defined as follows: positive (negative) moderate to strong phases/events were defined as those years which were in the upper (lower) quartile of the data; and neutral phases/events as those years found in the mid-quartile range. A short-term analysis, from 2003 to 2014, was conducted to determine the strengths and phases/events of each climate oscillation for each year, and then their effects on climate and carbon fluxes.

2.4 RESULTS AND DISCUSSION

The relationship between climate oscillations and environmental variables from the long-term Delhi climate station (climate ID: 6131983) data was first analyzed, followed by a breakdown of the strengths of each climate oscillation, for the years 2003 to 2014, and their potential impact on the local climate and carbon fluxes at the forest site.

2.4.1 Long-term analysis

In the Great Lakes region, previous studies have shown that both eastern and western climate oscillations affect the local climate and these influences are stronger during the winter (Bai et al., 2012; Bai et al., 2014; Wang et al., 2010). Annually, four possible conditions can occur depending on the strength of the various climate oscillations over the winter: (1) one strong oscillation occurring overpowering all other oscillations; (2) two or more strong oscillations; (3) strong oscillation(s) combined with regional effects (such as lake effect); and (4) weak to neutral phases of these oscillations, where regional effects are more noticeable. When only one oscillation is strong, its phase will generally be the most influential on the local climate. When two or more oscillations are strong, both may influence the local climate at the same time or they may be out of phase in terms of their effect on the region, with the possibility of their influences being nullified by each other. For example, Bai et al. (2011) noted that during the 2009/2010 winter over the Great Lakes region, a negative AO, which generally brings cooler temperatures, coincided with a strong El Niño event that typically is associated with warmer temperatures. The El Niño event dominated the mid-latitude region and prevented cold Arctic air masses from intruding from the north, leading to a warmer winter. Additionally, with the combination of the PNA, PDO and ENSO oscillations, which are highly correlated with each other, Ning and Bradley (2014b) showed that strong PNA and PDO events, those years with and without ENSO forcing, have a different impact on moisture flux patterns over northeastern United States.

This study showed a positive correlation between mean winter temperature and the NAO and AO indices, and a negative correlation with the EPO index (Table 2.2). These significant correlations confirm that positive (negative) phases of the NAO and AO indices can increase (decrease) winter temperatures in the region, where the opposite is true for the EPO index. The PNA and AO indices both had a significant correlation with total winter precipitation, where a negative (positive) phase of the PNA index and a positive (negative) phase of the AO index brought more (less) winter precipitation to the area. The EPO index has a significant positive correlation with autumn precipitation, increasing precipitation during positive phases.

Only the PNA and SOI indices showed a significant correlation with total winter snowfall. Negative (positive) phases of the PNA index generally produced more snowfall and El Niño (La Niña) events decrease (increase) total snowfall amounts. Correlations between the PNA index and winter precipitation is

Table 0.2. Correlation between climate oscillations and mean seasonal temperature, seasonal precipitation and total winter snowfall, from 1950 to 2014. * and ** indicate significance at 95% and 99%, respectively. PDO = Pacific Decadal Oscillation, PNA = Pacific-North American, ENSO = El Niño Southern Oscillation, NAO = North Atlantic Oscillation, AO = Arctic Oscillation, and EPO = Eastern Pacific Oscillation. Winter (November to March), spring (April and May), summer (June to August) and autumn (September and October).

Mean Seasonal Temperature						
Season	**PDO**	**PNA**	**ENSO**	**NAO**	**AO**	**EPO**
Winter	-0.23	-0.02	-0.12	0.33**	0.29*	-0.48**
Spring	-0.18	0.15	-0.05	-0.13	-0.23	-0.13
Summer	0.08	0.12	-0.14	0.19	0.10	-0.19
Autumn	0.03	0.12	-0.09	-0.07	-0.06	-0.15
Total Seasonal Precipitation						
Winter	-0.23	-0.37**	0.09	0.19	0.29*	-0.13
Spring	-0.08	-0.10	0.08	-0.09	-0.12	0.15
Summer	0.02	0.06	-0.10	0.20	0.14	-0.13
Autumn	0.22	0.17	-0.03	0.08	-0.01	0.30*
Total Winter Snowfall						
Winter	-0.15	-0.34**	0.33**	-0.03	0.08	0.02

consistent to those findings by Coleman and Rogers (2003) for the Ohio River Valley. No significant correlations were found between the oscillation indices and spring, summer and autumn temperature, and spring and summer precipitation (Table 2.2).

2.4.2 Occurrence, strengths and impacts of climate oscillations

Mean winter time-series for each climate oscillation from 1950 to 2014 is shown in Figure 2.2. ENSO events have fluctuated over the last twelve years between El Niño and La Niña episodes, with four strong events (one El Niño and three La Niña events) occurring between 2007 and 2011. The PDO was strongly positive in 2002/03 and 2003/04, and strongly negative in 2008/09, 2010/11 and 2011/12. The PNA was also strongly positive in 2002/03 and in 2009/10. Strong negative phases of the PNA occurred in 2008/09 and 2011/12. A strong positive EPO event occurred in 2002/03 and 2013/14, with strong negative events in the winter of 2007/08, 2009/10 and 2011/12. The NAO was strongly positive in 2003/04, 2006/07, 2007/08, 2011/12 and 2013/14, with two strong negative events in 2009/10 and 2010/11. The AO had strong positive events in 2006/07, 2007/08, 2011/12 and 2013/14, with strong negative events in 2009/10 and 2012/13. No strong events were observed during the 2005/06 winter for any of the climate oscillation indices.

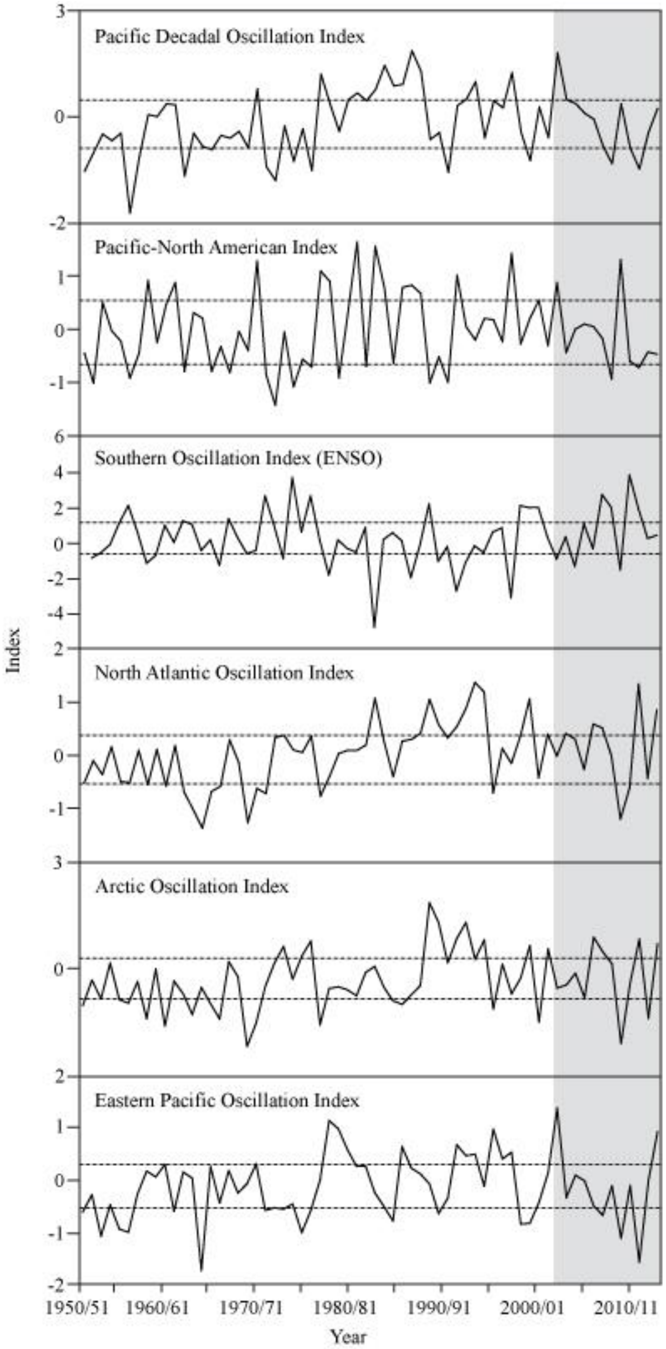

Figure 2.2. Time-series of mean winter index values for each climate oscillation from 1950 to 2014. Grey shading denotes the last twelve years (2002/03 to 2013/14) of the time-series used for comparison with observed carbon flux data using the eddy-covariance technique. Dashed lines for each oscillation represent the upper and lower quartile thresholds used to determine moderate to strong phases/events.

As previously discussed, the strength of an oscillation determines how influential it may be in the Great Lakes region. Each oscillation has its own level of strength, and the occurrence of more than one strong oscillation can complicate how the local climate responds (e.g. Bai et al., 2012; Bai et al., 2014). As the correlation between climate indices and seasonal variables (temperature and precipitation) show a stronger connection to the winter, and not the spring, summer and autumn, this section will only focus on winter climate response to climate oscillations.

Deviation of the mean winter temperature, total winter precipitation and total winter snowfall, for years between 2002/03 and 2013/14 from the long-term mean values (1935 to 2014) are shown in Figure 2.3. Impacts of these oscillations were considered using the generalizations outlined in Table 2.1. In 2002/03, there were strong positive events of the PDO, PNA and EPO and an El Niño event; the first three indices generally produce cooler and drier conditions, whereas the El Niño event typically produces warmer and drier conditions. It is most likely that the cooler conditions from the PDO, PNA and EPO should overpower warmer temperatures typically brought in by an El Niño event. Climate data from the Delhi station for 2002/03 showed that most likely, cooler conditions from the PDO, PNA and EPO overpowered warmer temperatures typically brought in by an El Niño event. In 2002/2003, winter temperatures were cooler than the mean (-1.4°C), and less precipitation and snowfall (-191 mm and -69 mm, respectively)

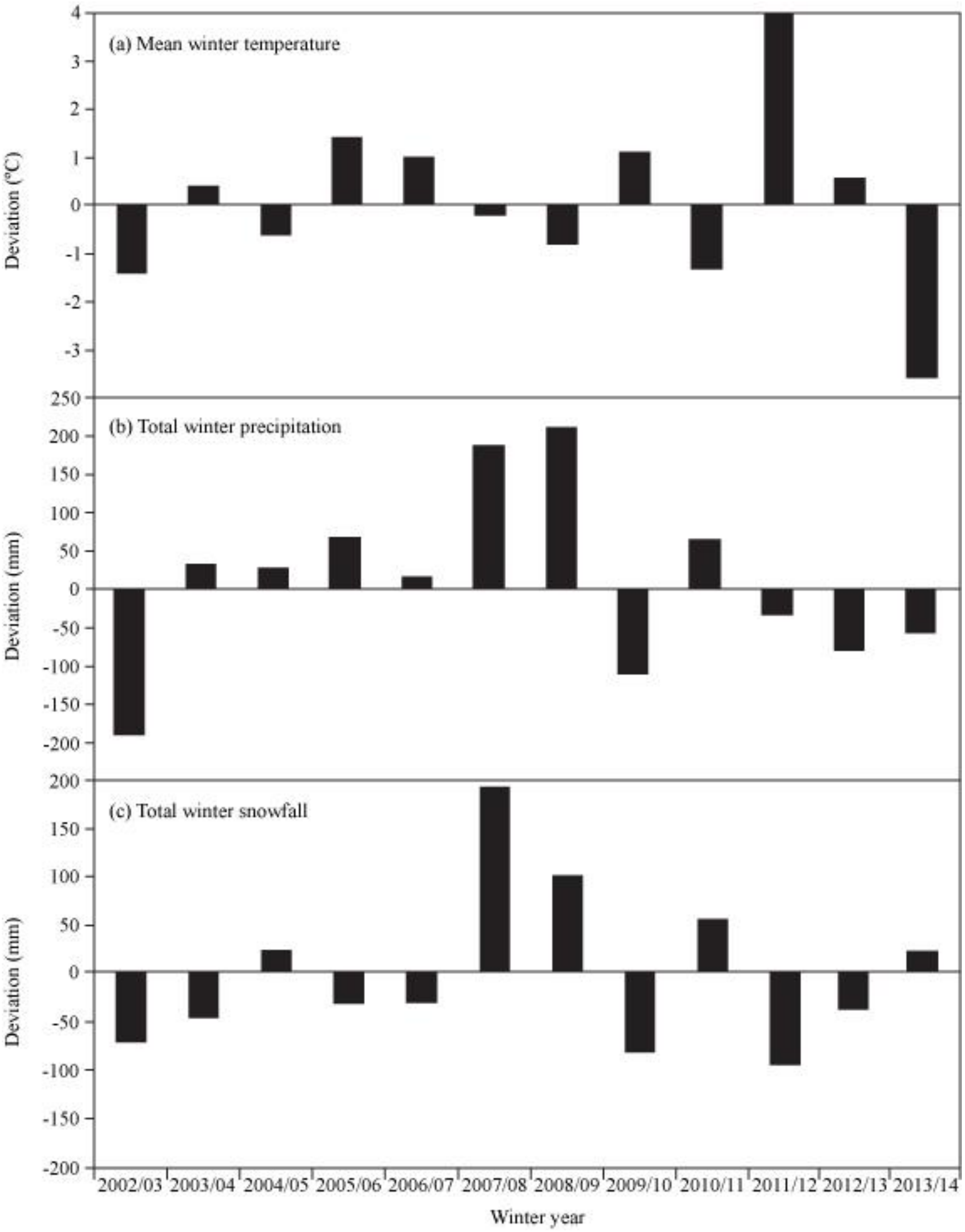

Figure 2.3. Deviations of 2002/03 to 2013/14 data from the long-term (1935 to 2014) winter (a) mean temperature, (b) total precipitation, and (c) total snowfall.

were measured (Figure 2.3). In 2003/04, there was a strong positive PDO and NAO event, with a positive PDO event producing cooler and drier conditions, whereas a positive NAO event creates warmer and wetter conditions. These conflicting events could result in one overpowering the other, or a mix of the two, neutralizing both their influences. Winter temperature and precipitation were slightly above the mean (0.4°C and +30 mm, respectively), with total snowfall lower than normal (-44 mm). This year, the positive NAO event likely had a stronger influence on the local climate. For 2004/05, only a strong El Niño event occurred, however, winter temperatures were cooler (-0.6°C), and winter precipitation and total snowfall were higher (+26 mm for both). Here, the El Niño event did not seem to have much influence on the local climate. In 2005/06, no strong oscillations were present, therefore, regional climate effects or possibly weak to moderate oscillation events could have been influencing factors. Additionally, winter temperatures were slightly warmer (1.4°C), whereas winter precipitation was higher (+66 mm) and total snowfall (-29 mm) lower. Average to warmer temperatures in 2005/06 was found by Bai et al. (2012), as ice cover over the Great Lakes was below-normal. In 2006/07, both the NAO and AO indices had strong positive events, which should bring warmer and wetter conditions to the area. Winter temperatures were warmer (1.0°C), with more precipitation (+13 mm) and less total snowfall (-28 mm). In 2007/08, a strong La Niña event, along with positive phases of the

NAO and AO, and a negative phase of the EPO were present. All events combined would generally be expected to produce warmer and wetter winter conditions in the region. When examining the local climate during the 2007/08 winter, temperatures were relatively close to their mean values (lower by -0.2°C), while total winter precipitation and snowfall were much higher (+186 mm and +195 mm, respectively). Ice thickness for the Great Lakes was found to be about normal (Bai et al., 2012). For 2008/09, another strong La Niña event occurred, along with strong negative phases of the PNA and PDO. The same general response of a warmer and wetter season would be expected, but as with 2007/08, this season had slightly cooler temperatures (-0.8°C) and was much wetter than normal (+210 mm total precipitation and +103 mm snowfall). Wang et al. (2010) found that the positive phase of the AO (which typically warms the region) behaved in an anomalous manner this year. In mid-winter, the Icelandic Low deepened and split, with one section moving to the Labrador Sea. This movement brought cold, dry Arctic air into the Great Lakes region for both December and January. In late-winter, the AO phase shifted back to a negative position. This was the reason why the average index value for this year did not show a strong event occurring.

For 2009/10, several strong oscillations occurred, including an El Niño event, positive PNA phase, and negative EPO, NAO and AO phases. Based on the general outcome predictions, there was a conflict as to how the local climate would be expected to respond. All phases of these oscillations tend towards lower amounts of precipitation in the Great Lakes region. Discrepancies arise in the expected winter temperature response as a positive PNA, and negative NAO and AO phases. dominating the area would tend towards cooler temperatures, while the occurrence of the negative EPO phase and El Niño event would have brought warmer temperatures. During 2009/10, drier conditions were evident, as both precipitation and snowfall were lower than normal (-110 mm and -79 mm, respectively). The seasonal temperatures were warmer than normal (+1.1°C), suggesting that the El Niño event or the negative EPO phase, or a combination of both, was likely the stronger influence in the region that season. Seager et al. (2010) determined that the strong El Niño event pushed the storm track further south, making warmer and drier conditions prevalent in the region. Without this strong ENSO event, the negative NAO would have produced colder temperatures.

In 2010/11, a strong La Niña event and a strong negative PDO phase, which should produce warmer and wetter conditions was observed. A strong negative NAO phase was also present which can cool and dry the region. Temperatures for that season were below-normal (-1.3°C) with an observed increase in precipitation and snowfall (+63 mm and +57 mm, respectively), producing similar colder and wetter conditions as observed in 2007/08. For 2010/11, perhaps the NAO had a stronger influence over the seasonal temperature than the La Niña and PDO events.

In 2011/12, strong negative phases of the PNA, PDO and EPO were present, along with positive phases of both the NAO and AO, and a strong La Niña event. The overall pattern prohibited movement of Arctic air mass into the mid-latitudes and enhanced southerly flow and warm advection from the Gulf of Mexico into the Great Lakes region. Similar observations were made by Bai et al. (2014) for this region. This led to record warmer and wetter seasonal values (4°C above the mean seasonal winter temperatures). In March 2012, an extreme warm event occurred in central and eastern North America for the majority of the month, which contributed to the above-normal mean winter temperatures. Tropical heating anomalies over the Indian Ocean and western Pacific, due to an exceptionally strong Madden-Julian Oscillation (an oscillation partly linked to ENSO), contributed to the flow anomalies that were the cause for the extreme warm temperatures (Dole et al., 2014). According to temperature data recorded at both the Delhi climate station and the local weather station, temperatures began to climb on March 11 and continued for fifteen consecutive days, peaking at 27°C and 25°C above the monthly normal at the Delhi climate station and local weather station, respectively. There was less winter precipitation (-34 mm), and less than normal total snowfall (-11 mm), as the majority of precipitation fell as rainfall due to the warmer winter temperatures. This is a contrast to expected general conditions based on the combination of strong oscillations for the region.

In 2012/13, a strong negative phase of the AO was observed, which typically allows colder, dry Arctic air into the region. Winter temperatures were slightly warmer than the mean (+0.5°C), although much cooler than the previous winter. Both precipitation and snowfall were below the mean, -80 mm and -42 mm, respectively. In 2013/14, strong positive phases of the NAO, AO, and EPO were observed. Both the NAO and AO typically bring in warmer and wetter conditions, however, the EPO is generally associated with cooler conditions. Mean winter temperature was considerably below-normal (-3.5°C), with less precipitation (-58 mm) but greater total snowfall (+19 mm). During this winter, cold temperatures were observed across much of North America due to anomalous meridional upper air flow, also referred to as a polar vortex (Clites et al., 2014).

Our analysis suggests that discrepancies between the predictions and outcomes are related to abrupt changes in climate patterns or one pattern becoming more dominant over another. Additionally, the regional climate of the Great Lakes can overshadow the influence of low frequency oscillations in some years on the local climate (such as thermal moderation and lake effect snow), which may add a large amount of uncertainty (Brown and Duguay, 2010).

2.4.3 Impacts of teleconnections on carbon fluxes

The impacts of low frequency climate oscillations on GEP, R_E and NEP using observed eddy-covariance flux data was evaluated from 2003 to 2014. Although the forested site was thinned in March 2012, shelterwood harvesting did not significantly impact the response of carbon fluxes and post-harvesting fluxes were within the range of interannual variability (Trant, 2014). Trant (2014) showed that the mean annual post-harvesting (2012 to 2014) GEP over the 2003 to 2014 period was 1518±78 g of C m^{-2} $year^{-1}$ as compared to pre-harvesting (2003 to 2011) GEP of 1384±121 g of C m^{-2} $year^{-1}$. Over the same period, mean post-harvesting NEP was 185±75 g of C m^{-2} $year^{-1}$ as compared to mean pre-harvesting NEP of 180±70 g of C m^{-2} $year^{-1}$, indicating that mean annual pre-harvesting NEP was not significantly different than post-harvesting NEP. Only mean post-harvesting annual R_E (1322±54 g of C m^{-2} $year^{-1}$) was higher, but not statistically different than mean pre-harvesting R_E (1195±101 g of C m^{-2} $year^{-1}$). Detailed analysis of shelterwood harvesting on carbon fluxes at this forested site has been reported by Trant (2014).

Carbon fluxes are predominantly influenced by temperature and precipitation at forest sites, including the one used in this study (Arain and Restrepo-Coupe, 2005; Khomik et al., 2010; Peichl et al., 2010a; b). Seasonal and annual temperature and precipitation correlations against GEP, R_E and NEP are summarized in Table 2.3. In this analysis, a correlation with winter GEP was not included as little photosynthesis generally occurs during the winter (i.e. November to March). Our analysis found that winter temperatures had significant positive correlations with winter R_E and NEP values; while only GEP and R_E had a significant positive response to spring temperatures. There were no significant correlations found between the carbon fluxes (GEP, R_E and NEP) and mean summer and autumn temperatures, and no significant correlations were found against seasonal total precipitation (Table 2.3). Annual NEP was found to be correlated with annual total precipitation, although no significant relationships were found between annual NEP and total annual rainfall or snowfall (data not shown). Winter NEP is inversely correlated with total snowfall. These results show that that during the winter and spring seasons, carbon fluxes were sensitive to changes in the local temperature and precipitation, similar to the results found by Zhang et al. (2011).

With significant correlations found in winter and spring, deviations from the mean winter and spring carbon fluxes from 2003 to 2014 are shown in Figure 2.4. In winter and spring, with a decrease in precipitation and an increase in temperature, carbon fluxes generally became positive and increased in magnitude. Cold temperatures and lower precipitation resulted in the decrease of fluxes in spring of 2005, whereas the opposite occurred in 2010.

Table 0.3. Correlation of gross ecosystem productivity (GEP), ecosystem respiration (R_E) and net ecosystem productivity (NEP) against seasonal and annual mean temperature and mean total precipitation and snowfall from 2002/2003 to 2013/2014. * and ** indicates significance at 95% and 99% , respectively. n/a denotes analysis not conducted. Winter (November to March), spring (April and May), summer (June to August) and autumn (September and October).

Mean Temperature			
Season	**GEP**	**R_E**	**NEP**
Winter	n/a	0.63*	0.67*
Spring	0.86**	0.88**	0.25
Summer	-0.22	0.09	-0.45
Autumn	0.17	0.15	0.03
Annual	-0.21	0.04	-0.42
Mean Total Precipitation			
Winter	n/a	-0.21	-0.58
Spring	-0.16	-0.10	-0.12
Summer	0.23	0.33	-0.12
Autumn	-0.33	-0.30	-0.07
Annual	0.13	-0.27	0.62*
Mean Total Snowfall			
Winter	n/a	-0.32	-0.66*

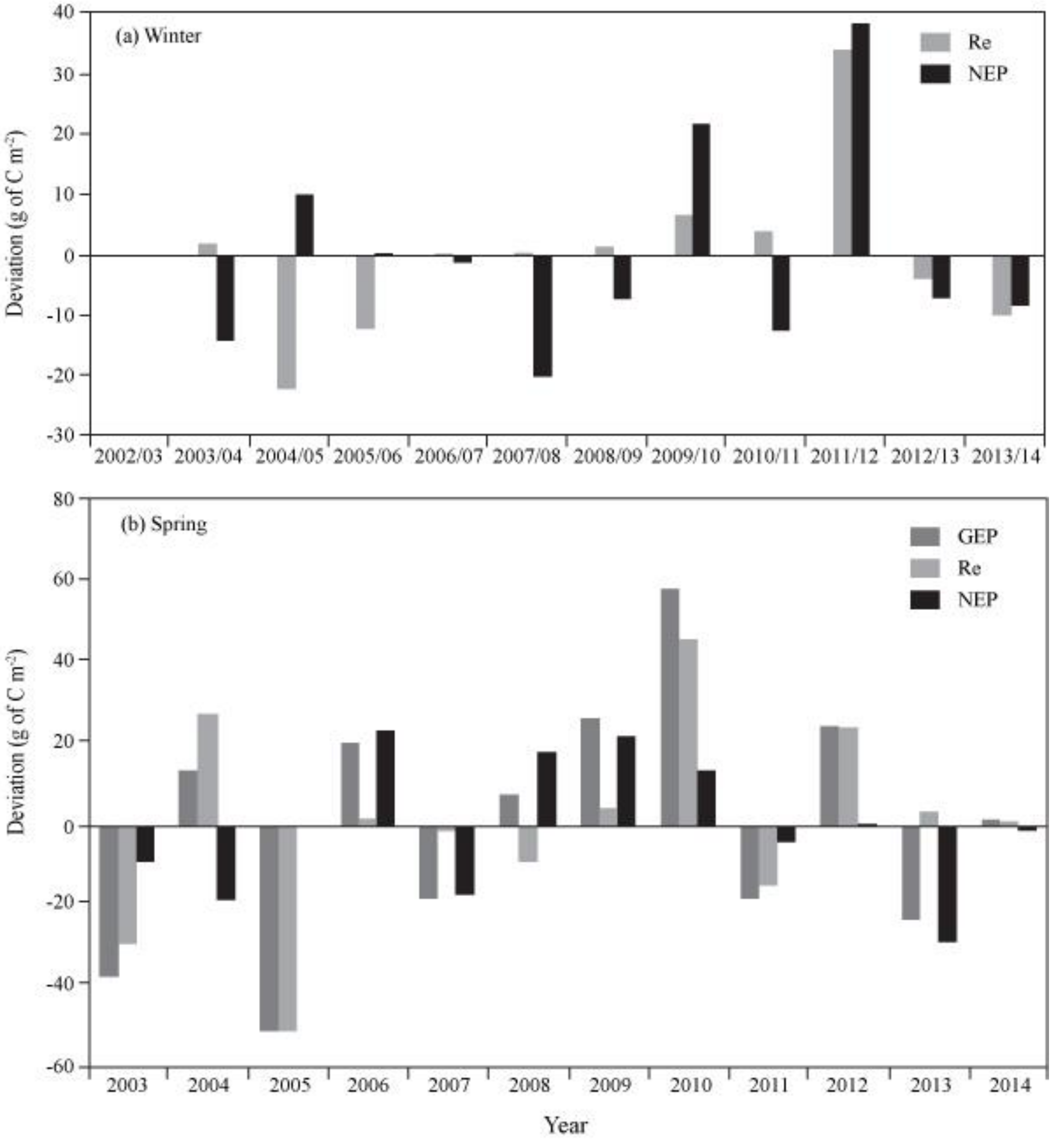

Figure 2.4. (a) Winter and (b) spring deviations from the mean gross ecosystem productivity (GEP), ecosystem respiration (R_E) and net ecosystem productivity (NEP) from 2003 to 2014.

For the 2011/12 winter season, the large increase in seasonal temperature (Figure 2.3) had a strong effect on R_E and NEP. These above-normal winter temperatures increased R_E (+34 g of C) and NEP (+39 g of C) over the annual means. The extreme warm event in March 2012, linked to strong climate oscillations, contributed to the seasonal increase in carbon fluxes. Cumulative R_E and daily mean temperature in March, measured at 2 m above the surface and below the forest canopy at the local weather station, for select years, are shown in Figure 2.5. More R_E occurs during March 2012 than other years (Figure 2.5a) even prior to the warming event which began around March 11 (Figure 2.5b). At the onset of the warming event, cumulative R_E steadily increased coinciding with the rise in temperature and this event alone released approximately 60 g of C from the forest from March 11 to March 30. As daily mean temperatures returned to the seasonal average, carbon re-release diminished.

As influences from climate oscillations are strongest during the winter months, snow storage and melt will be affected. Increases in the snow pack, seen in 2007/08 and 2008/09 winters (Figure 2.3), will most likely delay the onset of the growing season prolonging the melting period and increasing soil moisture. With the 2011/12 winter season, mild winters can enhance carbon uptake as the growing season length increases and more precipitation occurs as rain than snow, which immediately increases soil moisture. Additionally, increased winter precipitation may recharge soil water reserves and reduce the likelihood of water stress or drought occurring later in the summer (Malhi et al., 1999). The analyses showed that only the EPO index has a significant negative relationship with spring values of GEP (Table 2.4). The R_E has a significant correlation with both EPO and PDO during the winter, while only a significant negative correlation with EPO in the spring. NEP only has a significant negative correlation with EPO during the winter. This relationship in the winter and spring with the EPO is likely the result of the 2011/12 winter season, which had above-normal winter temperatures and an extreme warming event in March, which coincides with a strong negative phase of the EPO that significantly increased these fluxes (Figure 2.4 and 2.5a). The significant correlation between the PDO and winter respiration may be a result of respiration increasing (decreasing) in response to a decrease (increase) in the PDO phase. For the other seasons (spring, summer and autumn) and annually, both the EPO and PDO generally show a negative relationship with all three fluxes (GEP, R_E and NEP) although not significant, where a negative (positive) phase would promote (undermine) these carbon fluxes. Mixed weak relationships were found between the carbon fluxes and other indices (Table 2.4).

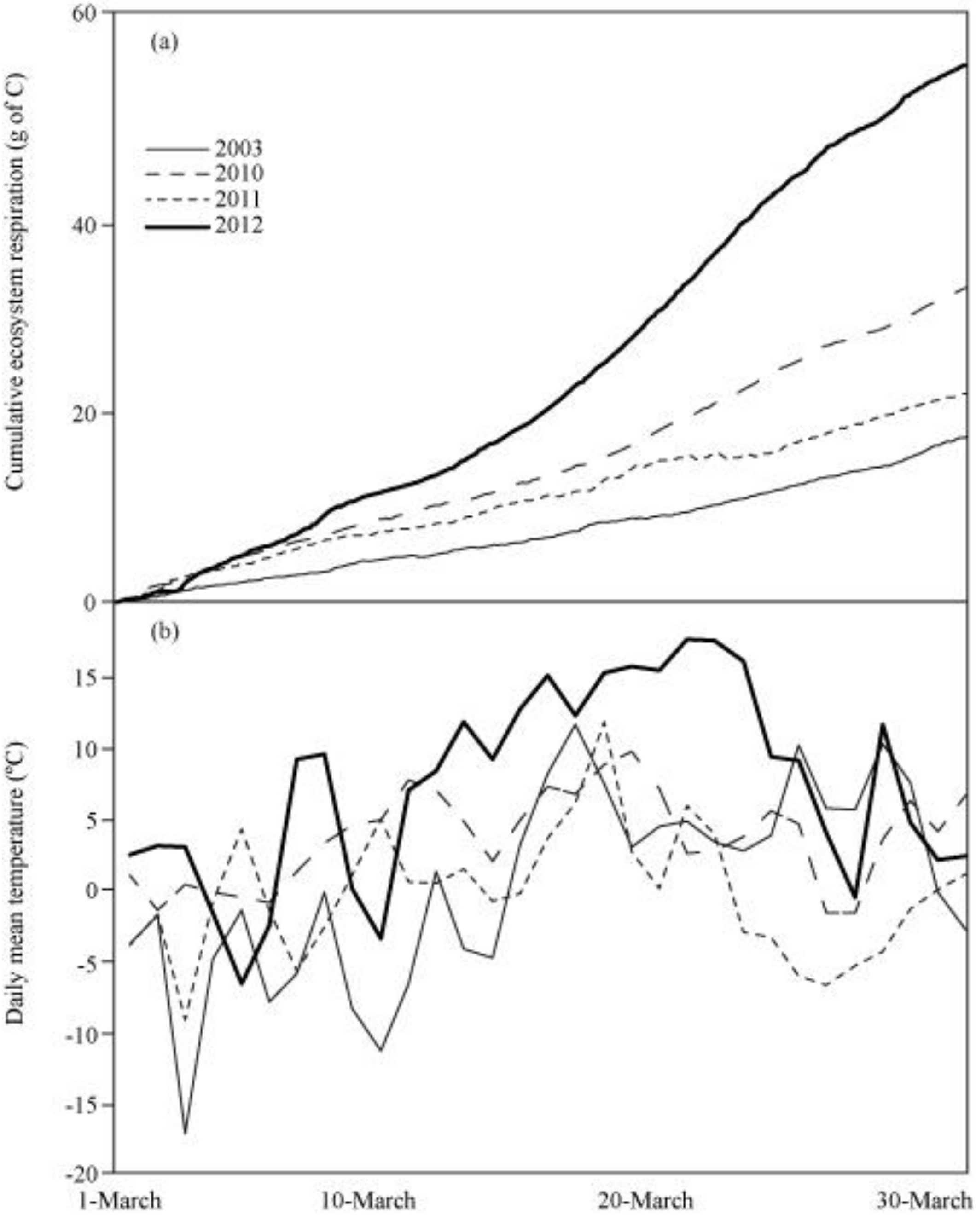

Figure 2.5. March (a) cumulative ecosystem respiration (R_E) and (b) daily mean temperature measured at 2 m above the surface and below the forest canopy for 2003, 2010, 2011 and 2012 at the local weather station. The 2012 data highlighted by the thicker black line.

Table 0.4. Correlation of seasonal and annual gross ecosystem productivity (GEP), ecosystem respiration (R_E) and net ecosystem productivity (NEP) against winter (November to March) mean climate oscillation indices, from 2003 to 2014. * and ** indicates significance at 95% and 99% confidence level, respectively**.** n/a denotes analysis not conducted.

Season	**PDO**	**PNA**	**ENSO**	**NAO**	**AO**	**EPO**
GEP						
Winter	n/a	n/a	n/a	n/a	n/a	n/a
Spring	-0.35	0.08	0.16	-0.16	-0.21	-0.59*
Summer	-0.14	0.06	0.20	-0.22	-0.18	0.13
Autumn	-0.31	-0.26	0.12	0.27	0.40	0.03
Annual	-0.35	-0.02	0.29	-0.14	-0.09	-0.14
R_E						
Winter	-0.62*	-0.18	0.36	0.25	0.17	-0.79**
Spring	-0.24	0.05	0.02	-0.12	-0.27	-0.60*
Summer	-0.05	0.37	-0.11	-0.07	-0.18	-0.16
Autumn	-0.19	-0.21	-0.12	0.56	0.47	-0.11
Annual	-0.26	0.16	-0.05	0.12	-0.02	-0.43
NEP						
Winter	-0.11	0.30	-0.35	0.12	-0.07	-0.62*
Spring	-0.26	-0.09	0.28	-0.13	-0.02	-0.18
Summer	-0.15	-0.32	0.44	-0.24	-0.06	0.38
Autumn	-0.15	-0.05	0.40	-0.55	-0.18	0.22
Annual	-0.23	-0.27	0.57	-0.41	-0.12	0.37

Climate oscillation influences on forest carbon exchange have been studied in other parts of North America. Grant et al. (2009) examined the impact that changes in temperature and precipitation associated with large-scale weather events have on NEP from flux towers along a transcontinental transect of forest stands in Canada. Collectively the results suggested that climate oscillations were reflected in NEP, although patterns may vary with climate zone, vegetation species, and topography. Overall warm episodes, which can be associated with ENSO and PDO events, were found to influence diurnal CO_2 exchange of temperate and boreal confiners, but had little effect on boreal deciduous forests (Grant et al., 2009). Such warm events extended the period of net carbon uptake increasing annual NEP at boreal coniferous and deciduous sites, but had the opposite effect at a temperate coniferous site. However, these changes in NEP are more likely to occur within regions most affected by ENSO and PDO events, areas closer to the western coast of Canada.

Hember and Lafleur (2008) investigated the connections between the NAO and the North Pacific Oscillation (NPO), another teleconnection pattern that is correlated with the ENSO, and surface air temperature, precipitation and carbon fluxes throughout a range of mid-latitude North American ecosystems by examining eddy-covariance flux data spanning 1994 to 2006. For spring, correlation in the southeast stations were positive between NAO and surface air temperature, while the NPO was positively (negatively) correlated with temperature (precipitation) in the northeast. They showed that GEP and R_E were positively correlated with NAO during the spring and summer at the south temperate maple tulip forest located in Indiana, USA. Eastern stations were inversely correlated with NPO during spring, while positively correlated during the summer. For this study, the NAO only showed a significant positive correlation with winter temperatures, whereas the ENSO was found to be significantly correlated with snowfall. No significant correlations were shown between NAO, ENSO, and carbon fluxes, however, spring and summer fluxes were negatively correlated with NAO and positively correlated with ENSO.

Zhang et al. (2011) examined annual flux measurements from a deciduous forest in central Massachusetts, a coniferous forest in Maine, and a mixed-hardwood deciduous forest in Indiana. Their results indicated that climate oscillations influence annual fluxes through their effects on the local surface climate. Minimum spring and summer temperatures, and autumn temperatures were significantly inversely correlated with annual net ecosystem exchange (NEE = -NEP) and annual GEP. Winter precipitation had a significant positive correlation with annual NEE and GEP, while autumn precipitation had a significant negative correlation with annual NEE only. They also found that annual GEP was significantly related with autumn Atlantic Multidecadal Oscillation (AMO), winter EPO, spring Multivariate ENSO Index (MEI) and PDO. Annual NEE was significantly correlated with autumn AMO and PDO, while annual respiration responds to previous autumn ENSO and PNA indices. This study showed strong correlations between winter and spring mean temperatures, and total snowfall against the winter and spring mean carbon fluxes. The current study found similar relationships between winter temperature and NEP, and autumn temperature and GEP (data not shown in Table 2.3). Zhang et al. (2011) noted a significant positive correlation between spring precipitation and annual NEP, which was not observed in this study. Against the indices, this study showed that the EPO was correlated with winter and spring carbon fluxes, and the PDO correlated to winter R_E. All other significant correlations were not found with the carbon fluxes for this study.

Although previous studies and the current study identified a relationship between local climate oscillations and carbon exchanges, differences in findings for this study and that of Grant et al. (2009), Hember and Lafleur (2008) and Zhang et al. (2011) could be attributed to differences in geographic locations, length of data sets, hydrological conditions and different forest species. This study found a significant link between climate oscillations and carbon fluxes, where local climate conditions were modified by the regional circulation in the Great Lakes region in eastern North America. Compared to other studies, the strong linkage was found between winter variables and winter and spring carbon fluxes, such as R_E and NEP. In particular, the effect of a warming event, which occurred in March 2012, on carbon fluxes demonstrated the severity of these impacts to warming events, or heat stresses, which are predicted to become more frequent and intense in the future (IPCC, 2012).

2.5 CONCLUSIONS

Long-term analyses of temperature and precipitation data in southern Ontario, Canada showed that local climate is strongly influenced by low frequency climate oscillations. Mean winter temperatures were positively correlated to the NAO and the AO indices and negatively correlated to the EPO index. Total winter precipitation was mainly influenced by the PNA and AO indices, while snowfall was found to have a strong relationship with the PNA index and ENSO events. These influences have shown to affect R_E and NEP during the winter and GEP and R_E during the spring. The only direct connection between carbon fluxes and climate oscillations was a significant correlation between winter and spring carbon fluxes and the EPO index, and the winter respiration and the PDO index. The winter of 2011/12 was unusual with strong negative phases of the PNA, PDO and EPO occurring along with positive phases of both the NAO and AO. These strong phases of climate oscillations generated record warm conditions where mean winter temperature was 4°C above the mean seasonal winter temperature, causing a decrease in snowfall despite a slight increase in total precipitation. These above-normal winter temperatures had a significant impact on winter carbon fluxes with increases observed in R_E (+34 g of C) and NEP (+39 g of C) over the seasonal mean. In March 2012, a warming event linked to climate oscillations, raised daily mean temperatures to 25°C above the 30-year mean monthly temperature and elevated temperatures persisted for fifteen consecutive days. This event alone released approximately 60 g of C from the study site over the month of March. As large-scale modes of climate variability are important drivers to changes in climate extremes, changes to these modes will affect the intensity, frequency and spatial patterns of extreme climatic events. These changes can partially offset carbon sinks or even cause net losses in carbon stocks in plantation forests in the Great Lakes region. The knowledge gained from the current research pertaining to carbon sequestration in forest ecosystems, and the effect of low frequency climate oscillations on forest growth and their survival, should aid in developing polices for better managements of forest ecosystems in eastern Canada. It will also aid in planning and developing realistic strategies to offset fossil fuel CO_2 emissions to improve environmental quality.

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CHAPTER 3

RESPONSE OF SOIL CO_2 EFFLUX TO SHELTERWOOD HARVESTING IN A MATURE TEMPERATE PINE FOREST

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Abstract: In forest ecosystems, soil CO_2 efflux is an important component of ecosystem respiration (R_E), which is generally driven by variability in soil temperature and soil moisture. Tree harvesting (or thinning) in forests can alter the soil variables, and consequently impact soil CO_2 efflux. This study investigated the response of total soil CO_2 efflux, and its components, to a shelterwood harvesting event of a mature temperate white pine (*Pinus strobus L.*) forest located in southern Ontario, Canada. The objective was to explore the response of soil CO_2 effluxes to changes in the forest microclimate, such as soil temperature and soil moisture, after shelterwood harvesting removed approximately one-third of the overstory canopy. No significant differences were found in both soil temperature and soil moisture between the pre-harvesting (2008 to 2011) and post-harvesting (2012 to 2014) periods. Despite similar soil microclimates, total soil CO_2 effluxes were significantly reduced by up to 37%. Soil CO_2 effluxes from heterotrophic sources were significantly reduced post-harvesting by approximately 27%, while no significant difference in the mineral-soil horizon sources were measured. An analysis of R_E, measured with an eddy-covariance tower over the study area, showed an increase post-harvesting. However, the overall net ecosystem carbon exchange showed no significant difference between pre- and post-harvesting. This was due to an increase in the gross ecosystem productivity post-harvesting, compensating for the increased losses (i.e. increased R_E). This study highlights the complexities of soil CO_2 efflux after a disturbance, such as a harvest. The knowledge gained from this

study adds to our understanding of how shelterwood harvesting may influence ecosystem carbon exchange and will be useful for forest managers focused on carbon sequestration and forest conservation.

3.1 INTRODUCTION

Forest ecosystems account for about 80% of the world's biomass and are considered an important sink for atmospheric carbon dioxide (CO_2) (Pan et al., 2011). Forest carbon cycle is strongly influenced by the environmental and edaphic factors (Allen et al., 2010; Bastos et al., 2013; Davidson et al., 1998; Heimann and Reichstein, 2008; Hogg et al., 2017; Jassal et al., 2012; Khomik et al., 2010; Piao et al., 2008; Wharton et al., 2012). In addition, silvicultural practices, such as forest thinning, can also influence forest carbon dynamics by altering soil temperature, soil moisture, soil organic matter, root biomass and microbial activity (D'Amato et al., 2011; Gustafsson et al., 2012; Magruder et al., 2013; Peng et al., 2008; Powers et al., 2009; Vesala et al., 2005). These silvicultural practices are being adopted in North America and across the world, not only to manage forests, particularly planted stands, but also to enhance their carbon sequestration capabilities and/or to restore them to their previous native species (Gustafsson et al., 2012).

In the Great Lakes – St. Lawrence Region of Canada, the Ontario Ministry of Natural Resources and Forestry (OMNRF) uses a number of silvicultural practices based on the characteristics of the current forest, its history and the desired forest condition. One such practice is a specific shelterwood harvest, which consists of selectively removing mature trees, while maintaining legacy trees (largest, best quality), in a series of two or more partial cuts, over the course of 10 to 30 years. Shelterwood harvest encourages natural regeneration and development of seedlings in partial shade (OMNRF, 1998; OMNRF, 2000; Trant, 2014). Shelterwood harvesting changes the tree spacing, number and size distribution, which can affect the dynamics of carbon uptake through above-ground productivity and carbon losses through soil respiration processes (Marchi et al., 2018; Vesala et al., 2005). More open-canopy alters the radiation dynamics, interception of precipitation, wind flow patterns, water vapour deficit, soil temperature and soil moisture (Stoffel et al., 2010; Vesala et al., 2005; Woods et al., 2006). Shelterwood harvesting also causes changes in root density and production altering autotrophic respiration, litterfall input and hence soil microbial activity. These changes may cause variations on soil CO_2 efflux, also referred to as soil respiration (Peng and Thomas, 2006).

Peng et al. (2008) found that amongst the studies that examined the impacts of forest harvesting on soil respiration, only a few have examined selective harvesting or thinning effects, with little information on spatial and temporal patterns of soil CO_2 efflux response in these forests, particularly in conifer stands. This is important because an increasing number of temperate conifer forests in eastern North America are being managed to enhance their carbon sequestration capabilities to restore them to their native species composition, enhance biodiversity, and conserve water resources. Therefore, there is an urgent need to enhance our understanding of the effects of silvicultural practices, in particular shelterwood harvest, on soil processes and carbon sequestration (Jandl et al., 2007).

This study investigated the response of soil CO_2 efflux and its components to a shelterwood harvesting event in a mature temperate white pine (*Pinus strobus L.*) forest, located near Lake Erie in southern Ontario, Canada. The purpose of the study was to explore the response of soil CO_2 efflux dynamics to changes in forest microclimate, such as soil temperature and soil moisture, after a partial (i.e. selective) thinning event. Both soil temperature and soil moisture have been previously shown to strongly influence soil respiration at this site (Khomik et al., 2010). Therefore, seasonal variability of soil temperature, soil moisture and soil CO_2 efflux were measured using an automated soil CO_2 chamber system (Drewitt et al., 2002; Jassal et al., 2005) and comparisons were made between the pre- (2008 to 2011) and post-harvesting (2012 to 2014) periods. A previous study at the site showed an approximate 10% decrease in the total ecosystem respiration (R_E) after thinning, using the eddy-covariance technique (Trant, 2014). Given that soil respiration is a major component of R_E at the site (Khomik et al., 2010) the impact of shelterwood harvesting on the forest carbon fluxes were measured using the automated soil CO_2 chamber system and eddy-covariance method.

3.2 METHODS

3.2.1 *Study Site*

This study was conducted at a mature white pine forest near Turkey Point Provincial Park on the northern shore of Lake Erie in southern Ontario, Canada. This site is part of the Turkey Point Flux Station (TPFS), or Turkey Point Observatory of the Global Water Futures (GWF) program, and global Fluxnet. The forest is dominated (>82%) by eastern white pine (*Pinus strobus L.*), which was planted in 1939 to stabilize local sandy soils. As of 2014, the average tree height was about 21.8±1.7 m, and stand density was about 421±166 stems·ha^{-1}. Other tree species include 11% balsam fir (*Abies balsamea L. Mill*) and species native to the hardwood forests of North American Eastern Temperate Forest Ecoregion, including 4% Oak (*Quercus velutina L.*, *Q. alba L.*), 2% Red Maple (*Acer rubrum L.*) and some wild black cherry trees (*Prunus serotina Ehrh*). The understory consists of ferns (*Pteridium aquilinum L.*), mosses (*Polytrichum spp.*), poison ivy (*Rhus radicans L.ssp.*) and Rubus species. The topography at the site is fairly flat with well drained sandy soil (Brunisolic Gray Brown Luvisol, following the Canadian Soil Classification system), which is composed of ~98% sand, 1% silt, and 1% clay. Further site details are provided in Arain and Restrepo-Coupe (2005) and Peichl et al. (2010a; b).

3.2.2 Shelterwood harvest treatment

This forested site is managed by the Ontario Ministry of Natural Resources and Forestry (OMNRF) under the shelterwood silvicultural system

(OMNRF, 1998; OMNRF, 2000; Trant, 2014). This system is characterized by two or more partial thinning treatments over a few decades to allow for regeneration and development of seedlings in partial shade. The forest was first thinned in 1983, where one-third of trees were randomly removed. In March of 2012, a second partial cut was conducted and approximately 30% of the dominant overstory trees were selectively removed to improve light and water availability and stimulate growth of the remaining trees. The long-term aim of the management is to convert the pine plantation back to the native stand with mixed-wood ecosystem characteristics (OMNRF, 1998). A mechanical harvester was used to cut, de-limb and section the selected trees. To reduce soil compaction and disturbance, the harvester used pre-existing multi-use trails throughout the stand. Thinning residues (e.g. limbs, bark, crowns) were placed on the ground ahead of the harvester to limit compaction when trails were not accessible (Trant, 2014). After shelterwood harvesting the average basal area was reduced by 13% (Kula, 2014), while peak leaf area index was reduced by 35% (Trant, 2014). It was estimated that the thinning added 43 t of foliage, 200 t of live branches and 129 t of dead branches to the forest floor, while an additional 515 t of roots were made inactive from growth and production (Trant, 2014). These components will decompose and are expected to contribute significantly to the ecosystem respiration.

3.2.3 *Soil CO_2 efflux measurements*

Soil CO_2 efflux was measured using an automated chamber system. The main advantage of an automated chamber system, compared to manual measurements, is the ability to take continuous long-term measurements of soil CO_2 efflux, which are useful in studying short- and long-term responses of soil CO_2 efflux to environmental variables (Jassal et al., 2012). The chamber system occupied an approximate area of 50 m × 50 m and was located about 100 m north from the eddy-covariance flux tower at the site. The experimental area was chosen to complement the ecosystem carbon flux measurements by the flux tower. The individual chamber locations, although few in replicates due to logistical and budgetary constraints, were placed in locations within the footprint of the eddy-covariance flux tower measurements, with the aim to provide data on the soil CO_2 efflux contribution to the overall ecosystem respiration. Chambers were placed away from tree stems, in order to avoid large roots, particularly for those chambers measuring heterotrophic respiration. Additionally, the chambers were located away from thick understory and placed on level ground. As soil respiration is comprised of contributions from mineral-soil, organic litter and roots with their associated mycorrhizae, chambers were strategically placed to elucidate the CO_2 contribution from these different components to the total soil respiration.

The chamber system was established to measure total soil CO_2 efflux, including those from root exclusion and forest floor removal plots. The chamber system was installed in June 2008, with the initial installation of four chambers. One chamber was established to measure heterotrophic respiration (F_{RT1}) in the trenched plot whereby live tree roots were severed in the area surrounding the chamber (Nicholas, 2011). The other three chambers were used as control chambers and measured total soil CO_2 efflux (F_S: F_{S1}, F_{S2}, and F_{S22}). One of the control chambers (F_{S22}) was altered in May 2009 to measure the contributions from the mineral-soil horizon (F_{LR}) to total soil respiration, by removing the Litter-Fermenting-Humified (LFH) layer. Additional litter, which may have settled inside this chamber, was periodically removed, including any material from the harvesting event. Additionally, in May 2009, two more control chambers (F_{S3} and F_{S4}) were added to the chamber system to measure total soil respiration. Finally, in May 2010, two additional chambers were installed, one as a control (F_{S5}) and an additional chamber to measure heterotrophic respiration (F_{RT2}); increasing the total number of chambers in the chamber system to eight. Any vegetation that grew inside the collars between visits was removed to eliminate any potential photosynthesis effect.

Soil CO_2 efflux was continuously measured at each chamber location, using automated non-steady state chambers developed by the University of British Columbia (Drewitt et al., 2002; Gaumont-Guay et al., 2008; Jassal et al., 2005; Jassal et al., 2012). Each chamber consisted of a collar and a chamber lid that fit onto the collar. The collar was constructed from a polyvinyl chloride (PVC) cylinder and were inserted approximately 2 to 4 cm into the ground, to avoid cutting roots near the soil surface. Additionally, the height of the collar was maintained at a minimum height above the ground so when the chamber lid was open, there was no reduction in air flow over the soil surface within the collar (Jassal et al., 2012). The chamber lid consisted of a transparent plastic dome fixed to a metal frame and the lid was same diameter as the collars. A torsion spring provided enough force to close the chamber dome, while a chamber-mounted, two-way pneumatic cylinder opened the dome when compressed air was pushed through the tubing (Gaumont-Guay et al., 2008; Nicholas, 2011). When the dome closed during measurements, a foam gasket sealed the dome around the collar on which the chamber rested, providing an air-tight seal. A small fan inside the chamber circulated the air, preventing stagnation zones and microclimate formation. When a chamber was not in use, the chamber lid was kept open as to allow precipitation and litter to naturally settle into the collar area.

The chambers were controlled by a computer housed in a datalogger box, which collected and processed data, and controlled the instrument pumps and gas analyzer equipment. Cycling through the chambers, including opening and closing their lids, was done with custom MATLAB (The MathWorks Inc.) programs. Each chamber closed individually for a one-minute interval, where the CO_2 concentration was sampled and measured by an infrared gas analyzer (model LI-840, Li-COR Inc.). Measurements were cycled through the eight collars for a total of three cycles per half-hour period. Thus, each half-hour consisted of three, minute-long CO_2 concentration measurements per chamber. One half-hourly value was produced for each collar by averaging the three measurements taken during that half-hour. These average values were used for computation and analysis. Concentrations measured during the fifteen seconds following dome closure were discarded to ensure the samples tubes were relatively free from air from the previous sampled chamber and to account for the time taken for the chamber lid to completely close.

In order to measure soil CO_2 efflux (F_S) with these chambers, the measured time rate of change in chamber headspace CO_2 mole fraction or mixing ratio are estimated (Jassal et al., 2012). The soil CO_2 efflux (µmol CO_2 m^{-2} s^{-1}) from each chamber was calculated as:

$$Fs = \rho_a \frac{Ve}{A} \frac{dSc}{dt} \tag{3.1}$$

where ρ_a is air density in the chamber headspace (µmol m^{-3}), Ve is the effective volume of the chamber (m^3), A is the area (m^2) of the soil surface covered by the chamber collar, and dSc/dt is the time rate of change of CO_2 mixing ratio in the chamber headspace (µmol CO_2 mol^{-1} dry air s^{-1}).

The mean Ve value was calculated using the following equations by injecting CO_2 through the top of the chamber domes for one minute and recording the CO_2 concentration change (in ppm):

$$Ve = \frac{IRT}{PV(S_c - S_m)} \tag{3.2}$$

where S_c (µmol CO_2 µmol^{-1} dry air s^{-1}) is the rate of CO_2 concentration increase during the calibration period, I (µmol CO_2 s^{-1}) is the rate of injection of CO_2 during the calibration period, S_m (µmol CO_2 µmol^{-1} dry air s^{-1}) is the rate of change of CO_2 concentration, P is the atmospheric pressure (Pa), V is the volume of the chamber (m^3), T is the chamber air temperature (K), and R is the universal gas constant, (8.314 J µmol^{-1} K^{-1}).

According to the work done by Jassal et al. (2012), an accurate estimation of the effective chamber volume will help in acquiring good measurements of soil CO_2 efflux. Effective chamber volume varies seasonally due to snow and litterfall, and the expansion and contraction of the chambers and collars due to variation in air temperature.

Due to limited weekly field site visits, chamber volumes were measured a few times during the growing season, primarily to check changes in the litter-layer within the collars and to remove vegetation. The mean value of *Ve*, calculated using Equation 3.2, was 0.069 m^3 during the snow-free season (Nicholas, 2011). During the snow season snow depth measurements were taken inside the chambers, when possible, and used to adjust *Ve*.

Measuring soil CO_2 efflux during the winter was challenging as snowfall can completely fill the collars, heavy ice and snow on the dome can force the chamber to remain shut for extended periods of time and cold temperatures can create leaks in sample tubes or compressed air lines (Nicholas, 2011). Frequent site visits were made during the winter season to remove snow from the top of the dome and around the collar rim to ensure a continued tight seal for measurements.

3.2.4 *Data processing and quality control of soil flux data*

Soil CO_2 efflux data were quality controlled with a custom script written in MATLAB software. In the script, the slope of time versus efflux in the high frequency (1 Hz) data for each measurement interval was estimated. This was a process by which half-hourly soil CO_2 efflux data were reviewed to identify and subsequently remove non-representative measurements. For the quality assessment, first the root-mean-square-error (RMSE) of the linear fit of chamber headspace CO_2 mole fractions versus time was calculated. Next, by applying a threshold value of 1 for the ratio of RMSE to the slope of the linear fit of dSc/dt, all values above this threshold were considered unreliable and were removed following a method similar to Jassal et al. (2012). A visual inspection was also conducted to remove any remaining erroneous data. Negative fluxes were discarded as any vegetation, whether by moss or small plants, were removed on the soil surface within the chambers.

Besides data removal through quality control and data omitted from the throughfall exclusion experiment, data loss also occurred from chamber malfunctions, calibrations, the lack of compressed air available, which controls the ability of the chamber dome to close, and winter snow and ice accumulation.

Beyond the quality control described above, measurements from the automated chambers were compared with soil respiration measurements using a portable photosynthesis/soil respiration system (model LI-6400, LI-COR Inc., NE, USA) which can measure soil CO_2 effluxes levels from individual points (Figure 3.1). Portable soil chamber systems are able to better capture spatial variability, while the automatic chamber system can better capture temporal variability in soil CO_2 fluxes. Time stamp of the individual measurements showed that soil respiration values measured by the automatic chamber system were comparable to those measured by the portable LI-6400 system.

In 2013, hydraulic problems arose in the two-way pneumatic cylinder, resulting in approximately 80 to 90% of annual data loss for most of the chambers (Table 3.1), all occurring during the growing season (Table 3.2). Further mechanical problems in the two-way pneumatic cylinder occurred in 2014, with about 70 to 90% of the data removed (Table 3.1). The percentage of data retained decreased over the years due to a combination of equipment age, malfunctions and removal of the chamber system (November 20, 2011 and March 27, 2012) for forest thinning operations (Table 3.3). Note that a throughfall exclusion experiment was conducted at the site from April to July of 2009, where an area, 20 m × 20 m, was covered by flat-bottomed aluminum troughs (MacKay et al., 2012). Three chambers were covered and used for the experiment (Nicholas et al., 2011) and data during this time period was omitted from this study.

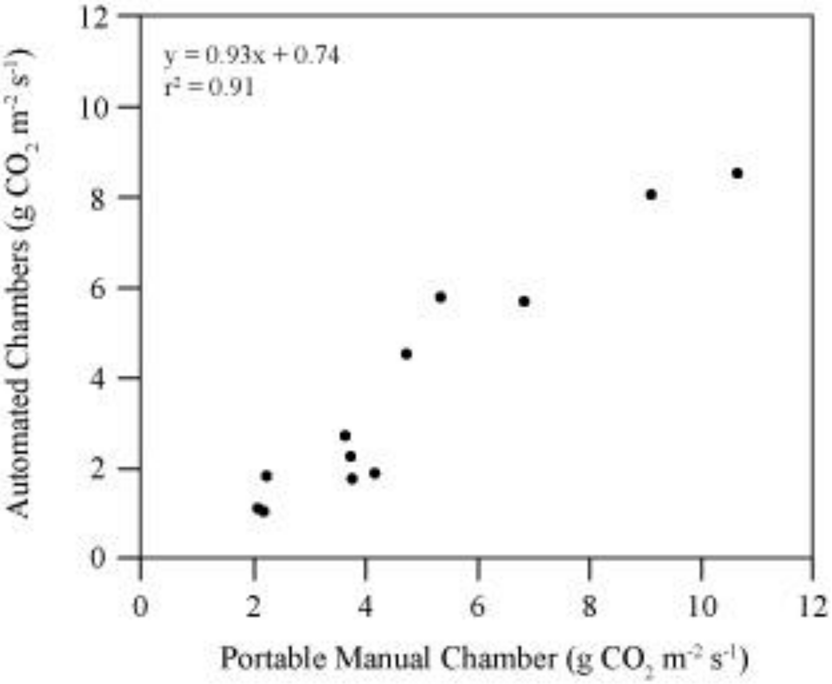

Figure 6. Comparison of half-hourly total soil CO2 efflux values from the automated chamber system against the portable manual soil chamber system (model LI-6400, LI-COR Inc.) in 2014**.**

Table 0.1. Percentage of half-hourly total soil CO_2 efflux values, and its components, for each year, from 2008 to 2014. (Dash indicates that the collar was not installed).

Year	F_{LR}	F_{S1}	F_{RT1}	F_{S2}	F_{S3}	F_{S4}	F_{S5}	F_{RT2}
2008	82	74	39	76	-	-	-	-
2009	69	63	68	61	51	40	-	-
2010	75	74	44	75	78	67	52	24
2011	54	45	42	55	46	42	58	52
2012	26	20	22	23	22	22	23	23
2013	12	13	8	11	24	27	10	11
2014	33	21	17	13	18	19	33	7

Table 0.2. Percentage of half-hourly total soil CO_2 efflux values, and its components, for the growing season (April to October), from 2008 to 2014. (Dash indicates that the collar was not installed).

Year	F_{LR}	F_{S1}	F_{RT1}	F_{S2}	F_{S3}	F_{S4}	F_{S5}	F_{RT2}
2008	83	77	25	78	-	-	-	-
2009	78	74	74	45	62	45	-	-
2010	81	78	35	77	81	64	67	23
2011	73	59	57	72	54	57	71	71
2012	42	32	36	38	36	36	38	39
2013	0	0	0	0	28	30	0	0
2014	39	18	16	9	15	16	36	2

Table 0.3. Percentage of half-hourly total soil CO_2 efflux values, and its components, for the winter season (November to March), from 2008 to 2014. (Dash indicates that the collar was yet installed).

Year	F_{LR}	F_{S1}	F_{RT1}	F_{S2}	F_{S3}	F_{S4}	F_{S5}	F_{RT2}
2008-09	70	53	59	75	-	-	-	-
2009-10	54	61	59	82	80	72	-	-
2010-11	55	53	46	58	64	54	72	52
2011-12	2	0	0	0	0	0	0	0
2012-13	14	18	10	14	8	13	12	13
2013-14	32	32	22	31	22	27	36	21

3.2.5 *Eddy-covariance flux and ancillary data*

Water, carbon and energy fluxes were measured using a closed-path eddy-covariance system. Flux measurements were made at 20 Hz above the canopy on top of a scaffolding tower at a height of 28 m. Above canopy fluxes of carbon and water were calculated using a methodology outlined by Baldocchi et at. (1988):

$$F = \overline{w'}\overline{c'} \tag{3.3}$$

where F refers to the flux of energy, carbon or water, w' is the covariance of vertical wind speed, measured by a sonic anemometer (CSAT3, Campbell Scientific Inc. (CSI)), and c' is the covariance of CO_2 or water vapour concentrations, measured by an infrared gas analyzer (Li-7000, LI-COR Inc.). Half-hourly net ecosystem productivity (NEP) was calculated by adding CO_2 flux and the CO_2 storage, which is calculated as change in CO_2 concentration over time in the air column below the eddy-covariance sensors (Brodeur, 2014). Ecosystem respiration (R_E) was estimated using a non-linear logistic relationship between nocturnal half-hourly CO_2 fluxes during high turbulence conditions and soil temperature at 2 cm depth (Brodeur, 2014; Peichl et al., 2010a; b). Gross ecosystem productivity (GEP) was determined by adding measured NEP to model daytime R_E. Soil temperature and soil moisture were measured at 5, 10, 20, 50 and 100 cm depths at two locations, near the chamber system location. For the analysis, soil temperature at a depth of 5 cm and soil moisture, averaged between 0 and 30 cm, were used. All flux and meteorological data were quality controlled and averaged at half-hourly intervals (Brodeur, 2014). For further details regarding instrumentation, the eddy-covariance tower is described by Arain and Restrepo-Coupe (2005) and Peichl et al., (2010a; b).

3.2.6 *Statistical analysis of pre- and post-harvesting*

To determine whether the shelterwood harvesting event can be detected within the variability of the data, a comparison was made for the pre- (2008 to 2011) and post-harvesting (2012 to 2014) periods. Half-hourly soil temperature and soil moisture values were averaged daily to allow for a comparison of adequate sized data points between the two periods (Hungate et al., 1996; Johnson, 1999). As soil CO_2 efflux measurements had several half-hourly gaps after quality control, only available measurements taken at 11:00 for each day were compared, as the mean daily soil CO_2 efflux value was found to correspond best to values around 11:00 at this site, and in previous studies (Cueva et al., 2017). This study was not a truly replicated experiment and the results needed to be viewed and interpreted cautiously due to pseudoreplication (Hurlbert, 1984).

The soil temperature, soil moisture, soil CO_2 efflux, and eddy-covariance flux data (all denoted by x' in Equation 3.4) were normalized using the min-max normalization method where the values are rescaled to range between 0 and 1 using the following equation:

$$x' = \frac{x - \min(x)}{\max(x) - \min(x)} \qquad (3.4)$$

where min(x) and max(x) are the minimum and maximum values of x for the complete data (2008 to 2014). An F-test was then applied to assess whether the variances of the two periods were equal or not, followed by a t-test to determine if there was any significant difference ($p\text{-value} < 0.05$) in the means.

Furthermore, as these variables change seasonally, the relative contribution of component fluxes to the total soil CO_2 efflux were examined for different months of the year. Following similar work to Heinemeyer et al. (2011), two-month averages of diurnal variability of soil temperature and soil CO_2 efflux rates, for each individual chamber, were constructed for March to April, July to August, August to September, and December to January. To check for changes, diurnal patterns from pre- and post-harvesting years were compared. Comparisons for the diurnal patterns from pre- and post-harvesting were made for selected two-month seasons in those years with a similar climate: when their two-month average temperature and total precipitation were similar.

3.3 RESULTS

3.3.1 *Seasonal variations in soil temperature and soil moisture*

During the study, the site experienced a large seasonal variation in temperature with air temperatures reaching above 30°C in the summer, and well below 0°C in the winter. This variation was also reflected in the soil temperatures measured at 5 cm depth, but less pronounced.

Daily soil temperatures fluctuated around 0°C in the winter, increased by April, peaked in July or August, and then decreased again by October (Figure 3.2a). The mean soil temperature for the seven-year data period from 2008 to 2014 was 9.3±7.5°C, with the highest soil temperature of 23°C recorded during the summer of 2013, and the lowest temperature of -2.9°C recorded during the winter of 2012. However, in 2012, above-normal temperatures occurred in March; increasing soil temperatures well above 10°C for approximately twenty days.

Daily soil moisture showed large variations during the seven-year study period, which is attributed to rainfall and/or snowmelt events, and influenced by the well-drained sandy soil at this site. Annually, soil moisture increased during the winter season, with maximum values observed after snowmelt (Figure 3.2b). Little variability in soil moisture was observed during the 2012 growing season, as very few rainfall events occurred, and this year was deemed a drought year (Skubel et al., 2015). The mean soil moisture over the seven-year study period was 0.12±0.03 m^3/m^3, with the highest value of 0.25 m^3/m^3 recorded during the spring of 2013 due to a combination of snowmelt and precipitation. In each growing season, soil moisture levels in this well-drained soil declined to approximately 0.065 m^3/m^3.

3.3.2 *Seasonal variability of soil CO_2 efflux*

From 2008 to 2014, the soil CO_2 efflux rates showed seasonal patterns following seasonal soil temperature, where CO_2 effluxes were low during winter months and high during the summer months. There was a large amount of variability in soil CO_2 effluxes during the growing season and much less during the winter months. Soil CO_2 efflux from the control chambers (Figure 3.2c and 3.2d) showed the highest daily average values at 3.1±2.0, 2.5±1.8, 3.2±2.2,

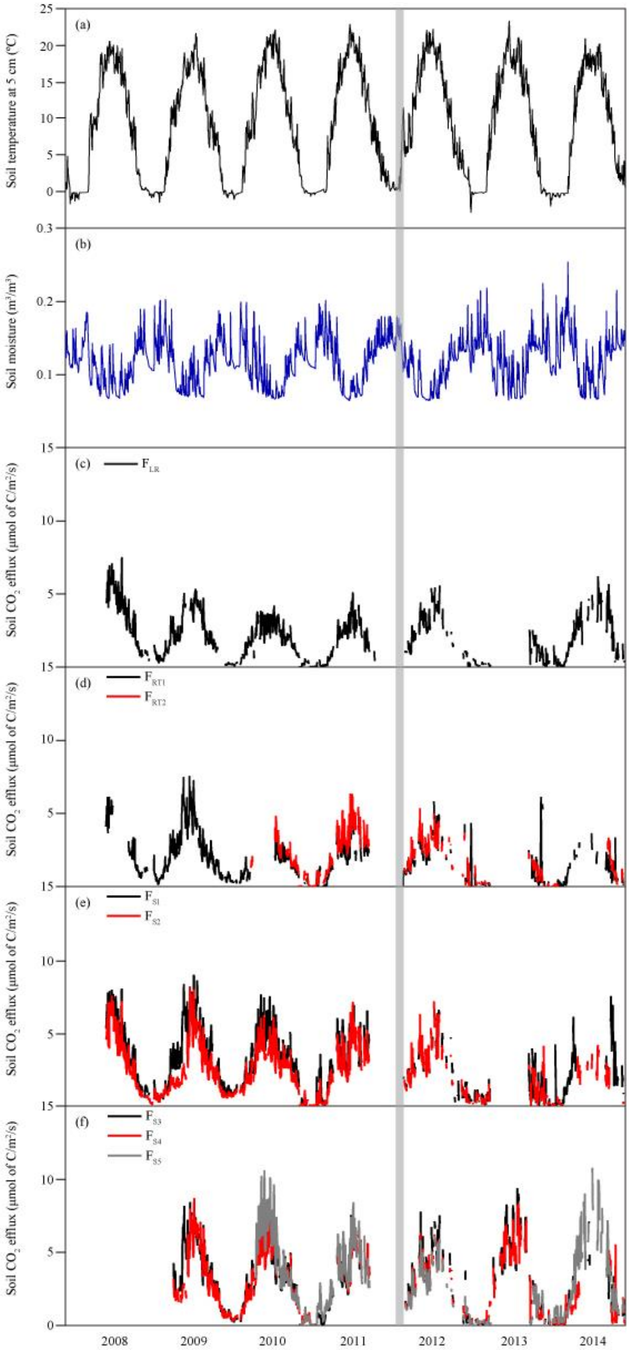

Figure 7. Daily values of (a) soil temperature at 5 cm depth, (b) soil moisture averaged between 0 and 30 cm depth. (c) soil CO_2 efflux from mineral-soil (F_{LR}), (d) soil CO_2 efflux from root trenched (F_{RT1} and F_{RT2}), (e) soil CO_2 efflux from initial group of control (F_{S1} and F_{S2}) chambers and (f) soil CO_2 efflux from latest group of control (F_{S3}, F_{S4} and F_{S5}) chambers, from 2008 to 2014. Grey line denotes timing of the shelterwood harvesting event.

3.1±2.1 and 3.3±2.4 μmol of C m^2 day^{-1} for F_{S1}, F_{S2}, F_{S3}, F_{S4} and F_{S5}, respectively. Both F_{S1} and F_{S2} showed high CO_2 effluxes in 2009, compared with the following years. Soil CO_2 effluxes for F_{S4} were higher in 2010 and 2014 compared with the other control chambers. Equipment malfunctions limited the number of observations for 2013 and 2014, with chamber data only from F_{S3} and F_{S5} in 2013, and F_{S4} in 2014, passing quality control. No observations during the 2011/12 winter season were recorded due to the removal of the automatic chamber system for shelterwood harvesting.

Before removal of the litter-layer, F_{LR} produced similar soil CO_2 effluxes to both F_{S1} and F_{S2} (Figure 3.2e). After litter-layer removal, soil CO_2 effluxes dropped and remained lower than the control chambers, producing daily average value of 2.2±1.5 μmol of C m^2 day^{-1} over six years.

The two chambers where soil was trenched to exclude tree roots also produced lower values of soil CO_2 efflux compared to the control chambers, with daily average values of 2.0±1.5 and 2.2±1.5 μmol of C m^2 day^{-1} for F_{RT1} and F_{RT2}, respectively. These averaged daily soil CO_2 efflux values were similar to those from F_{LR} for most years, except 2009 and 2014.

Mean averaged half-hourly soil CO_2 efflux rates, compared to soil temperature and soil moisture, showed diurnal fluctuations that mimic the diurnal pattern of soil temperature, while responding to increases in soil moisture (Figure 3.3). An example from June 2009 showed that soil CO_2 effluxes from the control (F_S), heterotrophic (F_{RT}) and mineral-soil horizons (F_{LR}) increased with an increase in both soil temperature and soil moisture. However, in this example, June 17 showed an instance where the mineral-soil component decreased at the onset of a rainfall event. Conversely and less common, instances of a decrease in the mineral-soil component occurred in the following years.

In January and December 2013, large pulses of soil CO_2 effluxes were observed from F_{LR}, $F_{RT1,}$ F_{RT2} and F_{S2} chambers, often exceeding values recorded during growing seasons (Figure 3.2c, d and e). Based on observed meteorological data, these soil CO_2 efflux pulses seem to have corresponded to an increase in soil temperature, from temperatures below freezing to a few degrees above 0ºC, and an increase in soil moisture (from 0.14 to 0.20 m^3/m^3), both of which may have stimulated microbial activity. These rare observations could also be due to the melting of ice or dense snow cover on top of the soil within the chambers, which was blocking the release of soil CO_2 accumulated in the near-surface soil layer.

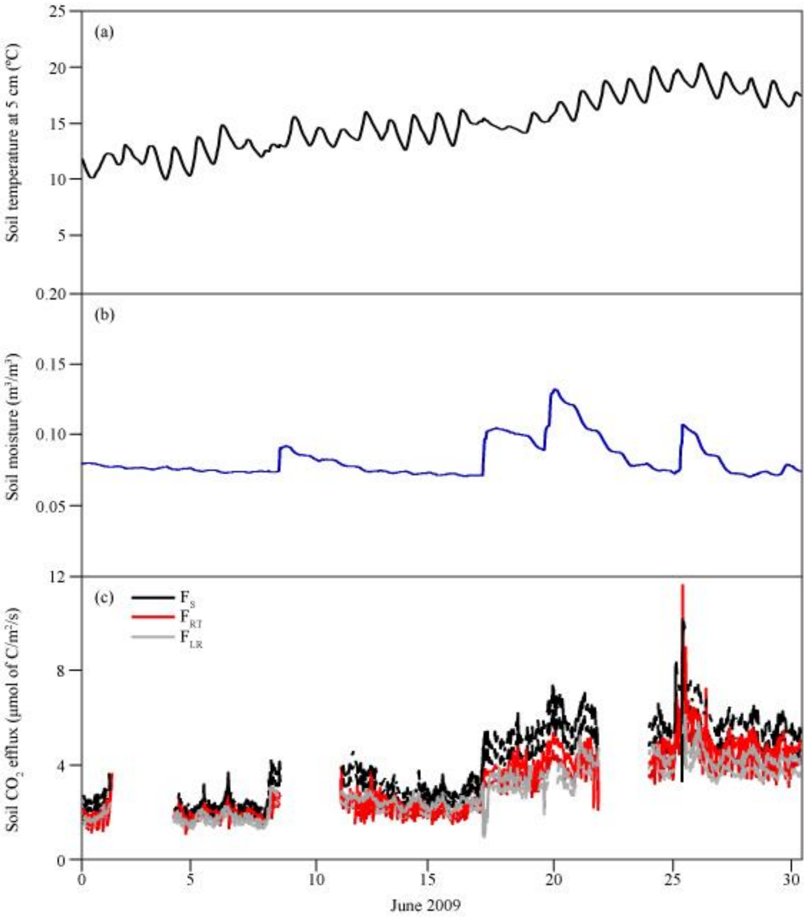

Figure 8. Daily (a) soil temperature at 5 cm depth, (b) soil moisture averaged between 0 and 30 cm depth, and (c) total soil CO_2 efflux (F_S), heterotrophic efflux (F_{RT}) and mineral-soil efflux (F_{LR}) for June 2009. Data gaps in (c) are the result of quality control and equipment malfunction (Section 3.2.4).

3.3.3 *Comparison of soil temperature and moisture pre and post shelterwood harvesting*

Diurnal pattern of soil temperatures over the two-month period for each season of the study showed a typical decrease overnight, then an increase coinciding with an increase in daylight, with peak values in late-afternoon (Figure 3.4a, 3.4g, 3.4m and 3.4s). Diurnal pattern of mean soil moisture over the July to August and August to September months showed increased overnight values that decreased as the day progressed (Figure 3.4h and 3.4n). For the March to April and December to January months, mean diurnal values of soil moisture stayed relatively flat (Figure 3.4b and 3.4t). Soil moisture for the winter season was higher than the growing season, due to the lack of evapotranspiration and an increase in precipitation events.

Mean values of soil temperature over the pre- and post-harvesting time periods were 9.4±7.5°C and 9.1±7.5°C, respectively, while the mean values of soil moisture were 0.11±0.03 m^3/m^3 and 0.12±0.03 m^3/m^3, respectively. A t-test with normalized daily data did not show a significant difference for both soil variables (Table 3.4). While soil temperature did not change much between the pre- and post-harvesting periods, an increase of 3% in soil moisture was observed, even with the data from a 2012 drought year included in the post-harvesting period.

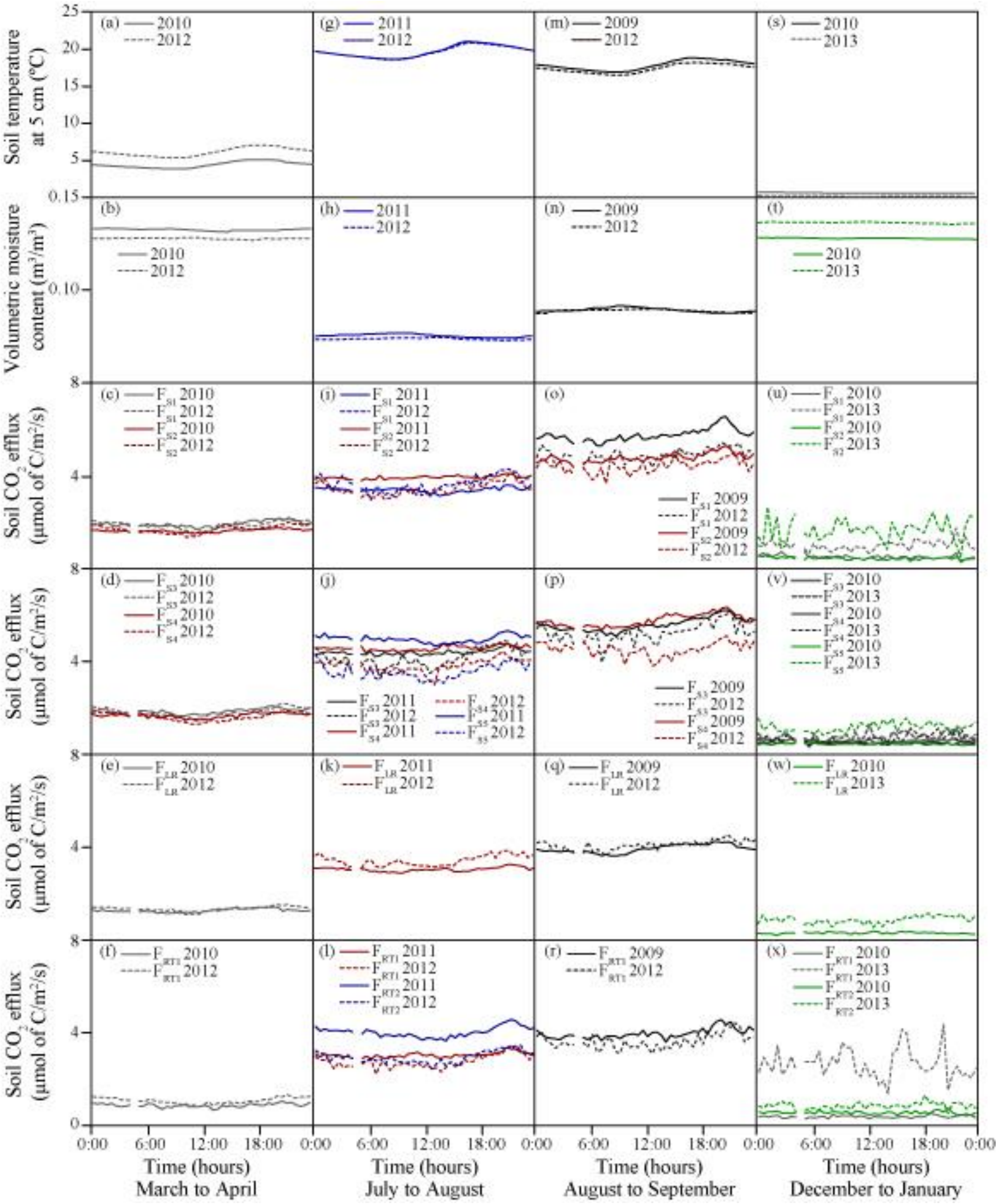

Figure 9. Two-month averages of the diurnal pattern for (a) soil temperature at 5 cm depth; (b) soil moisture averaged between 0 to 30 cm depth, and soil CO_2 efflux from (c) mineral-soil efflux (F_{LR}), (d) root trenched efflux (F_{RT}), (e) initial group of control (F_{S1} and F_{S2}) chambers and (f) latest group of control (F_{S3}, F_{S4} and F_{S5}) chambers. Diurnal average patterns from pre- (solid line) and post- (dotted line) harvesting event in 2012 are compared with the same months that have similar average temperature and total precipitation.

Table 0.4. Mean soil temperature, mean soil moisture and soil CO_2 efflux, from total soil CO_2 efflux (F_{S1}, F_{S2}, F_{S3}, F_{S4} and F_{S5}), heterotrophic respiration (F_{RT1} and F_{RT2}) and mineral-soil horizon respiration (F_{LR}), for pre- (2008 to 2011) and post-harvesting (2012 to 2014) periods, along with the percent difference. * denotes significant difference (α=0.05) of the normalized daily means for soil temperature and soil moisture and for those soil CO_2 efflux measurements taken at 11:00between pre- and post-harvesting periods.

Environmental Variable	Pre-Thinning (no. of observations)	Post-Thinning (no. of observations)	% Difference	t-statistic	p-value
Mean Soil Temperature (°C)	9.4±7.5 (n=1274)	9.1±7.5 (n=1095)	-3%	-0.90	0.37
Mean Soil Moisture (m^3/m^3)	0.11±0.03 (n=1274)	0.12±0.03 (n=1095)	3%	2.48	0.01
Soil CO_2 Efflux Component Chamber	Mean Soil CO_2 Efflux (μmol of C $\cdot m^2 \cdot s^{-1}$)		% Difference	t-statistic	p-value
F_{S1}	3.39±2.04 (n=780)	2.13±1.58 (n=183)	-37%	11.0*	0.00
F_{S2}	2.58±1.78 (n=919)	2.13±1.54 (n=152)	-17%	3.60*	0.00
F_{S3}	3.24±2.17 (n=642)	3.20±2.28 (n=219)	-1%	0.90	0.18
F_{S4}	3.29±2.10 (n=566)	2.73±2.09 (n=233)	-17%	3.59*	0.00
F_{S5}	3.74±2.50 (n=392)	2.63±2.18 (n=277)	-30%	6.73*	0.00
F_{LR}	2.09±1.32 (n=651)	2.06±1.49 (n=240)	-2%	0.41	0.68
F_{RT1}	2.14±1.56 (n=644)	1.57±1.27 (n=145)	-26%	5.52*	0.00
F_{RT2}	2.47±1.55 (n=274)	1.81±1.24 (n=130)	-27%	4.30*	0.00

3.3.4 Comparison of soil CO_2 effluxes pre and post shelterwood harvesting

Two-month diurnal average of soil CO_2 effluxes for each season showed differences in the range of magnitude. Although soil CO_2 efflux had more variability than soil temperature and soil moisture, for each season, the soil CO_2 efflux increased during the daylight, peaking around 21:00, about three to five hours after soil temperature peaks. Afterwards, emissions decreased overnight, with the lowest usually observed in the early morning hours. Soil CO_2 effluxes at approximately 11:00 were found to be close to the mean daily average for total soil CO_2 efflux and its components.

For March to April, representing spring, a comparison was made for years 2010 and 2012, which had similar observed climate. For the control (Figure 3.4c and 3.4d) and the litter-layer removed chambers (Figure 3.4e) minimal change was observed as the magnitude and diurnal pattern were similar. One trenched chamber (the other trenched chamber not being installed until May 2010), showed an increase with the average diurnal soil CO_2 efflux increasing from 0.86±0.07 to 1.09±0.11 μmol of C m^2 s^{-1} (Figure 3.4f).

In July to August, a comparison of 2011 to 2012 showed a decrease in the average diurnal soil CO_2 efflux for the control chambers, but an increase in the standard deviation (Figure 3.4i and 3.4j). The largest change was measured for F_{S5}, with a decrease in the average by 1.42 μmol of C m^2 s^{-1} and an increase in the standard deviation of 0.13 μmol of C m^2 s^{-1}. F_{LR} had an increase in both the mean and standard deviation, (3.09±0.07 μmol of C m^2 s^{-1} to 3.45±0.18 μmol of C m^2 s^{-1}; Figure 3.4k), while both F_{RT1} and F_{RT2} decreased in the average diurnal soil CO_2 efflux, and an increase in the standard deviation (3.07±0.09 μmol of C m^2 s^{-1} to 2.72±0.23 μmol of C m^2 s^{-1} and 4.02±0.18 μmol of C m^2 s^{-1} to 2.93±0.21 μmol of C m^2 s^{-1} for F_{RT1} and F_{RT2}, respectively; Figure 3.4l).

Comparing the two-month diurnal average of August to September, between 2009 and 2012, the four available control chambers all had a decrease in soil CO_2 effluxes (Figure 3.4o and 3.4p). The largest difference was seen in F_{S4}, with a daily average increase of 1.24 μmol of C m^2 s^{-1} and an increase in the daily average standard deviation of 0.02 μmol of C m^2 s^{-1}. F_{LR} had a slight increase in both the daily average and standard deviation post-harvesting (Figure 3.4q). The daily average soil CO_2 efflux for F_{RT1} decreased by 0.29 μmol of C m^2 s^{-1} between 2009 and 2012, while the standard deviation increased by 0.10 μmol of C m^2 s^{-1} (Figure 3.4r). Despite similar soil temperature and soil moisture conditions, soil CO_2 effluxes decreased post-harvesting during the growing season, except for F_{LR}, which showed an increase.

For the December to January months, a comparison of 2010 to 2013 two-month diurnal averages showed an increase in soil CO_2 effluxes for all five control chambers, with the largest average increase of 1.25±0.33 μmol of C m^2 s^{-1} seen in F_{S2} (Figure 3.4u and 3.4v). F_{LR} had an increase in the daily average soil CO_2 efflux, with an average of 0.84 μmol of C m^2 s^{-1} in 2013, compared to an average of 0.31 μmol of C m^2 s^{-1} in 2010 (Figure 3.4w). Both F_{RT1} and F_{RT2} had a higher soil CO_2 efflux in 2013, compared with 2010. F_{RT2} increased from an average of 0.57±0.06 μmol of C m^2 s^{-1} in 2010, to an average of 0.82±0.12 μmol of C m^2 s^{-1} in 2013. F_{RT1} had the largest difference between the two years, with an average of 0.37±0.05 μmol of C m^2 s^{-1} in 2010, to an average of 2.58±0.52 μmol of C m^2 s^{-1} in 2013. The difference was attributed to a large amount of CO_2 released at the end of December 2013, which persisted for approximately five days and was comparable to values observed during the growing seasons. Note that the number of observations recorded during the winter months were much less than during the growing season, so these recordings in December 2013 have a stronger influence on the winter average.

Comparison of the soil CO_2 effluxes measured pre- and post-harvesting showed that there was a reduction from each of the chambers after the shelterwood harvest event (Table 3.4). Reduction in total soil CO_2 efflux ranged from 1 to 37%, with the highest difference from F_{S1}, and minimal change in F_{S3} measurements. The chamber with the litter-layer removed (F_{LR}) had a 2% reduction, while both measurements from the trenched chambers (F_{RT1} and F_{RT2}) decreased by approximately 27%. A significant difference for all but two of the chambers were shown for the normalized data (Table 3.4). Both F_{S3} and F_{LR} showed no significant difference between pre- and post-harvesting periods (Table 3.4).

To assess the impact of shelterwood harvesting on overall carbon budget of the forest, an analysis of net ecosystem productivity (NEP) using the eddy-covariance flux data from the nearby flux tower was conducted following methods outlined by Trant (2014). Using data from 2008 to 2014, as compared to the 2003 to 2012 for the analysis by Trant (2014), showed that thinning, following harvesting, did not significantly impact the response of NEP. Post-harvesting NEP fluxes were within the range of interannual variability over the study period (2008 to 2014). Mean pre-harvesting NEP was 237±81 g C m^{-2} $year^{-1}$ as compared to post-harvesting NEP of 185±97 g C m^{-2} $year^{-1}$. Since NEP is a difference of two large fluxes, Gross Ecosystem Productivity (GEP) and ecosystem respiration (R_E), the differences between these components were also considered. Mean annual post-harvesting (2012 to 2014) GEP was 1519±78 g C m^{-2} $year^{-1}$ as compared to pre-harvesting (2008 to2011) GEP of 1479±74 g C m^{-2} $year^{-1}$. In contrast, mean annual post-harvesting total R_E (1326±55 g C m^{-2} $year^{-1}$) was also higher than pre-harvesting R_E (1233±152 g C m^{-2} $year^{-1}$). Based on the results, there was no significant impact on the flux measurements following harvesting, however, both the GEP and R_E fluxes increased after the shelterwood harvesting, while NEP fluxes decreased.

3.4 DISCUSSION

Soil temperature and soil moisture have been shown to be a major influence on the soil CO_2 efflux dynamics (Borken et al., 2002; Monson et al., 2006; Ryu et al., 2009). Results from this study were in agreement with these studies as soil CO_2 efflux followed the seasonal pattern of soil temperature, with the lowest CO_2 effluxes observed during the winter, and highest during the growing season. Observed half-hourly soil CO_2 efflux rates showed diurnal fluctuations that mimic the diurnal pattern of soil temperature, while responding to increases in soil moisture. Pulses of soil CO_2 efflux were observed during the winter of 2013/2014, with measured values similar to those observed during the growing season. These pulses corresponded to an increase in soil temperature and melting of the snow cover. The formation of dense snow and/or ice lenses in a snowpack can impede the upward diffusion of soil CO_2 efflux (Monson et al., 2006' Winston et al., 1997). Monson et al. (2006) observed a similar increase in soil CO_2 efflux due to meltwater, from the above snow cover layer, and an increase in soil temperature, which increased the activity of soil microbes during spring for a subalpine ecosystem. DeForest et al. (2009) observed pulses in soil respiration following full leaf senesce that corresponded with an input of fresh litter decay. This pattern did not occur annually and may be attributed to a warmer season. In relation to the possibility of warming soils, these pulses could have implications on the cycling of nutrients released from the decomposition of the litter-layer (DeForest et al., 2009). Improvement of collecting robust and accurate soil CO_2 efflux data, during the winter season, would help in the understanding the response of soil CO_2 efflux to winter environmental factors, such as snow and ice accumulation and freeze/thaw events.

Forest management practices, such as shelterwood harvesting, produces disturbances that alters the physical structure, microclimate, physiochemical properties, and biological activity of forest soils and vegetation, affecting soil respiration (Peng et al., 2008; Ryu et al., 2009; Selig et al., 2008; Sullivan et al., 2008; Tang et al., 2005; Wang et al., 2013). Mixed results (Concilio et al., 2005; Concilio et al., 2006; Ma et al., 2004; Ohashi et al., 1999; Peng et al., 2006; Sullivan et al., 2008; Tang et al., 2005; Wang et al., 2013) highlights the need for comprehensive and long-term studies to explore the impact of shelterwood harvesting, which is being increasingly adopted to manage forest ecosystems, on soil CO_2 efflux and the overall carbon budget. This study is an important step forward in this direction, where soil CO_2 efflux measurements were conducted, over the seven-year period (2008 to 2014), in a managed temperate forest.

Comparison of soil temperature and soil moisture, between pre- and post-harvesting periods, showed no significant difference, although there was a small (0.3±7.5°C) decrease in soil temperature, while soil moisture showed a small increase (0.01±0.03 m^{-3} m^{-3}) following post-harvesting. Despite similar soil microclimates, total soil CO_2 effluxes were significantly reduced by up to 37%. Soil CO_2 effluxes from heterotrophic sources were significantly reduced post-harvesting by approximately 27%, while no significant difference in the mineral-soil horizon sources were measured. Although root respiration may decrease with a decrease in root biomass post-harvesting, root respiration from the remaining trees may also increase as a result of increased photosynthetic rate and growth of new roots due to a decrease in competition for water and nutrients among remaining trees (Tang et al., 2005). In the thinned stands, dead roots from harvested trees can also contribute more soil CO_2 efflux after the treatment (Tang et al., 2005). In contrast, Shabaga et al. (2015) found that partial harvesting, in northern mixed deciduous forests, increased soil respiration due to an increase in soil temperatures, and enhanced decomposition rates of harvested residue, such as leaves/needles and woody debris left on ground, and the decomposition of dead roots. They also reported that post-harvesting; forest understory re-growth contributed to a reduction in soil temperatures, but influenced patterns of soil CO_2 efflux by increasing summer respiration and Q_{10} values, likely due to increased root respiration. Wang et al. (2013) found no change in soil respiration between control and thinned larch plantation sites, however, the distribution in different components were altered, as both heterotrophic and autotrophic respiration were higher at the thinned site, whereas litter decomposition was lower. Wang et al. (2013) attributed the higher levels of heterotrophic respiration from the mineral-soil to a change in soil temperature, something not found in this study.

As suggested by Sullivan et al. (2008), a possible reason for the lower soil CO_2 efflux can be explained by a decline in live tree root respiration being greater than a possible increase in soil CO_2 efflux, resulting from an increase in root production by surviving trees, and a possible increase in heterotrophic respiration associated with the decomposition of newly dead roots and understory plant litter. The shelterwood harvest may have altered the soil environment by changing amounts and sources of below-ground carbon for microbial metabolism, thus reducing soil CO_2 efflux (Sullivan et al., 2008). Since this study only examined the short-term effects of shelterwood harvesting, the increase in quantity of high quality carbon available for microbial decomposition, usually associated with this harvesting, may not yet be available for decomposition (Sullivan et al., 2008). Tang et al. (2005) found that soil respiration decreased by 13% post-thinning at their site, for data normalized with soil temperature and soil moisture. They note that due to natural climate variability and the effect of thinning on soil temperature and moisture, no significant trends were found in actual non-normalized soil respiration post- thinning. They suggested that the net effects of thinning on soil respiration are the result of complex and numerous interacting factors. Tang et al. (2005) argued that apart from thinning impacts, that changes soil temperature, soil moisture and biotic factors, soil CO_2 efflux in a thinned forest may also be influenced by climate variability (such as drought in 2012 for this site).

An analysis of the eddy-covariance flux data, specifically net ecosystem productivity (NEP), gross ecosystem productivity (GEP) and ecosystem respiration (R_E) from the nearby flux tower showed that harvesting had no significant impact on the response of NEP, while both GEP and R_E increased post-harvesting. Previous work at this site by Nicholas (2011), over a three-year period (2008 to 2010), showed that the annual soil CO_2 efflux values totaled approximately 92% of the annual ecosystem respiration. Since there was a reduction in soil CO_2 efflux post-harvesting while both GEP and R_E increased, it appears that the surge in R_E was primarily caused by the increase in autotrophic respiration due to the enhancement in forest growth. An increase in R_E can also be attributed to a disparity between flux footprints for the eddy-covariance fluxes, which encompass the whole stand, as compared to a localized area where the soil CO_2 efflux chambers were installed. Soil CO_2 efflux chambers often overestimate soil respiration as compared to R_E measured by the eddy-covariance system (Braendholt et al., 2017). This study highlights the complexities and challenges associated with soil CO_2 efflux studies and the need for comprehensive and long-term field studies, to better understand the response of these fluxes to a change in the future climate.

3.5 CONCLUSION

Silvicultural practices, such as shelterwood harvesting, can have a profound impact on forest carbon dynamics by changing soil temperature, and soil water content, soil organic matter, root biomass and microbial activity. These silvicultural practices are being adopted in North America and across the world, not only to manage forests, particularly planted stands, but also to enhance their carbon sequestration capabilities, and/or to restore them to their previous native species, such as mixed woodland in the Great Lakes region. By comparing soil CO_2 effluxes between pre- and post-harvesting years, this study showed that soil temperature and soil moisture did not show any significant changes post-harvesting compared to pre-harvesting. Despite similar climate conditions, soil CO_2 effluxes post-harvesting were lower than pre-harvesting. Further collection of soil CO_2 efflux measurements as the forest re-establishes are needed to determine the full effect of the shelterwood system on soil respiration, and particularly, the impact to heterotrophic and autotrophic respiration following post-harvesting. Results presented in this study are useful for forest management practitioners, specifically those focused on carbon sequestration and forest conservation. The knowledge gained from this research will help in developing polices for better management of forest ecosystems in eastern Canada, in addition to assisting in planning and developing realistic strategies to offset fossil fuel CO_2 emissions to improve environmental quality.

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CHAPTER 4

EVALUATING THE RESPONSE OF SEASONAL AND ANNUAL DYNAMICS OF SOIL CO_2 EFFLUX TO ENVIRONMENTAL FACTORS, INCLUDING LOW FREQUENCY CLIMATE OSCILLATIONS AND FOREST MANAGEMENT, IN A TEMPERATE PINE FOREST

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Abstract: Climate variability and forest management practices influence the rate of soil CO_2 efflux; however, their combined effects are complex and not well understood. This study investigated the response of soil CO_2 efflux to seasonal and interannual climate variability and shelterwood harvesting in a mature wine pine (*Pinus strobus L.*) forest, near Lake Erie in southern Ontario, Canada. By developing a Gaussian-Gamma specification model, the particular focus was to explore how major components of soil CO_2 efflux, such as autotrophic and heterotrophic respiration, and soil CO_2 efflux from the mineral-soil horizon, may respond directly and collectively to soil temperature, soil moisture, shelterwood harvesting and climate oscillations. Soil CO_2 effluxes were highest when soil temperatures were between 16 and 23°C, and while the soil moisture content was between 0.08 and 0.15 m^3/m^3. Soil CO_2 efflux was found to be the major contributor to the total forest carbon emissions, contributing, on average, 78±9% to the annual total ecosystem respiration (R_E) of the forest. It was estimated that the shelterwood harvesting treatment reduced the heterotrophic respiration contribution to total soil CO_2 efflux from 80±6% (2008 to 2011) to 53±21% (2012 to 2014), while increasing autotrophic respiration contribution from 20±5% to 47±21%, with minimal change to the contribution from the mineral-soil horizon. By incorporating climate oscillations that prevail in the region, and their active years into the model, the relationship between soil CO_2 efflux and environmental factors were improved. Results from this study will aid to better understand the influence of climate variability, affected by local and large-scale

climate oscillations, and shelterwood harvesting have on soil CO_2 effluxes to more accurately estimate soil CO_2 effluxes to develop polices for better forest management focused on carbon sequestration and forest conversation.

4.1 INTRODUCTION

Soil carbon dioxide (CO_2) efflux is a major component of the carbon cycle in forest ecosystems (Schlesinger and Andrews, 2000). It is comprised of CO_2 efflux from the litter-surface, autotrophic respiration, by living roots, and heterotrophic respiration from the decomposition of soil organic matter by soil microbes (Elberling, 2007). Several studies suggest that the forest carbon cycle, including soil respiration, is strongly influenced by climatic variability (Bastos et al., 2013; Heimann and Reichstein, 2008; Wharton et al., 2012). An increase in the length of the growing season from warmer seasonal temperatures (Piao et al., 2008) is expected to increase carbon assimilation, with more CO_2 fertilization and higher nitrogen deposition (Black et al., 2000; Reay et al., 2007). Extreme climatic events, such as droughts, flooding, heatwaves, duration and timing of the winter and freeze/thaw periods, are defined as an occurrence of an unusual climate period that alters the ecosystem structure (Reichstein et al., 2013) and are expected to severely affect the forest carbon cycle (Amiro et al., 2010; Davis et al., 1997; Granier et al., 2007; Reichstein et al., 2013; Vose et al., 2012; Wullschleger and Hanson, 2006). With increasing atmospheric CO_2 levels, forest ecosystems may sequester more carbon into the vegetation and soils (Norby et al., 2005; Thomas et al., 1996), or release more carbon from the soils through an increase in root density and higher below-ground allocation of carbon (Hungate et al., 1997; Luo et al., 1996).

Soil CO_2 efflux, also referred to as soil respiration, is influenced by many factors, such as soil temperature, soil moisture, soil pH, soil organic matter, nutrient availability, fine root biomass, vegetation type and microbial activity (Ryu et al., 2009). Soil temperature and soil moisture are the major controls and explain most of the temporal and spatial variations in soil respiration (Borken et al., 2002). As temperatures rise, gross ecosystem productivity (GEP) and soil organic matter decomposition rates increase, producing more soil CO_2 efflux (Lloyd and Taylor, 1994). Soil respiration rates are generally lower under dry conditions due to lower root respiration and diminished microbial activity (Londo et al., 1999). Therefore, monitoring soil respiration, in conjunction with physical soil properties and environmental parameters, are important to understand their response to climate variability and forest management practices (Peng et al., 2008).

In North America, low frequency climate oscillations, such as the Pacific Decadal Oscillation (PDO), the Pacific-North American (PNA) teleconnection pattern, the El Niño-Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), the Arctic Oscillation (AO) and the Eastern Pacific Oscillation (EPO) are important atmospheric phenomenon that influence seasonal and annual climate variations (Bonsal et al., 2001; Shabbar et al., 1997; Shabbar and Bonsal, 2004;

Zhang et al., 2011). Time scales of these climate oscillations range from interannual to interdecadal, and their influence are typically stronger during the winter, but can persist into the spring and summer. Long-term (over 50 years) climate data from the Great Lakes region is strongly influenced by low frequency climate oscillations (Thorne and Arain, 2015; Chapter 2). They found that mean winter temperatures were positively correlated to the NAO and the AO, while negatively correlated with the EPO. Total winter precipitation was primarily influenced by the PNA and the AO, while snowfall was found to have a strong negative relationship with the PNA and the ENSO events. Additionally, oscillations have shown to influence ecosystem respiration (R_E) at this forest site during the winter and spring (Thorne and Arain, 2015; Chapter 2). Winter and spring temperatures, and precipitation, may also influence soil respiration dynamics during the summer, where enhanced water input contributes to groundwater recharge and increased soil moisture levels could reduce the likelihood of a late-summer water stress (Malhi et al., 1999). Therefore, low frequency climate oscillations may have an impact on soil CO_2 effluxes, and overall carbon balance, in forest ecosystems in North America.

Silvicultural practices, such as shelterwood harvesting, can also impact forest carbon dynamics by altering soil temperature, soil water content, soil organic matter, root biomass and microbial activity (D’Amato et al., 2011; Gustafsson et al., 2012; Magruder et al., 2013; Peng et al., 2008; Powers et al., 2009; Vesala et al., 2005). The aim of these silvicultural practices is to enhance carbon sequestration capabilities of a forest and/or to restore planted stands to their native species (Gustafsson et al., 2012).

In the Great Lakes – St. Lawrence Region of Canada, the Ontario Ministry of Natural Resources and Forestry (OMNRF) uses a number of silvicultural practices based on the characteristics of the current forest, its history and the desired forest condition. One such practice is a specific shelterwood harvest, which consists of selectively removing mature trees, while maintaining legacy trees (largest, best quality), in a series of two or more partial cuts, over the course of 10 to 30 years (OMNRF, 1998; OMNRF, 2000). Shelterwood harvesting encourages natural regeneration and development of seedlings in partial shade (OMNRF, 1998; OMNRF, 2000; Trant, 2014). By changing the tree spacing, number and size distribution, shelterwood harvesting can affect the dynamics of carbon sequestration and soil respiration processes (Marchi et al., 2018; Vesala et al., 2005). Additionally, open-canopy alters the radiation dynamics, interception of precipitation, wind flow patterns, water vapour deficit, soil temperature, and soil moisture (Stoffel et al., 2010; Vesala et al., 2005; Woods et al., 2006). Shelterwood harvesting, or selective thinning, also changes root density and production affecting autotrophic respiration, litterfall input and hence soil microbial activity, which can lead to large variations in soil respiration (Peng and Thomas, 2006).

Peng et al. (2008) suggested that amongst the studies that examined forest harvesting impacts on soil respiration, only a few have examined selective harvesting, with little information on temporal and spatial patterns of soil CO_2 efflux responses, particularly in conifer stands. Additional information is required to better understand the effects of shelterwood harvesting on soil processes and soil carbon sequestration. This is important because an increasing number of temperate conifer forests in eastern North America are being managed to enhance their carbon sequestration capabilities to restore them to their native species composition, enhance biodiversity and conserve water resources. Therefore, there is an urgent need to enhance our understanding of the effects of silvicultural practices, in particular shelterwood harvest, on soil processes and carbon sequestration (Jandl et al., 2007).

This study investigated the response of soil CO_2 efflux to the combined effects of climate variability, including those caused by climate oscillations, and shelterwood harvesting in a mature white pine (*Pinus strobes L.*) forest, located near Lake Erie in southern Ontario, Canada from 2008 to 2014. This study used an automated soil CO_2 chamber system, which measures continuous (integrated over a time scale of 30 minutes for this study) soil CO_2 efflux to capture rapid changes in seasonal variations and environmental controls on soil CO_2 efflux (Cai et al., 2010; Drewitt et al., 2002; Gaumont-Guay et al., 2006a; b; Janssens et al., 2000; Jassal et al., 2005; Tang et al., 2005). The use of the automated soil CO_2 chamber system provided a much more reliable estimate of seasonal and annual soil CO_2 efflux, as compared to conventional manual chambers that have low temporal resolution (Bolstad et al., 2004; Khomik et al., 2010). A previous study at the site showed that despite similar soil microclimate, pre- (2008 to 2011) and post-harvesting (2012 to 2014) total soil CO_2 effluxes were significantly reduced by up to 37% (Chapter 3). Soil CO_2 effluxes from heterotrophic sources were significantly reduced post-harvesting by approximately 27%, while no significant difference in the mineral-soil horizon sources were measured. An analysis of R_E, measured with an eddy-covariance tower over the study area, showed an increase post-harvesting. However, the overall net ecosystem carbon exchange showed no significant difference between pre- and post-harvesting (Chapter 3). With these previous results, a Gaussian-Gamma specification model was developed and used to estimate annual total soil respiration and quantify the response of soil CO_2 efflux to soil temperature, soil moisture, climate variability and shelterwood harvesting.

4.2 METHODS

4.2.1 *Study Site*

This site is part of the Turkey Point Flux Station (TPFS), or Turkey Point Observatory of the Global Water Futures (GWF) program, and global Fluxnet. The forest is dominated (>82%) by eastern white pine (*Pinus strobus L.*), which was planted in 1939 to stabilize local sandy soils. As of 2014, the average tree height was about 21.8±1.7 m, and stand density was about 421±166 $\text{stems}\cdot\text{ha}^{-1}$. Other tree species include 11% balsam fir (*Abies balsamea L. Mill*) and species native to the hardwood forests of North American Eastern Temperate Forest Ecoregion, including 4% Oak (*Quercus velutina L.*, *Q. alba L.*), 2% Red Maple

(*Acer rubrum L.*) and some wild black cherry trees (*Prunus serotina Ehrh*). The understory consists of ferns (*Pteridium aquilinum L.*), mosses (*Polytrichum spp.*), poison ivy (*Rhus radicans L.ssp.*) and Rubus species. The topography at the site is fairly flat with well drained sandy soil (Brunisolic Gray Brown Luvisol, following the Canadian Soil Classification system), which is composed of ~98% sand, 1% silt, and 1% clay. Further site details are provided in Arain and Restrepo-Coupe (2005) and Peichl et al. (2010a; b).

4.2.2 Shelterwood harvesting

This forested site is managed by the Ontario Ministry of Natural Resources and Forestry (OMNRF) under the shelterwood silvicultural system (OMNRF, 1998; OMNRF, 2000; Trant, 2014). This system is characterized by two or more partial thinning treatments over a few decades to allow for regeneration and development of seedlings in partial shade. The forest was first thinned in 1983, where one-third of trees were randomly removed. In March of 2012, a second partial cut was conducted and approximately 30% of the dominant overstory trees were selectively removed to improve light and water availability and stimulate growth of the remaining trees. The long-term aim of the management is to convert the pine plantation back to the native stand with mixed-wood ecosystem characteristics (OMNRF, 1998). A mechanical harvester was used to cut, de-limb and section the selected trees. To reduce soil compaction and disturbance, the harvester used pre-existing multi-use trails throughout the stand. Thinning residues (e.g. limbs, bark, crowns) were placed on the ground ahead of the harvester to limit compaction when trails were not accessible (Trant, 2014). After shelterwood harvesting the average basal area was reduced by 13% (Kula, 2014), while peak leaf area index was reduced by 35% (Trant, 2014). Potential effects associated with the exchange of CO_2 between the soil, vegetation and atmosphere is a by-product of shelterwood harvesting, but can be crucial for the development or modification of carbon sequestration polices and influenced by climate change scenarios.

4.2.3 *Soil CO_2 efflux measurements*

Soil CO_2 efflux was measured using automated non-steady state chambers developed by the University of British Columbia (Drewitt et al., 2002; Gaumont-Guay et al., 2008; Jassal et al., 2005; 2012). The soil CO_2 efflux chamber system (chamber system) occupied an area of 50 m × 50 m and was installed approximately 100 m north from the eddy-covariance flux tower at TPFS. The advantage of automated chamber systems to manual measurements is their ability to record continuous long-term measurements and thousands of measurements annually, which is important when assessing short- and long-term responses of soil CO_2 efflux (Jassal et al., 2012). The experimental area was chosen to complement the ecosystem carbon flux measurements from the eddy-covariance flux tower. The individual chamber locations, although few in replicates due to logistical and budgetary constraints, were placed in locations within the footprint of the flux tower measurement, with the objective to provide representative spatial data of soil CO_2 efflux contribution to the overall ecosystem respiration carbon flux. Chambers were situated away from tree stems to avoid large roots, particularly for those chambers measuring heterotrophic respiration. Additionally, the chambers were located away from thick understory and placed on level ground. As soil respiration is comprised of contributions from mineral-soil, organic litter and roots with their associated mycorrhizae, chambers were strategically placed to elucidate the CO_2 contribution from these different components to the total soil respiration.

The chamber system was established to measure total soil CO_2 efflux, including fluxes from root exclusion and forest floor removal plots. The chamber system was installed in June 2008, with the initial installation of four chambers: one chamber to measure heterotrophic respiration (F_{RT1}) in a trenched plot whereby live tree roots were severed in the area surrounding the chamber (Nicholas, 2011); and three chambers used as control plots to measure total soil CO_2 efflux (F_S: F_{S1}, F_{S2} and F_{S22}). One of the control chambers (F_{S22}) was altered in May 2009 by removing the Litter-Fermenting-Humified (LFH) layer and used to measure the contributions from the mineral-soil horizons (F_{LR}) to total soil CO_2 efflux. Additional litter, which settled inside this chamber, was periodically removed, including any material remaining from the thinning event. In May 2009, two more control chambers (F_{S3} and F_{S4}) were added to the chamber system to measure total soil CO_2 efflux. In May 2010 two additional chambers were installed in the chamber system, one used as a control (F_{S5}) and the other to measure heterotrophic respiration (F_{RT2}). At this point eight chambers comprised the chamber system. Vegetation that grew inside the chamber collars were removed between site visits to eliminate potential photosynthesis effect.

Half-hourly soil CO_2 efflux measurements from the control chambers (F_{S1}, F_{S2}, F_{S3}, F_{S4} and F_{S5}) were averaged to determine total soil CO_2 efflux (F_S). Both F_{RT1} and F_{RT2} were averaged to represent heterotrophic respiration (F_{RT}). Autotrophic respiration (F_A) was calculated as the difference between F_S and F_{RT}. The single chamber with the LFH layer removed (F_{LR}) represented the CO_2 efflux from the mineral-soil horizon.

4.2.4 *Data processing and quality control of soil efflux data*

Soil CO_2 efflux data were quality controlled with a custom script written in MATLAB software. In the script, the slope of time versus CO_2 efflux in the high frequency (1 Hz) data was estimated for each measurement interval. This was a process by which half-hourly soil CO_2 efflux data were reviewed to identify and subsequently remove non-representative measurements. First, the root-mean-square-error (RMSE) of the linear relationship of chamber headspace CO_2 mole fractions versus time was calculated. A threshold value of 1 for the ratio of RMSE to the slope of the linear relationship of *dSc/dt* was applied. All values above this threshold were considered unreliable and were removed following a method similar to Jassal et al. (2012). A visual inspection was also conducted to remove any remaining erroneous data. Negative fluxes were discarded as any vegetation, whether by moss or small plants, were removed on the soil surface within the chambers.

In November 2011, the chamber system was temporarily removed prior to the shelterwood harvesting operation and was reinstalled in March 2012. Post-harvesting measurements were recorded from March 2012 until the end of 2014. In addition to data loss during shelterwood harvesting, data loss were also caused by chamber malfunctions, calibrations, the lack of compressed air available, which controls the ability of the chamber dome to close, and winter snow and ice problems associated which ice accumulation in the collar and chamber. Further quality control details for the automatic chamber system are described in Chapter 3.

4.2.5 *Eddy-covariance flux and ancillary data*

Water, carbon, and energy fluxes were measured using a closed-path eddy-covariance system. Flux measurements were made at 20 Hz above the forest canopy on top of a scaffolding tower at a height of at 28 m. Above canopy fluxes of carbon and water were calculated using a methodology outlined by Baldocchi et at. (1988):

$$F = \overline{w'}\overline{c'} \qquad (3.3)$$

where F refers to the flux of energy, carbon or water, w' is the covariance of vertical wind speed, measured by a sonic anemometer (CSAT3, Campbell Scientific Inc. (CSI)), and c' is the covariance of CO_2 or water vapour concentrations, measured by an infrared gas analyzer (Li-7000, LI-COR Inc.). Respiration (R_E) was estimated using a non-linear logistic relationship between nocturnal half-hourly CO_2 fluxes during high turbulence conditions and soil temperature at 2 cm depth (Brodeur, 2014; Peichl et al., 2010a; b). Soil temperature and soil moisture were measured at 5, 10, 20, 50 and 100 cm depths at two locations, near the chamber system location. For analysis, soil temperature at a depth of 5 cm and soil moisture, averaged between 0 and 30 cm, were used. All flux and meteorological data were quality controlled and averaged at half-hourly intervals (Brodeur, 2014). For further details regarding instrumentation, the eddy-covariance tower is described by Arain and Restrepo-Coupe (2005) and Peichl et al., (2010a; b).

4.2.6 *Statistical analysis*

A site-specific model was developed to predict the temporal variability in soil CO_2 efflux using several environmental variables, such as soil temperature and soil moisture, based on the work done by Khomik et al. (2009). Comparing models, such as the Q_{10}, Arrhenius and several empirical models, Khomik et al. (2009) determined the Gamma specification model provided the best estimated fit to soil CO_2 efflux data collected using manual chamber measurements. Refining the Gamma specification model (Khomik et al., 2009), a Gaussian-Gamma specification model (GG model) was developed using the soil CO_2 efflux data for this study. This updated model contains the Gaussian specification for soil temperature and the Gamma specification for soil moisture and is expressed as:

$$y_i = e^{\beta_0+\beta_{11}T_i+\beta_{12}T_i^2+\beta_{21}M_i+\beta_{22}\ln(M_i)} + \varepsilon_i \qquad (4.1)$$

where y_i is the expected value of the ith observation of soil CO_2 emissions (μmol of CO_2 $m^{-2}s^{-1}$); β_0, β_{11}, β_{12}, β_{21} and β_{22} are unknown coefficients to be estimated; T_i is the ith observation of soil temperature (°C); M_i is the ith observation of soil moisture (m^3/m^3); and ε_i is the mean of the error term. The coefficient β_0 represents biotic factors related to soil microbes, soil organic carbon content, root biomass and root respiration (Tang et al., 2005). The β_0 coefficient is site-specific and may change after a significant disturbance, such as the shelterwood harvesting, which alters biotic factors.

The GG model allows for a proportional increase in soil CO_2 emissions related to a unit increase in soil temperature, which becomes smaller at higher temperatures. This aspect provided a better fit than the Gamma specification model for temperature adopted by Khomik et al. (2009). For soil moisture, the Gamma specification allowed for the shape of the relationship to be concave-downward and non-symmetric, which captures the effects of soil moisture on the activities of microorganisms and CO_2 diffusion in the soil (Lou and Zhou, 2006).

The relationship between soil temperature, soil moisture and soil CO_2 efflux was tested using a sensitivity analysis, by omitting one of the explanatory variables (e.g. soil temperature or soil moisture) and examining the resulting effects of the output from the GG model. Results indicated a decrease in the correlation between soil CO_2 efflux, and either soil temperature or soil moisture, when one of the explanatory variables was removed. High temperatures generally coincided with low soil moisture levels, and low temperatures generally coincided with high moisture levels. Omitting soil moisture in the GG model caused the predicted soil CO_2 efflux to be overestimated at low temperatures and underestimated at high temperatures. Omitting soil temperature in the GG model resulted in soil CO_2 efflux to be overestimated at low moisture levels and underestimated at high soil moisture levels. Therefore, both explanatory variables need to be included in the GG model for the most accurate results. Soil temperature and soil moisture can be highly correlated variables and prone to multicollinearity, however, analysis showed that this is not an issue for the data analyzed for this study. The data for this study conformed to the indispensability reasoning, where highly correlated explanatory variables will not lead to a serious overlap, and that the omission of either soil temperature or soil moisture will lead to serious distortions in the predicted relationship between soil CO_2 efflux and the remaining explanatory variable.

To incorporate the influence of the shelterwood harvesting, an additional component was added to the GG model (GGTHIN model), shown in Eq. (4.1):

$$y_i = e^{\beta_0+\beta_{11}T_i+\beta_{12}T_i^2+\beta_{21}M_i+\beta_{22}\ln(M_i)+\beta_{31}THIN} + \varepsilon_i \tag{4.2}$$

where β_{31} is an unknown coefficient to be estimated; and *THIN* is a binary variable that is a value of 1 for soil CO_2 efflux estimated for the pre-harvesting years (2008 to 2011) and a value of 0 for the soil CO_2 efflux estimated for the post-harvesting years (2012 to 2014).

To help investigate the combined effects of shelterwood harvesting and climate oscillations on the soil CO_2 efflux, additional components for each climate oscillation were added to the GGTHIN model (GGTOS model), shown in Eq. (4.2):

$$y_i = e^{\beta_0+\beta_{11}T_i+\beta_{12}T_i^2+\beta_{21}M_i+\beta_{22}\ln(M_i)+\beta_{31}THIN+\beta_{41}PDO+\beta_{51}PNA+\beta_{61}SOI+\beta_{71}NAO+\beta_{81}AO+\beta_{91}EPO} + \varepsilon_i \tag{4.3}$$

where β_{41}, β_{51}, β_{61}, β_{71}, β_{81} and β_{91} are the unknown coefficients to be estimated for each climate oscillation index. These climate oscillation indices are created based on the differences between the relative controlling variables for each climate oscillation (e.g. sea-surface temperature, air pressure) and were obtained from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (NAO index and PNA index), the Bureau of Meteorology (ENSO, as the Southern Oscillation Index (SOI)), and the Joint Institute for the Study of the Atmosphere and Ocean (PDO index and EPO index). For all analyses, only winter index values (averaged between November to March) were used as these climate signals are notably stronger during the winter. For this study, the strength of a climate oscillation were defined as follows: positive (negative) moderate to strong phases were defined as those years which were in the upper (lower) quartile of the data; and neutral phases as those years found in the mid-quartile range. *PDO*, *PNA*, *SOI*, *NAO*, *AO* and *EPO* are binary variables with a value of 1 to represent the calendar year (January to December) that coincided with a strong phase (positive or negative) of the climate oscillations and a value of 0 for neutral phases.

In each of the models, a corresponding t-statistic was estimated for any of the included components: soil temperature, soil moisture, the shelterwood harvesting event and climate oscillations. When the degrees of freedom was greater than 50, and the magnitude of the t-statistic is not less than 2.0, then the corresponding coefficient was significantly different from zero, and should not be omitted.

Each model was used to explore the relationship between soil CO_2 effluxes (F_S, F_{RT} and F_{LR}) and the environmental variables for the growing season (April to October), winter (November to March) and annual (January to December) time scales, and to gap-fill missing soil CO_2 effluxes in the time series using the statistical program SAS v9.4.

4.3 RESULTS

4.3.1 *Seasonal variations in soil temperature and soil moisture*

During the study, the site experienced a large seasonal variation in air temperature, reaching temperatures above 30°C in the summer, and well below 0°C in the winter. This variation was also reflected in the soil temperatures measured at 5 cm depth, but less pronounced.

Daily soil temperature fluctuated around 0°C in the winter, increased by April, peaked in July or August, and then decreased again by October (Figure 4.1). The mean daily soil temperature for the seven-year study period from 2008 to 2014 was 9.3±7.5°C, with the highest soil temperature of 23°C recorded during the summer of 2013, and the lowest of -2.9°C recorded during the winter of 2012. However, in 2012, above-normal temperatures occurred in March; increasing soil temperatures well above 10°C for approximately twenty days.

Daily soil moisture showed large variations throughout study (2008 to 2014), which is attributed to rainfall and/or snowmelt events, and influenced by the well-drained sandy soil at the site. Annually, soil moisture increased during the winter season, with maximum values observed after snowmelt (Figure 4.1b). Little variability in soil moisture was observed during the 2012 growing season, as very few rainfall events occurred, and this year was deemed a drought year (Skubel et al., 2015). The mean soil moisture over the seven-year study period was 0.12±0.03 m^3/m^3, with the highest value of 0.25 m^3/m^3 recorded during the spring of 2013 due to a combination of snowmelt and precipitation. In each growing season, soil moisture levels in this well-drained soil declined to approximately 0.065 m^3/m^3.

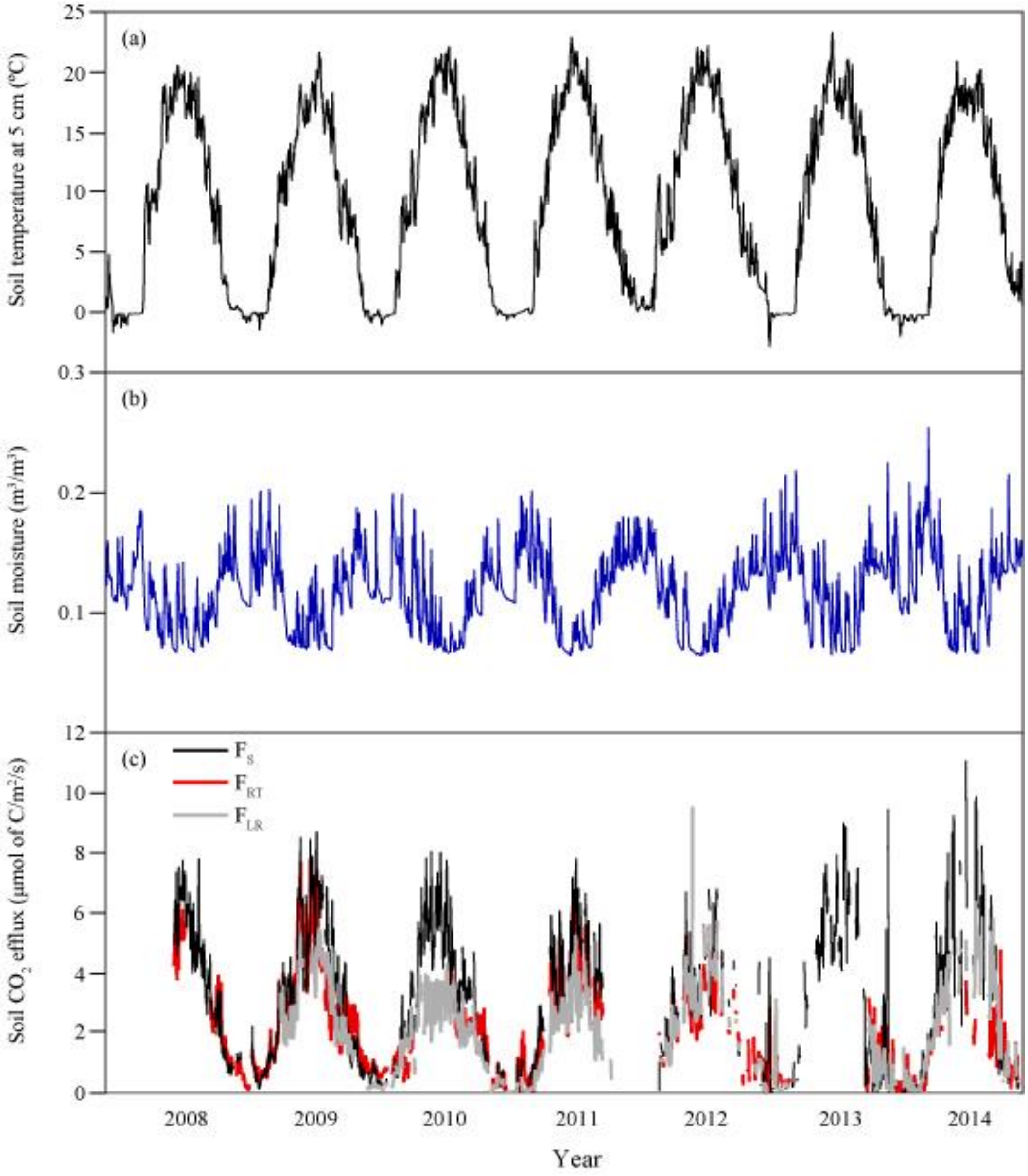


Figure 10. Daily mean (a) soil temperature at 5 cm depth, (b) soil moisture averaged between 0 and 30 cm depths, and (c) soil CO_2 efflux from control (F_S), root trenched efflux (F_{RT}) and mineral-soil efflux (F_{LR}) chamber plots, located in the automated chamber system, from 2008 to 2014. Data gaps in (c) are the result of quality control and equipment malfunction (Section 4.2.4).

4.3.2 *Seasonal variability of soil CO_2 efflux*

Soil CO_2 efflux showed distinct seasonal dynamics during the seven-year study (2008 to 2014); following the seasonal pattern of soil temperature (Figure 4.1c). Soil CO_2 effluxes were lowest during the winter and highest during the summer. Large and small variability in soil CO_2 effluxes occurred during the growing and winter seasons, respectively. Spring and autumn soil CO_2 effluxes were close to zero due to cooler soil temperatures and higher soil moisture levels. As the spring progressed into summer, soil CO_2 effluxes increased responding to increases in soil temperature and rainfall events. In autumn, soil temperatures decreased and soil moisture levels increased due to more rainfall events and decreasing evaporation levels. Soil CO_2 efflux decreased to minimal levels with the onset of winter.

Temporal trend in soil CO_2 effluxes compared to soil temperature and soil moisture, showed diurnal fluctuations that mimic the diurnal pattern of soil temperature, while responding to increases in soil moisture (Figure 4.2). An example from June 2009 showed that soil CO_2 effluxes from the control (F_S), heterotrophic (F_{RT}) and mineral-soil horizon (F_{LR}) chambers increased with a rise in both soil temperature and soil moisture. However, on June 17, 2009, the soil CO_2 efflux from the mineral-soil component decreased at the onset of a rainfall event. Although not common, other instances of a decrease in the soil CO_2 efflux from the mineral-soil component occurred in the following years.

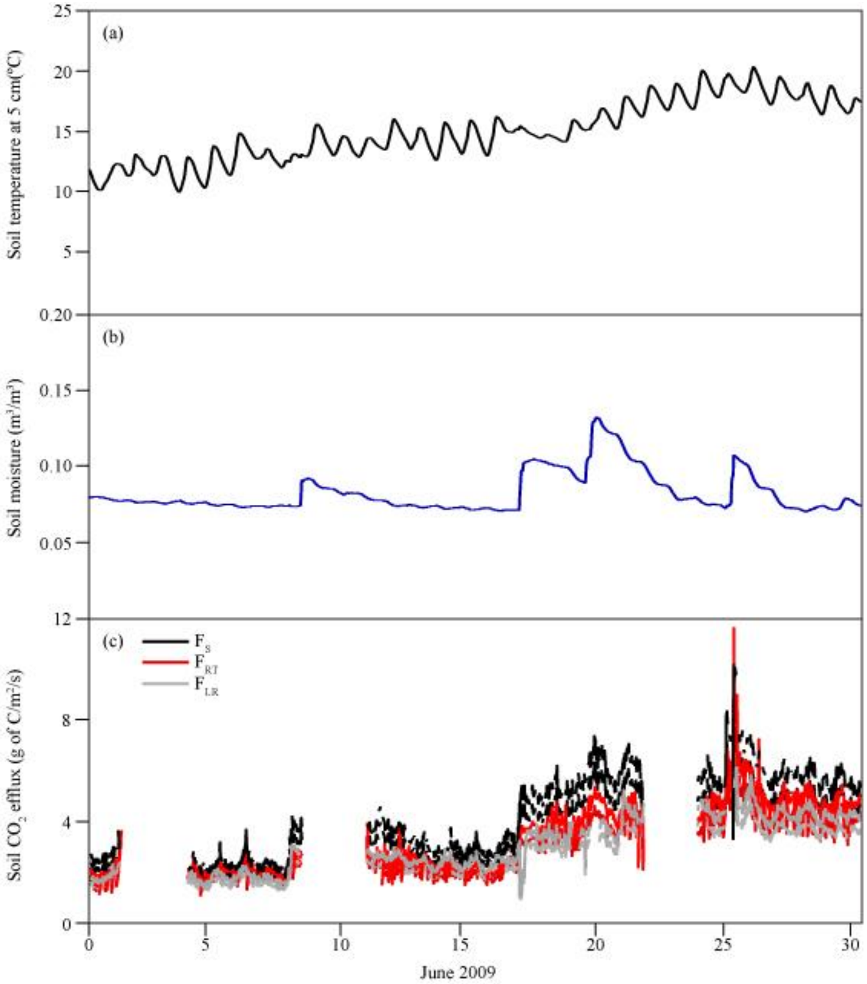


Figure 11. Daily mean (a) soil temperature at 5 cm depth, (b) soil moisture averaged between 0 and 30 cm depths, and (c) total soil CO_2 efflux (F_S), heterotrophic efflux (F_{RT}) and mineral-soil efflux (F_{LR}) for June 2009. Data gaps in (c) are the result of quality control and equipment malfunction (Section 4.2.4).

In January and December in 2013, large pulses of soil CO_2 effluxes were measured, which periodically exceeded values recorded during the growing seasons (Figure 4.1c). Meteorological data indicated these soil CO_2 efflux pulses corresponded to an increase in soil temperature transitioning from below to above freezing, and an increase in soil moisture. These infrequent observations could also be associated to the melting of dense snow or ice on the surface of the soil, or ice in the shallow layers of the soil within the chambers. This snow or ice may have acted as a barrier preventing entrapped CO_2 from being released causing a pulse of soil CO_2 efflux (Chapter 3).

4.3.3 *Response of soil CO_2 efflux to soil temperature and soil moisture*

A site-specific GG model (Equation 4.1) incorporating soil temperature and soil moisture variables, was used to explore the response of soil CO_2 efflux for each of the three components of the soil CO_2 efflux (F_S, F_{RT} and F_{LR}) during the growing season, winter and annually.

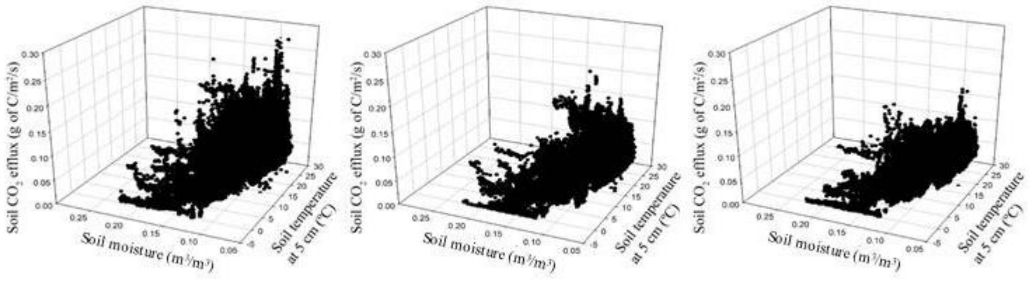
Three-dimensional relationships between soil CO_2 efflux, soil temperature and soil moisture, indicated that soil CO_2 effluxes were higher for temperatures between 16 and 23°C, and soil moisture between 0.08 to 0.15 m^3/m^3 (Figure 4.3a). Lower soil CO_2 efflux were predicted for temperatures below 5°C and soil moisture between 0.10 to 0.30 m^3/m^3 during the growing season (Figure 4.3a). F_S and F_{RT} had similar positive relationships with soil temperature and soil moisture.

F_{LR} showed a gradual increase in soil CO_2 effluxes with increasing soil temperature and decreasing soil moisture.

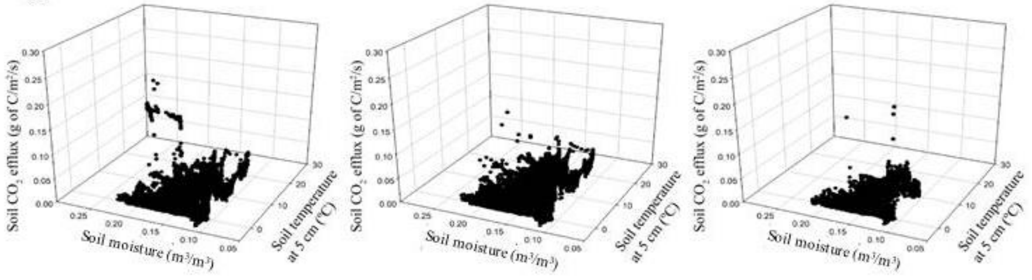
For the winter, soil CO_2 effluxes were lower than those measured during the growing season (Figure 4.3b). Changes in soil moisture had little influence on soil CO_2 efflux. Of the three components of soil CO_2 effluxes, F_{LR} had the lowest measured soil CO_2 efflux.

The relationship between soil CO_2 efflux, soil temperature, and soil moisture annually were similar to the growing season (Figure 4.3c). Soil CO_2 effluxes increased as soil temperatures rose and was higher at soil moisture between 0.08 to 0.15 m^3/m^3.

(a) Growing season



(b) Winter



(c) Annually

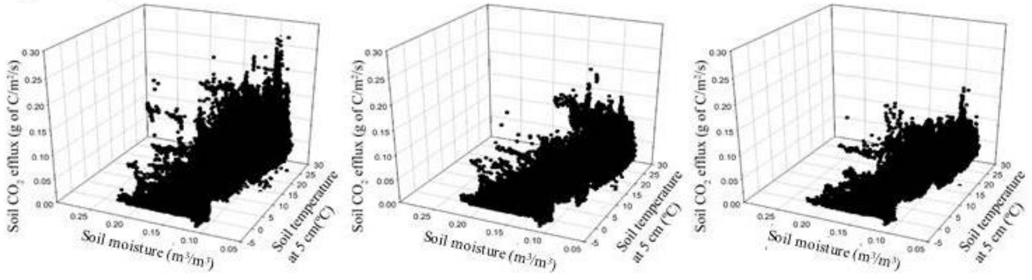


Figure 12. Three-dimensional relationships of soil CO_2 efflux (F_S), root trenched efflux (F_{RT}) and mineral-soil horizon efflux (F_{LR}) modelled against soil temperature at 5 cm depth, and soil moisture averaged between 0 and 30 cm depths, for the (a) growing season (April to October), (b) winter (November to March) and (c) annually (January to December).

4.3.4 *Exploring response of soil CO_2 efflux to environmental variables with the Gaussian-Gamma specification models*

To explore influences from shelterwood harvesting and climate oscillations, in addition to environmental factors (soil temperature and soil moisture), soil CO_2 effluxes were estimated (F_S, F_{RT} and F_{LR}) using the three Gaussian-Gamma specification models: GG (Equation 4.1), GGTHIN (Equation 4.2) and GGTOS (Equation 4.3).

Over the growing season, soil CO_2 effluxes had high correlations with soil temperature (adjusted R-square of 0.525, 0.583 and 0.564 for F_S, F_{RT} and F_{LR}, respectively) and a small correlation with soil moisture (adjusted R-square of 0.179, 0.159 and 0.281 for F_S, F_{RT} and F_{LR}, respectively) (Table 4.1). Including both soil temperature and soil moisture against soil CO_2 efflux, increased the correlation, with an adjusted R-square of 0.680, 0.736 and 0.578 for F_S, F_{RT} and F_{LR}, respectively. The addition of the shelterwood harvesting component increased the adjusted R-square values (0.719, 0.744 and 0.605 for F_S, F_{RT} and F_{LR}, respectively). Incorporating climate oscillations yielded the highest adjusted R-square values of 0.786, 0.774 and 0.610 for F_S, F_{RT} and F_{LR}, respectively.

Table Error! No text of specified style in document.**1.** Correlation using different Gaussian-Gamma specification models with half-hourly measurements from total soil CO_2 efflux (F_S), root trenched efflux (F_{RT}) and mineral-soil horizon efflux (F_{LR}), incorporating one or more of the following components: soil temperature, soil moisture, shelterwood harvesting treatment and climate oscillations. Relationships were examined for the growing season (April to October), winter (November to March) and annually (January to December).

Growing Season (April to October)			
	$\mathbf{F_S}$	$\mathbf{F_{RT}}$	$\mathbf{F_{LR}}$
Soil Temperature	0.525	0.583	0.564
Soil Moisture	0.179	0.159	0.281
Soil Temperature and Moisture (GG model)	0.680	0.736	0.578
Soil Temperature, Soil Moisture and Thinning Treatment (GGTHIN model)	0.719	0.744	0.605
Soil Temperature, Soil Moisture, Thinning Treatment and Climate Oscillations (GGTOS model)	0.786	0.774	0.610
Winter Season (November to March)			
	$\mathbf{F_S}$	$\mathbf{F_{RT}}$	$\mathbf{F_{LR}}$
Soil Temperature	0.680	0.256	0.625
Soil Moisture	0.026	0.024	0.039
Soil Temperature and Moisture (GG model)	0.684	0.257	0.629
Soil Temperature, Soil Moisture and Thinning Treatment (GGTHIN model)	0.686	0.258	0.664
Soil Temperature, Soil Moisture, Thinning Treatment and Climate Oscillations (GGTOS model)	0.712	0.302	0.690
Annually (January to December)			
	$\mathbf{F_S}$	$\mathbf{F_{RT}}$	$\mathbf{F_{LR}}$
Soil Temperature	0.765	0.507	0.570
Soil Moisture	0.346	0.208	0.314
Soil Temperature and Moisture (GG model)	0.820	0.581	0.577
Soil Temperature, Soil Moisture and Thinning Treatment (GGTHIN model)	0.824	0.662	0.584
Soil Temperature, Soil Moisture, Thinning Treatment and Climate Oscillations (GGTOS model)	0.841	0.704	0.639

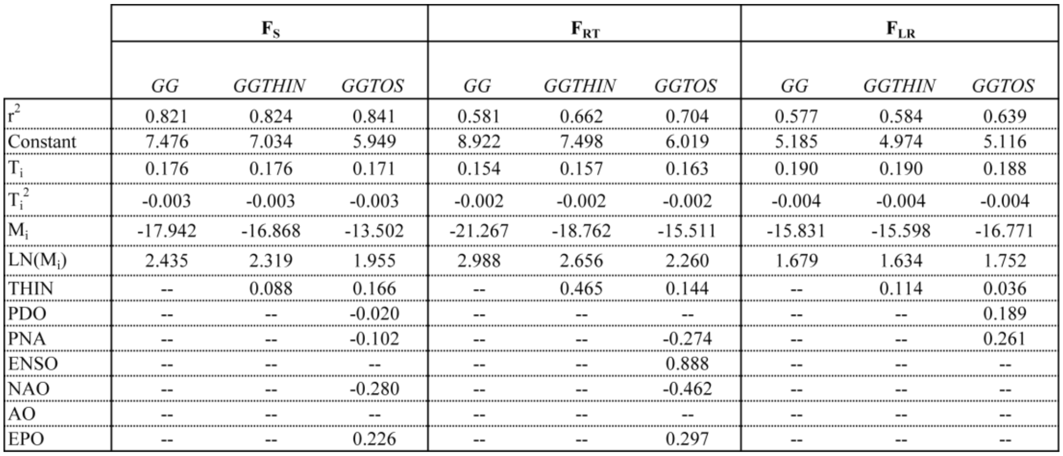
Adjusted R-squared values were lower for the winter, particularly for F_{RT}, which had the lowest values (Table 4.1). Individually, soil CO_2 efflux had the highest adjusted R-squared values with soil temperature (0.680, 0.256 and 0.625 for F_S, F_{RT} and F_{LR}, respectively), and smaller R-squared values with soil moisture (0.026, 0.024 and 0.039 for F_S, F_{RT} and F_{LR}, respectively). Including both soil temperature and soil moisture slightly improved the adjusted R-squared values (0.684, 0.257 and 0.629 for F_S, F_{RT} and F_{LR}, respectively). The GGTHIN model minimally improved the adjusted R-square values (0.686, 0.258 and 0.664 for F_S, F_{RT} and F_{LR}, respectively), with the most notable difference in F_{LR}. The addition of climate oscillations increased the adjusted R-square values to 0.712, 0.302 and 0.690 for F_S, F_{RT} and F_{LR}, respectively. Soil moisture levels had negligible influence on soil CO_2 efflux from the winter season.

Annually, soil CO_2 efflux had high correlation with soil temperature (adjusted R-square of 0.765, 0.507 and 0.570 for F_S, F_{RT} and F_{LR}, respectively), and smaller adjusted R-squared values with soil moisture (0.346, 0.208 and 0.314 for F_S, F_{RT} and F_{LR}, respectively). Incorporating both soil temperature and soil moisture improved the adjusted R-squared values to 0.821, 0.581 and 0.577 for F_S, F_{RT} and F_{LR}, respectively. For the GGTHIN model, adjusted R-squared values increased to 0.824, 0.662 and 0.584 for F_S, F_{RT} and F_{LR}, respectively. The additional of climate oscillation components increased R-squared values to 0.841, 0.704 and 0.639 for F_S, F_{RT} and F_{LR} (Table 4.1).

Comparing the estimated parameter coefficients among the three models (GG, GGTHIN and GGTOS), soil temperature and soil moisture were similar in terms of sign and magnitude (Table 4.2). The addition of the shelterwood harvest event and climate oscillation components increased the adjusted R-squared value for each soil CO_2 efflux component. For the GGTOS model, t-statistics for the climate oscillation components which were less than 2.0 in magnitude were omitted, notably ENSO for F_S, PDO for F_{RT} and ENSO, NAO and EPO for F_{LR}.

Estimated annual total soil CO_2 efflux, and its components, differed among the models (Figure 4.4). Annual total soil CO_2 effluxes from 2008 were omitted as the study began midway through the year. The average total annual soil CO_2 efflux (F_S) predicted among the three models was fairly consistent at approximately 1000 g of C m^{-2} $year^{-1}$ (1017±41, 1000±48 and 994±72 g of C m^{-2} $year^{-1}$ for the GG, GGTHIN and GGTOS models, respectively). The largest total annual soil CO_2 efflux difference was approximately 100 g of C m^{-2} $year^{-1}$ between the GG and GGTOS models in 2012 (Figure 4.4a). Total annual soil CO_2 efflux for F_{RT} and F_{LR} were lower than F_S. From 2009 to 2011, each component had similar annual soil CO_2 effluxes among the models (from 2009 to 2011: F_{RT}, 793±62, 825±50 and 807±79 g of C m^{-2} $year^{-1}$ for the GG, GGTHIN and GGTOS models, respectively, and F_{LR}, 616±26, 620±25 and 618±49 g of C m^{-2} $year^{-1}$ for the GG, GGTHIN and GGTOS models, respectively.

Table Error! No text of specified style in document.**2.** Parameter coefficients and correlation for each model used to estimate annual total soil CO_2 efflux (F_S), heterotrophic respiration efflux (F_{RT}) and mineral-soil respiration efflux (F_{LR}) for: soil temperature and soil moisture (GG model), soil temperature, soil moisture and shelterwood harvesting event (GGTHIN model) and soil temperature, soil moisture, shelterwood harvesting event and climate oscillation components (GGTOS model). For those climate oscillations without a value in the GGTOS model, the associated t-statistics was less than 2.0 in magnitude were removed from the model. Note: PDO = Pacific Decadal Oscillation, PNA = Pacific-North American, ENSO = El Niño Southern Oscillation, NAO = North Atlantic Oscillation, AO = Arctic Oscillation and EPO = Eastern Pacific Oscillation. Dash represents those components that were not included in the model run, or for the GGTOS model, represents a coefficient that did not pass the t-test at the 95% confidence level and was removed from the model.



	F_S			**F_{RT}**			**F_{LR}**		
	GG	*GGTHIN*	*GGTOS*	*GG*	*GGTHIN*	*GGTOS*	*GG*	*GGTHIN*	*GGTOS*
r^2	0.821	0.824	0.841	0.581	0.662	0.704	0.577	0.584	0.639
Constant	7.476	7.034	5.949	8.922	7.498	6.019	5.185	4.974	5.116
T_i	0.176	0.176	0.171	0.154	0.157	0.163	0.190	0.190	0.188
T_i^2	-0.003	-0.003	-0.003	-0.002	-0.002	-0.002	-0.004	-0.004	-0.004
M_i	-17.942	-16.868	-13.502	-21.267	-18.762	-15.511	-15.831	-15.598	-16.771
$LN(M_i)$	2.435	2.319	1.955	2.988	2.656	2.260	1.679	1.634	1.752
THIN	--	0.088	0.166	--	0.465	0.144	--	0.114	0.036
PDO	--	--	-0.020	--	--	--	--	--	0.189
PNA	--	--	-0.102	--	--	-0.274	--	--	0.261
ENSO	--	--	--	--	--	0.888	--	--	--
NAO	--	--	-0.280	--	--	-0.462	--	--	--
AO	--	--	--	--	--	--	--	--	--
EPO	--	--	0.226	--	--	0.297	--	--	--

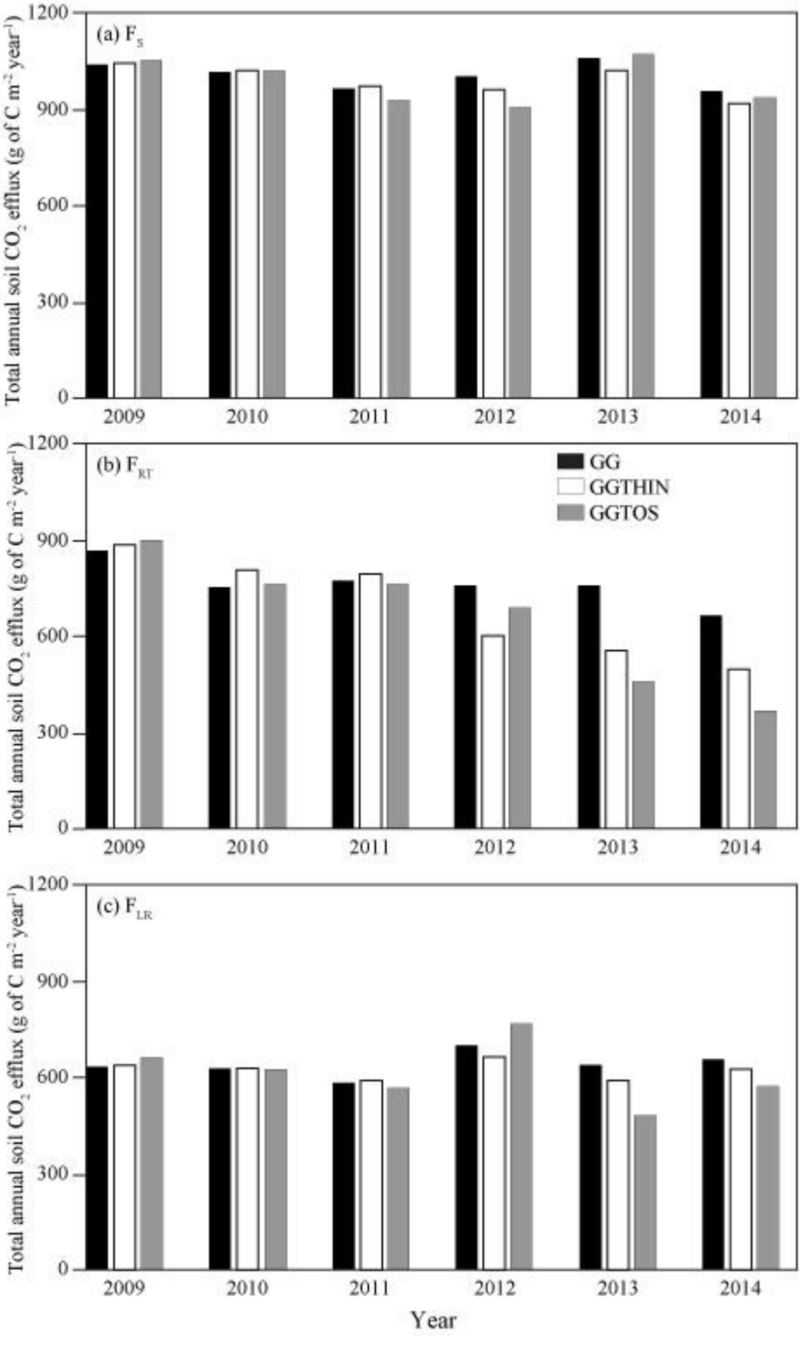


Figure 13. Annual predicted values of total soil CO_2 efflux (F_S), root trenched efflux (F_{RT}) and mineral-soil horizon efflux (F_{LR}), using the GG model, which incorporates soil temperature and soil moisture components, the GGTHIN model which includes an additional shelterwood harvesting component and the GGTOS model, with additional climate oscillation components.

In 2012, the predicted annual total F_{RT} and F_{LR} from the GGTHIN model was lower than those estimated using the GG and GGTOS models (F_{RT}: 753, 598 and 692 g of C m^{-2} $year^{-1}$ for GG, GGTHIN and GGTOS models, respectively; F_{LR}: 702, 667 and 768 g of C m^{-2} $year^{-1}$ for GG, GGTHIN and GGTOS models, respectively). In 2013 and 2014, the predicted annual total of F_{RT} and F_{LR} using the GGTOS model was lower compared with the other two models, despite having higher correlations (Table 4.2). By incorporating climate oscillations, such as PNA, NAO and EPO within the parameter coefficients for the GGTOS model, higher correlations were obtained. In 2013, the AO was the only active teleconnection and a negative coefficient was produced after numerous calibrations in the SAS program. This coefficient drastically reduced predicted soil CO_2 efflux, therefore, this climate oscillation was omitted from the calibrations for the GGTOS model.

Using the GGTOS model (Equation 4.3) with parameters from Table 4.2, the dynamics of total soil CO_2 efflux (F_S), heterotrophic (F_{RT}) mineral-soil horizon (F_{LR}) and autotrophic ($F_A = F_S - F_{RT}$) respiration for the growing season, winter season, and annually during the study period (2008 to 2014) were explored. Totals for 2008 were omitted as data collection began in June. For the growing season, mean CO_2 effluxes of F_S ranged from 749 g of C m^{-2} $season^{-1,}$ (in 2012), to 924 g of C m^{-2} $season^{-1}$ (in 2013), averaging 854±73 g of C m^{-2} $season^{-1}$ (Figure 4.5). Heterotrophic respiration was lower than F_S, and ranged from 295 g of C m^{-2} $season^{-1}$ (2014) to 725 g of C m^{-2} $season^{-1}$ (2009), averaging 531±170 g of C m^{-2} $season^{-1}$. Predicted values from the mineral-soil horizon produced values ranging from 399 g of C m^{-2} $season^{-1}$ (2013) to 635 g of C m^{-2} $season^{-1}$ (2012), averaging 524±86 g of C m^{-2} $season^{-1}$. Using the difference between F_S and F_{RT} to calculate F_A, it was estimated that autotrophic respiration ranged from 189 g of C m^{-2} $season^{-1}$ (2009) to 573 g of C m^{-2} $season^{-1}$ (2013), averaging 323±184 g of C m^{-2} $season^{-1}$. In 2013 and 2014, calculated autotrophic respiration was higher than heterotrophic respiration.

Total soil CO_2 efflux values were lower during the winter compared to the growing season, ranging from 107 g of C m^{-2} $season^{-1}$ (2010/11) to 181 g of C m^{-2} $season^{-1}$ (2011/12), averaging at 142±24 g of C m^{-2} $season^{-1}$. Predicted winter totals for F_{RT} ranged from 97 g of C m^{-2} $season^{-1}$ (2013/14) to 182 g of C m^{-2} $season^{-1}$ (2009/10), averaging at 134±34 g of C m^{-2} $season^{-1}$. The first three winter totals for F_{RT} winter were higher than F_S. Seasonal totals for F_{LR} ranged from 75 g of C m^{-2} $season^{-1}$ (2010/11) to 124 g of C m^{-2} $season^{-1}$ (2011/12). Calculated seasonal F_A were negative for 2008/09 to 2010/11 (-46 to -6 g of C m^{-2} $season^{-1}$) and positive for the last three seasons 2011/12 to 2013/2014 (30 to 40 g of C m^{-2} $season^{-1}$) (Figure 4.5).

The annual totals of soil CO_2 efflux were similar in pattern and magnitude to the growing season (Figure 4.5). F_S totals ranged from 912 g of C m^{-2} $year^{-1}$ (2012) to 1081 g of C m^{-2} $year^{-1}$ (2013), averaging 994±72 g of C m^{-2} $year^{-1}$. F_{RT} totals ranged from 370 g of C m^{-2} $year^{-1}$ (2014) to 898 g of C m^{-2} $year^{-1}$ (2009), averaging 657±201 g of C m^{-2} $year^{-1}$. Calculated F_A totals ranged from 163 g of C m^{-2} $year^{-1}$ (2009) to 621 g of C m^{-2} $year^{-1}$ (2013), and averaged 337±206 g of C m^{-2} $year^{-1}$. F_{LR} totals ranged from 483 g of C m^{-2} $year^{-1}$ (2013) to 768 g of C m^{-2} $year^{-1}$ (2012), and averaged 609±100 g of C m^{-2} $year^{-1}$.

Average annual soil CO_2 efflux values showed that total soil CO_2 efflux and mineral-soil horizon were similar between pre- to post-harvesting periods (1009±65 g of C m^{-2} $year^{-1}$ to 979±90 g of C m^{-2} $year^{-1}$ for F_S and 618±49 g of C m^{-2} $year^{-1}$ to 600±149 g of C m^{-2} $year^{-1}$ for F_{LR}, respectively). Heterotrophic respiration was higher pre- (807±166 g of C m^{-2} $year^{-1}$) than post-harvesting (507±166 g of C m^{-2} $year^{-1}$). Autotrophic respiration increased from 202±59 g of C m^{-2} $year^{-1}$ to 471±219 g of C m^{-2} $year^{-1}$ post-harvesting

Using estimated annual and seasonal totals, the average percent contribution of heterotrophic (F_{RT}), autotrophic (F_A) and mineral-soil (F_{LR}) effluxes to the total soil CO_2 efflux (F_S) were calculated. For the growing season, F_{RT} and F_A contributed 63±20% and 37±20% to F_S, respectively. F_{LR} contributed 62±13% to F_S. In winter, heterotrophic respiration was the main contributor to total soil CO_2 efflux (95±24%), while F_A contributed 5±24%. F_{LR} contributed 67±5% to total winter soil CO_2 efflux. Annually, F_{RT} and F_A contributions were 66±20% and 34±20% to F_S, respectively, while F_{LR} contributed 62±13% to F_S.

Comparisons of percent contribution, pre- and post-harvesting, showed that heterotrophic respiration was significantly reduced post-harvesting, but this was not manifested in changes in the mineral-soil horizon or total soil CO_2 effluxes. Percent contribution from autotrophic respiration increased post-harvesting. Annually, F_{RT} (F_A) decreased (increased) from 80±6% (20±6%) to 53±20% (47±20%) from pre- to post-harvesting. The largest change in contribution post-harvesting occurred during the growing season, where F_{RT} (F_A) decreased (increased) from 76±7% (24±7%) to 49±22% (51±22%).

4.3.5 *Soil CO_2 efflux contribution to ecosystem respiration*

Ecosystem respiration (R_E) measured from the eddy-covariance flux tower and soil CO_2 efflux measured from the chamber system were compared to explore the contribution from the various components of soil CO_2 efflux towards the total R_E (Figure 4.5).

For the growing season, R_E ranged from 942 g of C m^{-2} $season^{-1}$ (2011) to 1229 g of C m^{-2} $season^{-1}$ (2010) (Figure 4.5a). The average during the study(2009 to 2014, 2008 omitted) was 1115±108 g of C m^{-2} $season^{-1}$, with five of the six years reaching over 1000 g of C m^{-2} $season^{-1}$. Total soil CO_2 efflux contributed a significant amount to R_E, ranging from 63% in 2012 to 88% in 2009, and on average contributed 77±10% to R_E. Heterotrophic respiration contributed, on average, 49±18% to R_E, with these contributions lower post-harvesting (range of 25 to 47%). Autotrophic respiration contributed 29±16% to R_E, with contributions higher post-harvesting. The average contribution from the mineral-soil horizon to R_E was 30±12%.

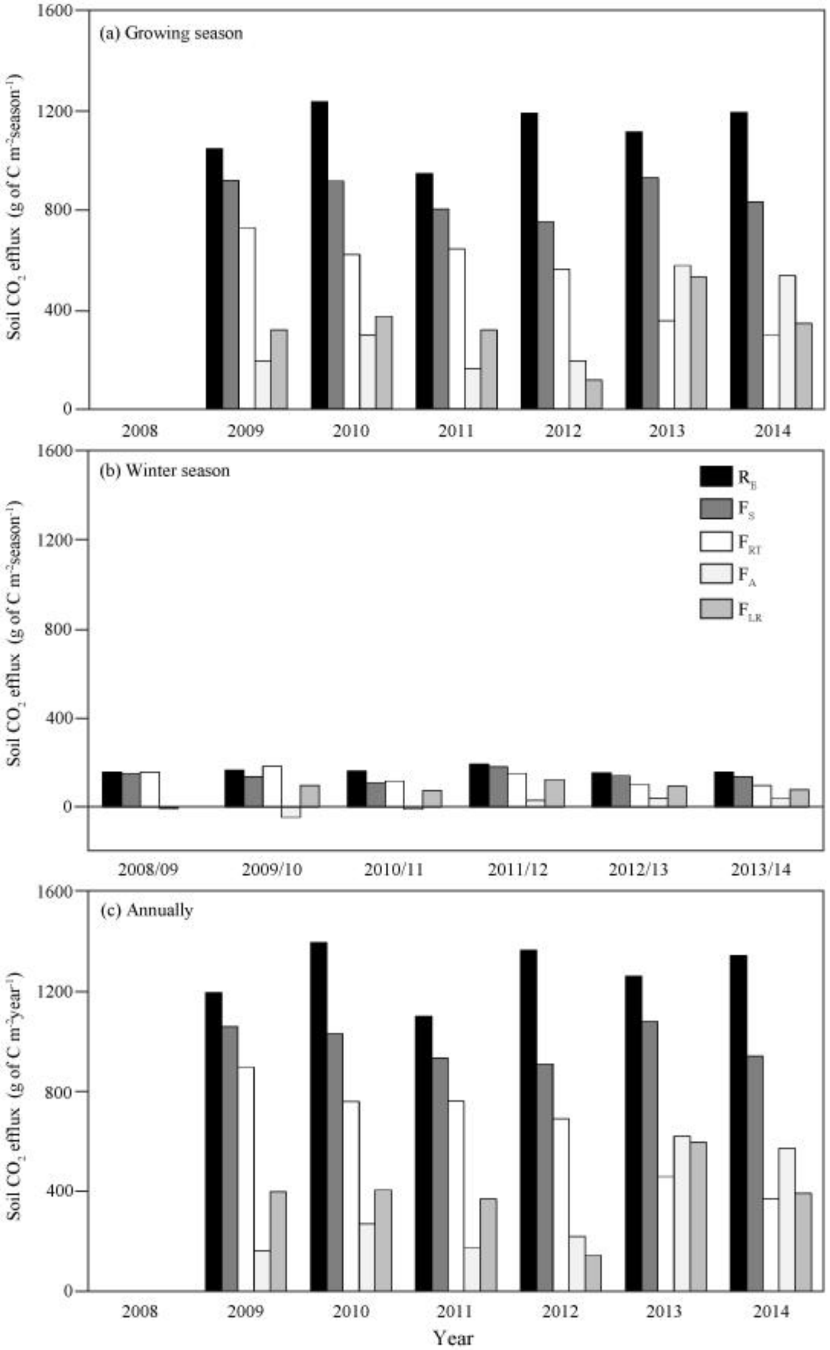


Figure 14. Comparison of ecosystem respiration (R_E), total soil CO_2 (Fs), heterotrophic (F_{RT}), autotrophic (F_A) and mineral-soil horizon (F_{LR}) effluxes for the (a) growing season, (b) winter season and (c) annually from 2008 to 2014.

During the winter season, F_S contributed the majority to total ecosystem respiration (R_E) and ranged from 155 g of C m^{-2} $season^{-1}$ (2012/13) to 193 g of C m^{-2} $season^{-1}$ (2011/12), averaging 165±14 g of C m^{-2} $season^{-1}$ (Figure 4.5b). In the winter, F_S contributed 86±11% to R_E. F_{RT} for the 2009/2010 winter season was higher than R_E and F_S. In 2008/09 and 2010/2011 winter seasons, F_{RT} was greater than F_S. Seasonal average percent contribution of F_{RT} to R_E was 81±19%. Calculated F_A was negative for the first three winter seasons (2008/09 to 2010/11) because F_{RT} was greater than F_S. The last three winter seasons (2011/12 to 2013/14) F_{RT} contributed on average 22±6% to R_E. During the winter, F_A is expected to nearly cease and F_S approximately equaled F_{RT}. The average mineral-soil horizon contribution to R_E was 28±7%.

Annually, R_E ranged from 1102 g of C m^{-2} $year^{-1}$ (2011) to 1399 g of C m^{-2} $year^{-1}$ (2010), averaging 1279±114 g of C m^{-2} $year^{-1}$ (Figure 4.5c). The F_S contributed a large proportion to R_E, with values ranging from 67% (2012) to 89% (2009), averaging 78±9%. The F_{RT} contributed almost half of the respiration to R_E (average of 52±18%). Calculated F_A and F_{LR} contributed an average of 26±16% and 30±12%, respectively to R_E.

Mean annual totals for R_E increased pre- to post-harvesting. The post-harvesting mean annual total R_E (1326±55 g of C m^{-2} $year^{-1}$) was higher than pre-harvesting R_E (1233±152 g of C m^{-2} $year^{-1}$). This manifested in the growing season which increased from 1071±146 g of C m^{-2} $year^{-1}$ for pre-harvesting to

1160±44 g of C m^{-2} $year^{-1}$ post-harvesting. Based on the results, total soil CO_2 efflux was the largest component of R_E, and although there was a reduction in soil CO_2 efflux post-harvesting, there was no significant impact on the flux measurements by the shelterwood harvest.

4.4 DISCUSSION

Results from this study were in agreement with previous studies showing that soil temperature and soil moisture are a major influence on the soil CO_2 efflux dynamics (Borken et al., 2002; Monson et al., 2006; Ryu et al., 2009). An analysis of soil CO_2 efflux, soil temperature and soil moisture showed soil CO_2 efflux followed seasonal patterns of soil temperature, with lower soil CO_2 efflux measured during the winter, and higher measured during the growing season. Soil CO_2 effluxes were also influenced by soil moisture and responded to soil saturation, droughts and rainfall/snowmelt events. At higher soil moisture, soil CO_2 efflux was lower, likely associated with decreases in air porosity in soils and available oxygen for microbes (Tang et al., 2005).

Low frequency climate oscillations, such as the Pacific Decadal Oscillation (PDO), the Pacific-North American (PNA) teleconnection pattern, the El Niño-Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), the Arctic Oscillation (AO) and the Eastern Pacific Oscillation (EPO) are important atmospheric phenomenon that influence seasonal and annual climate variations (Bonsal et al., 2001; Shabbar et al., 1997; Shabbar and Bonsal, 2004; Zhang et al., 2011). Thorne and Arain (2015) (Chapter 2) showed climate in the Great Lakes region is strongly influenced by these low frequency climate oscillations and these influences have shown to affect ecosystem respiration at the forest site during the winter and spring seasons. Forest management practices, such as shelterwood harvesting, produces disturbances that alter the physical structure, microclimate, physiochemical properties, and biological activity of forest soils and vegetation, affecting soil CO_2 efflux (Peng et al., 2008; Ryu et al., 2009; Selig et al., 2008; Sullivan et al., 2008; Tang et al., 2005; Wang et al., 2013).

Gaussian-Gamma specification models were developed to explain the temporal variability in soil CO_2 efflux with soil temperature, soil moisture, the shelterwood harvesting and climate oscillations. Comparison of output and estimated parameter coefficients among the three models (GG, GGTHIN and GGTOS models) found that the addition of the shelterwood harvest event and climate oscillation components increased the adjusted R-squared values for each soil CO_2 efflux component. Based on these results, the GGTOS model was used to explore the dynamics of total soil CO_2 efflux (F_S), heterotrophic (F_{RT}), mineral-soil horizon (F_{LR}) and autotrophic ($F_A = F_S - F_{RT}$) respiration for the growing season, winter season, and annually during the study period (2008 to 2014), and their contribution to ecosystem respiration (R_E) at the forested site.

Soil CO_2 efflux from forest ecosystems is a combination of autotrophic respiration (roots, mycorrhizal fungi and microorganisms in the rhizophere) and heterotrophic respiration (decomposition of litter and soil organic matter) in the organic and mineral-soil horizons (Hanson et al., 2000; Lorenz, 2010). Annual soil CO_2 efflux account for about two-thirds of the ecosystem respiration in forests, with the contribution roughly split between autotrophic and heterotrophic respiration (Hanson, 2000). This study showed that heterotrophic and autotrophic respiration contributed 63±20% and 37±20%, respectively, to soil CO_2 efflux during the growing season. About two-thirds of the total soil CO_2 efflux originated from the mineral-soil horizon (62±13%). Over the winter, heterotrophic respiration was the main contributor to soil CO_2 efflux (95±24%), with autotrophic respiration a minor contributor (5±24%). Annually, heterotrophic and autotrophic respiration components contributed 66±20% and 34±20% to the total soil CO_2 efflux, respectively. The litter-layer contribution for this study was higher than the 25% and 9% found by Wilson et al. (2014) in a Tennessee temperate deciduous forest, and by Ataka et al. (2014) in a warm temperate forest in Japan. However, DeForest et al. (2009) found that the litter-layer contributed 48±12% to the total soil CO_2 efflux at a temperate oak forest in Ohio.

This study showed that soil CO_2 efflux accounted for 78±9% of the annual ecosystem respiration (R_E), derived using eddy-covariance fluxes. Heterotrophic respiration contributed almost half of the respiration to R_E, while autotrophic and mineral-soil respiration contribution averaged 26±16% and 30±12%, respectively. These comparisons of ecosystem respiration to soil CO_2 efflux may differ due to discrepancies from spatial coverages between the chamber system and eddy-covariance flux (Drewitt et al., 2002; Law et al., 2001), difficulties estimating nighttime ecosystem respiration due to stable conditions with the eddy-covariance systems and applying methods to gap-fill these data (Baldocchi, 2008), forest floor heterogeneity of the chamber system at such small scales, and up-scaling the chamber measurements (Gaumont-Guay et al., 2006a; Xu and Qi, 2001).

A comparison of annual soil CO_2 effluxes between pre- and post-harvesting showed a minimal decrease in total soil CO_2 efflux, while heterotrophic (autotrophic) respiration decreased (increased) and mineral-soil respiration were similar post-harvesting. Given that soil CO_2 efflux at this site is the largest component of R_E, and there was a reduction in soil CO_2 efflux post-harvesting, while R_E increased, it appears that surge in R_E was primarily caused by the increase in autotrophic respiration due to enhancement in both the understory and/or growth in remaining trees after the harvesting. Sullivan et al. (2008) suggested shelterwood harvesting altered the soil environment by changing amounts and sources of below-ground carbon for microbial metabolism, thus reducing soil CO_2 efflux. Since this study only examined the short-term effects of shelterwood harvesting, the increase in quantity of high quality carbon available for microbial decomposition, usually associated with this harvesting, may not yet be available for decomposition (Sullivan et al., 2008). Tang et al. (2005) found that soil respiration decreased by 13% after thinning and noted that due to natural climate variability, and the effect of thinning on soil temperature and moisture, no clear trends were found in soil respiration values after thinning as a result of numerous complex interacting factors. Mixed results (Concilio et al., 2005; Concilio et al., 2006; Ma et al., 2004; Ohashi et al., 1999; Peng et al., 2006; Sullivan et al., 2008; Tang et al., 2005; Wang et al., 2013) highlights the need for more comprehensive and long-term studies, to explore the impact of shelterwood harvesting, which is being increasingly adopted to manage forest ecosystems, on soil CO_2 efflux and the overall carbon budget.

Soil CO_2 efflux measured from a manual soil CO_2 efflux chamber system, conducted bi-weekly or monthly along a 100 m transect located in the vicinity of the chamber system found annual soil CO_2 efflux values were 617±33 g of C m^{-2} $year^{-1}$ at this study site (Khomik et al. 2010). Annual soil CO_2 effluxes reported by Khomik et al. (2010) were lower compared to annual soil CO_2 efflux measurements for this study; however measurements had lower temporal resolution, but higher spatial coverage. This study had higher temporal resolution because of half-hourly data measurements, but lower spatial coverage. Results from both studies highlight the challenges balancing temporal and spatial variability associated with measuring soil CO_2 efflux in forest ecosystems, to obtain meaningful representative results to capture dynamics of weather and the large spatial variability associated with species structure and distribution, and forest floor organic matter accumulation.

4.5 CONCLUSION

Climate variability and silvicultural practices, such as shelterwood harvesting, can influence forest carbon dynamics; however, their combined effects are complex and not well understood. By developing a Gaussian-Gamma specification model, this study showed that soil CO_2 efflux responds to soil temperature, soil moisture, climate variability and the shelterwood harvesting event. Soil CO_2 efflux was found to be the major contributor in the forested site, with total soil CO_2 efflux contributing on average 78±9% to ecosystem respiration. Heterotrophic respiration contributed roughly 68±14%, while autotrophic respiration contributed 32±14%, to the total annual soil CO_2 efflux. Shelterwood harvesting reduced heterotrophic respiration contribution from 80±6% to 53±21%, while increasing autotrophic respiration contribution from 20±5% to 47±21% with minimal change in contribution from the mineral-soil horizon respiration. By incorporating climate oscillations that prevail in the region, and their active years into the model, the relationship between soil CO_2 efflux and environmental factors were significantly improved. Results from this study will aid to better understand the influence of climate variability, affected by local and large-scale climate oscillations, and shelterwood harvesting to more accurately estimate soil CO_2 effluxes to develop and modify polices for better forest management focused on carbon sequestration and forest conversation.

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CHAPTER 5 **CONCLUSIONS**

Soil CO_2 efflux is an important component of the forest carbon cycle and shown to be influenced by climate variability (Allen et al., 2010; Bastos et al., 2013; Heimann and Reichstein, 2008; Hogg et al., 2017; Noormets et al., 2008; Piao et al., 2008; Wharton et al., 2012). Climate change models predict potential changes in the frequency, and/or severity of extreme weather events, such as droughts, flooding, heatwaves, duration and timing of the winter, and freeze/thaw periods, which can have an influence on carbon fluxes from forest ecosystems (Amiro et al., 2010; Davis et al., 1997; Granier et al., 2007; IPCC, 2013; Reichstein et al., 2013; Vose et al., 2012; Wu et al., 2017; Wullschleger and Hanson, 2006; Zscheischler et al., 2014).

Seasonal dynamics of temperature and precipitation, which are influenced by climate oscillations, are major environmental controls on soil CO_2 efflux (Borken et al., 2002; Monson et al., 2006; Ryu et al., 2009). Soil temperature was the main driving factor for soil CO_2 efflux in this study. Soil moisture also influenced soil CO_2 efflux at times, such as during drought periods. Snowpack depth also influences carbon cycling by introducing an insulating layer that decouples the soil matrix from the atmosphere, generating a microclimate in the near-surface soil layer that affects the activity of decomposing soil organisms, thus altering the rate of soil respiration (Monson et al., 2006). A change in climate, through increased warming or increased variability in the timing and amount of winter precipitation, can lead to changes in snow cover thickness and extent, directly affecting soil CO_2 efflux during the winter (Schindlbacher et al., 2014). Improving the quality (temporal and spatial) of soil CO_2 efflux data collected in the winter would aid to better understand the response of soil CO_2 efflux to environmental factors, such as snow and ice accumulation and melt. In this study, large pulses of soil CO_2 effluxes were measured in the winter of 2013/2014, exceeding those recorded during the growing seasons. Meteorological data indicated these soil CO_2 efflux pulses corresponded to an increase in soil temperature transitioning from below to above freezing, and an increase in soil moisture. These infrequent observations could also be associated to the melting of dense snow or ice on the surface of the soil, or ice in the shallow layers of the soil within the chambers. This snow or ice may have acted as a barrier preventing entrapped CO_2 from being released causing a pulse of soil CO_2 efflux.

Soil CO_2 efflux is comprised of autotrophic respiration (from roots) and heterotrophic respiration, derived from the mycorrhizal fungi, microorganisms in the rhizophere, decomposition of litter and soil organic matter in the organic and mineral-soil horizons (Lorenz, 2010). Removal of the litter-layer may modify soil water content at the near-surface soil layer, which may affect heterotrophic activity and disturb the root-soil interface (Hanson et al., 2000). It is crucial to study these components of soil respiration since responses to environmental factors differ (Hanson et al., 2000; Scott-Denton et al., 2006). This study showed that heterotrophic and autotrophic respiration contributed approximately 68±14%

and 32±14% to the total annual soil CO_2 efflux, respectively, while about two-thirds of the soil CO_2 efflux originated from the mineral-soil horizon.

Silvicultural practices, such as shelterwood harvesting, can also influence forest carbon dynamics by altering soil temperature, soil moisture, soil organic matter, root biomass and microbial activity (D'Amato et al., 2011; Gustafsson et al., 2012; Magruder et al., 2013; Peng et al., 2008; Powers et al., 2009; Vesala et al., 2005). These silvicultural practices are being adopted in North America and across the world, not only to manage forests, particularly planted stands, but also to enhance their carbon sequestration capabilities and/or to restore them to their previous native species (Gustafsson et al., 2012). Shelterwood harvesting encourages natural regeneration and development of seedlings in partial shade (OMNRF, 1998; OMNRF, 2000; Trant, 2014). Shelterwood harvesting changes the tree spacing, number and size distribution, which can affect the dynamics of carbon uptake through above-ground productivity and carbon losses through soil respiration (Marchi et al., 2018; Vesala et al., 2005). Greater open-canopy alters the radiation dynamics, interception of precipitation, wind flow patterns, water vapour deficit, soil temperature and soil moisture (Stoffel et al., 2010; Vesala et al., 2005; Woods et al., 2006). Shelterwood harvesting or elective thinning also causes changes in root density and production impacting autotrophic respiration, litterfall input and hence soil microbial activity.

Limited information exists on temporal and spatial responses of soil CO_2 efflux response to shelterwood harvesting, particularly in conifer stands (Peng et al., 2008). Seasonal variability of soil temperature, soil moisture and soil CO_2 efflux were examined and pre- (2008 to 2011) and post-harvesting (2012 to 2014) periods were compared. This study found that soil temperature and soil moisture did not show any significant changes post-harvesting, when compared to pre-harvesting measurements. Despite similar soil microclimates, total soil CO_2 effluxes were significantly reduced by up to 37%. Soil CO_2 effluxes from heterotrophic sources were significantly reduced by approximately 27%, while no significant difference in the mineral-soil horizon sources were measured post-harvesting. Additional collection of soil CO_2 efflux measurements, both temporally and spatially, are needed to determine the full effect of adopting the shelterwood system on soil respiration, and particularly, the impact on both heterotrophic and autotrophic respiration in the years that follow post-harvest as the forest re-establishes.

Low frequency climate oscillations are important atmospheric phenomenon which influences seasonal and annual climate variations (Bonsal et al., 2001; Shabbar et al., 1997; Shabbar and Bonsal, 2004; Zhang et al., 2011). Thorne and Arain (2015) (Chapter 2) showed climate in the Great Lakes region is strongly influenced by low frequency climate oscillations and influences have shown to affect ecosystem respiration at the forest site during the winter and spring seasons. Climate oscillations are important drivers of extreme weather events, affecting their intensity, frequency, and spatial patterns, and can therefore cause large changes in carbon exchanges in forest ecosystems in northeastern parts of North America.

Gaussian-Gamma specification models were developed to explain the temporal variability in soil CO_2 efflux with soil temperature, soil moisture, the shelterwood harvesting and climate oscillations. Comparison of output and estimated parameter coefficients among the three models (GG, GGTHIN and GGTOS) found that the addition of the shelterwood harvesting event and climate oscillation components increased the predictive ability for each soil CO_2 efflux component. Heterotrophic respiration contributed approximately 68±14%, while autotrophic respiration contributed 32±14%, to the total annual soil CO_2 efflux. Shelterwood harvesting reduced the contribution from heterotrophic respiration from 80±6% to 53±21%, while contribution from autotrophic respiration increased from 20±5% to 47±21% with minimal change in the contribution from the mineral-soil horizon. By incorporating the climate oscillations that prevail in the region, and their active years into the model, the relationship between soil CO_2 efflux and environmental factors were significantly improved.

Given that soil CO_2 efflux at this site is the largest component of R_E, and there was a reduction in soil CO_2 efflux during the post-harvesting period, while both GEP and R_E increased, it appears that the increase in R_E was primarily attributed to the increase in autotrophic respiration due to enhancement both the understory and/or growth in remaining trees after the harvesting. An increase in

R_E may also be attributed to a disparity between flux footprints for the eddy-covariance fluxes, which encompass the entire forest stand, as compared to the localized area where the soil CO_2 efflux chambers were installed. Soil CO_2 efflux chambers often overestimate soil respiration as compared to R_E measured by the eddy-covariance system (Braendholt et al., 2017). This study highlights the complexities and challenges associated with assessing soil CO_2 effluxes and the need for more comprehensive and long-term field studies, to better understand the response of these fluxes to possible changes in the future climate. Results presented are useful for forest management practitioners, specifically those focused on carbon sequestration and forest conservation. The knowledge gained from this research will help in developing polices for better managements of forest ecosystems in eastern Canada, in addition to assisting in planning and developing realistic strategies to offset fossil fuel CO_2 emissions.

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