Assessing Regional Scale Fluxes of Mass, Momentum, and Energy with Small

Environmental Research Aircraft

By

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B.S. (San Diego State University) 1995

DISSERTATION
Submitted in partial satisfaction of the requirements for the degree of
DOCTOR OF PHILOSOPHY
in the
Joint Program in Ecology in the
OFFICE OF GRADUATE STUDIES
of the
UNIVERSITY OF CALIFORNIA
DAVIS
and the
DIVISION OF GRADUATE AFFAIRS
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2013
To my parents, Rogelio and Adele Zulueta
Acknowledgements

This dissertation would not have been made possible without the continuous support, encouragement, and patience from my advisor, Dr. Walter C. Oechel. The opportunity he has given me has allowed me to literally "take flight" and view the world, and science from a very different perspective. For this I am eternally grateful. I would like to thank my committee members, Dr. Kyaw Tha Paw U and Dr. Douglas Deutschman who were always supportive of my research regardless of the many directions it has taken throughout the years. Their technical and editorial comments have no doubt elevated the quality of my manuscripts. Special thanks to Dr. Kyaw Tha Paw U for taking the time and having the patience with me as I learned and worked through the critical micrometeorological theories and practices that I required for my research.

I would like to thank Dr. Steven Hastings, who many years ago took a chance and allowed me the opportunity to do some field research in Alaska. It was his energy and enthusiasm for the field work and the science that first hooked me and started me down this path. He always had the time to help and his uncanny ability to work endlessly was always an inspiration. I would also like to thank Pablo Bryant for always being there and supporting me throughout this whole ordeal. There was never anything that I asked for that you could not build, or any challenge I presented that you could not solve. You truly are my best man. I want to especially thank Joeseph Verfaillie Jr. for all the time and effort spent with me and the Sky Arrow. The commitment you put into getting the aircraft flux system functioning through every conceivable environmental condition and/or logistical constraint was truly amazing. Even through the most frustrating technical issues we encountered, I was always confident in your
ability to resolve the problems so I could continue flying and collecting data. If it wasn't for your calm and positive attitude through freezing arctic conditions, sweltering heat and humidity, and even tornados, I would not have been able to do this. A special thanks to Dr. George Vourlitis for introducing me to the eddy covariance technique and allowing me to learn first-hand its implementation in the field, and Dr. Beniamino Gioli for guidance and help with the most difficult challenges I encountered with the aircraft data and processing.

I truly appreciate the help from Dr. William T. Lawrence for his work with vastly complicated satellite data from which he was able to extract meaningful results for me, Dr. Henry Loescher for guidance and inspiration though the latter stages of writing my manuscripts, Jeffery Iles for his GIS wizardry, Ed Dumas, Dr. Steve Brooks, Dr. Jorg Hacker, and the late Dr. Timothy Crawford for all their help with the Sky Arrow and other aircraft-related intricacies.

My deepest thanks to all my fellow graduate students and field assistants who endured endless days and nights in the field with me, particularly, Masayoshi Mano, Glen Kinoshita, Hyojung Kwon, Hiroki Ikawa, Hongyan Luo, and Cheryl Laskowski. Field assistance from Rena Bryan and Cove Sturtevant are greatly appreciated. You all are what made the field work a great experience.

I would like to extend my appreciation to Greg Karnes for teaching me how to fly and instilling in me the confidence to overcome any in-flight situation, Christina Halterman for showing me what the Sky Arrow is truly capable of, and particularly to my A&P mechanics, Chuck Anclein, and the late Brad King, for keeping the Sky Arrow airworthy and flying safely. I completely trusted you with my life. To the engineers at Iniziative Industriali Italiane (3I), you have created an incredible aircraft, you all allowed me to soar.
None of the Alaska field work would have been possible without logistics help from Polar Field Services, the Barrow Arctic Science Consortium, and the people of Barrow, Alaska. I would like to especially acknowledge the extraordinary support and efforts from Marin Kuizenga who was able to provide logistical support regardless how demanding or financially infeasible the requests were. My research work in Baja California, Mexico, would not have been possible without support from the wonderful staff at the Centro de Investigaciones Biológicas del Noroeste, S.C (CIBNOR).

My earnest thanks to Lorraine Ahlquist and Nalu Kai for the wonderful companionship throughout the years, your support and patience allowed me to explore all avenues of my research work, and the inspiration to continually move forward.

Lastly I would like to extend my heartfelt gratitude to my family. To my parents, Roger and Adele Zulueta, you were always supportive of every endeavor I chose to explore. To my brother, Ogie, and my siters, Myla and Melanie, you have always been my center and my ground. To my extended family, Alyssa Lupo, Chris Aguillon, and Danny Ledesma, thank you for allowing me into your lives.

The initial development of the Sky Arrow 650TCN Environmental Research Aircraft, N272SA, was funded by a grant from the National Science Foundation under award DBI-9604793, with generous support from the San Diego State University Research Foundation. Subsequent research works were funded by grants from the National Science Foundation under awards, OPP-0436177, OISE-0072140, DGE-0139378, and the Department of Energy Western Regional Center of the National Institute for Climatic Change Research under award DE-FC02-06ER64159.
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Abstract

Natural ecosystems are rarely structurally or functionally homogeneous. This is true for the complex coastal regions of Magdalena Bay, Baja California Sur, Mexico, and the Barrow Peninsula on the Arctic Coastal Plain of Alaska. The coastal region of Magdalena Bay is comprised of the Pacific coastal ocean, eutrophic lagoon, mangroves, and desert ecosystems all adjacent and within a few kilometers, while the Barrow Peninsula is a mosaic of small ponds, thaw lakes, different aged vegetated thaw-lake basins (VDTLBs) and interstitial tundra which have been dynamically formed by both short- and long-term processes. We used a combination of tower- and small environmental research aircraft (SERA)-based eddy covariance measurements to characterize the spatial and temporal patterns of CO₂, latent, and sensible heat fluxes along with MODIS NDVI, and land surface information, to scale the SERA-based CO₂ fluxes up to the regional scale.

In the first part of this research, the spatial variability in ecosystem fluxes from the Pacific coastal ocean, eutrophic lagoon, mangroves, and desert areas of northern Magdalena Bay were studied. SERA-derived average midday CO₂ fluxes from the desert showed a slight uptake of -1.32 µmol CO₂ m⁻² s⁻¹, the coastal ocean also showed uptake of -3.48 µmol CO₂ m⁻² s⁻¹, and the lagoon mangroves showed the highest uptake of -8.11 µmol CO₂ m⁻² s⁻¹. Additional simultaneous measurements of NDVI allowed simple linear modeling of CO₂ flux as a function of NDVI for the mangroves of the Magdalena Bay region.

In the second part of this research, the spatial variability of ecosystem fluxes across the 1802 km² Barrow Peninsula region was studied. During typical 2006 summer conditions, the midday hourly CO₂ flux over the region was -2.04 x 10⁵ kgCO₂ hr⁻¹. The CO₂ fluxes among the
interstitial tundra, Ancient and Old VDTLBs, as well as between the Medium and Young VDTLBs were not significantly different. Combined, the interstitial tundra and Old and Ancient VDTLBs, represent ~67% of the Barrow Peninsula surface area, accounting for ~59% of the regional flux signal. Though the Medium and Young VDTLBs represent ~11% of the surface area, they account for a large portion, ~35%, of the total regional flux. The remaining ~22% of the surface area are lakes and contributed the remaining ~6% of the total regional flux. Previous studies treated vegetated areas of the region as a single surface type with measurements from a few study sites; doing so could underestimate the regional flux by ~22%.

The San Diego State University Sky Arrow 650TCN Environmental Research Aircraft proved to be an effective tool in characterizing land-atmosphere fluxes of energy, CO$_2$ and water across heterogeneous landscapes at the scale of 1 km, and was capable of discriminating fluxes from the various ecosystem and land surface types a few kilometers distant. Here, we demonstrate that SERA-based approaches have the ability to cover large spatial scales while measuring the turbulent fluxes across a number of surfaces and combined with ground- and satellite-based measurements provide a valuable tool for both scaling and validation of regional-scale fluxes.
Introduction

A key challenge to understanding regional carbon exchange between the biosphere and the atmosphere is one of adequate sampling, both temporally and spatially, to infer processes and annual budgets. This is exemplified by the difficulty in determining representative plot and tower measurements that can be extrapolated to larger landscapes and regions due to the spatial heterogeneity in the surface fluxes (Desai et al., 2008). Multiple factors cause spatial variability in the surface fluxes including; variation in edaphic factors, topography, hydrology, vegetation composition and function, land use, disturbance history, and so on. These variations are often not measured or monitored and their importance is not fully understood or explored. Assessing spatial variability in surface fluxes in complex and heterogeneous ecosystems is important to understanding the role they play in local and regional carbon balance.

While ecosystem models and satellite remote sensing are used to estimate fluxes at large spatial scales, the reliability of these models and satellite-derived products for upscaling from point measurements to regional scales requires assumptions that are difficult to validate, and empirical data to verify these data products and establish mechanistic linkages among spatial scales are often lacking (Desai et al., 2008). Measurements from ecosystem flux towers typically lack sufficient spatial density to adequately represent the region’s heterogeneity in structure and process, making the assimilation of plot level and tower measurements with regional and global estimates challenging (Desai et al., 2010). A key to understanding the regional patterns of and controls on mass and energy fluxes in terrestrial and aquatic ecosystems is the development of new methodologies that link and constrain multiple scales (Dolman et al., 2009).
One such methodology that can provide linkages among spatial scales, and the ability to measure regional-scale fluxes of mass and energy directly, has been the development and use of Small Environmental Research Aircraft (SERA; see e.g., Crawford et al., 1990; Crawford et al., 1996; Bange & Roth 1999; Crawford et al., 2001; Dumas et al., 2001; Gioli et al., 2004; Bange et al., 2006; Gioli et al., 2006; Hall et al., 2006). These SERAs have the ability to measure spatially integrated surface properties across large spatial scales and combined with tower-based flux measurements provide both a spatial and temporal integration of flux measurements (Brooks et al., 1996; Crawford et al., 1996; Oechel et al., 1998; Gioli et al., 2004; Mauder et al., 2007; Migletta et al., 2007). The complementary nature of tower- and aircraft-based flux measurements provides a mechanism for scaling both in time and in space, and since SERAs also have the ability for low level remote sensing of surface properties, combining satellite-derived data products, like the normalized difference vegetation index (NDVI; Tucker 1979) from the Moderate Resolution Imaging Spectroradiometer (MODIS) on the Terra satellite, provides the linkage to scale flux measurements from the plot-scale, to the landscape, and up to the region.

The San Diego State University (SDSU) Sky Arrow 650TCN Environmental Research Aircraft (registration number N272SA, serial number cn002; hereafter referred to as Sky Arrow) is a SERA that was specifically developed to incorporate instrumentation similar to those on tower-based eddy covariance (EC) systems, as well as instrumentation capable of low level remote sensing similar to those found on Earthward looking satellites. A full description of the SDSU Sky Arrow SERA platform can be found in Chapter 1 and in Dumas et al., (2001) with additional details described in Chapter 2. Two “sister” aircraft [one used in the Regional Assessment and Modelling of the Carbon Balance of Europe (RECAB) project and one operated by the University of Alabama], which have the same airframe but with slightly modified and
updated instrumentation packages, are also described in Gioli et al. (2004; 2006) and Hall et al., (Hall et al., 2006), respectively. The use of SERA is therefore a “bridge” from the direct surface flux measurements of tower-based eddy covariance (EC; Swinbank 1951; Baldocchi et al., 1988; Loescher et al., 2006) and satellite-based remote sensing.

The research presented here used the SDSU Sky Arrow SERA to measure regional-scale fluxes across two different regions with vastly different ecosystem types and spatial heterogeneities. The principal goal was to establish a methodology focused on the SERA-based EC measurements that could encompass the regional spatial heterogeneity of surface fluxes and be used to scale flux measurements to the region. The methods and procedures presented could be used to provide accurate and improved regional flux estimates to instruct, verify, and validate large-scale estimates from ecosystem models as well as other satellite-derived data products, and can also be instrumental in constraining the uncertainty in regional-scale fluxes.

The first paper (Zulueta et al., 2013) introduces and describes the SDSU Sky Arrow SERA platform and provides a detailed overview of aircraft-based flux measurements and procedures, which are used to examine the spatial variability in fluxes of CO₂, water vapor, and energy across the structurally and functionally complex coastal region of northern Magdalena Bay, Baja California Sur, Mexico. The Magdalena Bay region provides a case study for comparisons of ecosystem functioning of vastly different ecosystems with the same synoptic-scale climate patterns, and contrasting micro-climate, over relatively short distances. Direct measurements and regional estimates of mangrove forest net productivity are limited in this area, so an overall goal of this research was to estimate the midday net ecosystem CO₂ flux of the entire mangrove forest area of the Magdalena Bay region using a scaling methodology that combines tower- and aircraft-based EC measurements, and satellite remote sensing. Specific
questions addressed are (i) do aircraft-based flux measurements have sufficient sampling resolution to detect differences in process rates among adjacent coastal ecosystems (desert, mangrove, lagoon, and ocean) only a few kilometers apart (ii) and how do the magnitude and direction of the fluxes differ among the several ecosystems of the area?

The second paper (Zulueta et al., 2011) continued the SERA measurement methodology established in the first paper but across the arctic tundra ecosystems along the Barrow Peninsula of Alaska. The tundra landscape of the Barrow Peninsula is a mixed chronosequence of small ponds, thaw lakes, vegetated drained thaw-lake basins (VDTLBs), and interstitial tundra in various states of succession. This mosaic of different aged land surfaces is important to the region’s carbon balance though the focus of long-term ecosystem flux studies in the region has been on the upland, drier, interstitial tundra areas (e.g., Coyne & Kelley 1975; Oechel et al., 2000; Kwon et al., 2006), with limited studies of carbon fluxes over Alaskan arctic ponds and lakes (Coyne & Kelley 1974; Kling et al., 1991, 1992; Eugster et al., 2003), few on thaw lakes and VDTLBs (see e.g., Phelps et al., 1998; Harazono et al., 2003; Zona et al., 2010), and even less on measurements that integrate the large scale variability of the larger arctic landscape or region (see e.g., Brooks et al., 1996; Oechel et al., 1998). The Barrow Peninsula provides an opportunity to compare ecosystem fluxes across this heterogeneous landscape, and coupled with MODIS NDVI and VDTLB spatial information (Hinkel et al., 2003), we scale the fluxes of CO₂ to the Barrow Peninsula region. The objectives of this research are to i) determine the spatial pattern of CO₂ fluxes across the arctic landscape incorporating different aged VDTLBs and interstitial tundra, ii) determine the temporal pattern of CO₂ and energy fluxes of two different aged VDTLBs and interstitial tundra, and iii) assess the regional scale flux using a combination of tower-, aircraft-, and satellite-based measurements. We also address the question: are aircraft-
based flux measurements suitable to accurately characterize the spatial fluxes along the heterogeneous landscape of the Arctic Coastal Plain?
References


Chapter 1

Aircraft regional-scale flux measurements over complex landscapes of mangroves, desert, and marine ecosystems of Magdalena Bay, Mexico

In Press in the Journal of Atmospheric and Oceanic Technology

Citation:
Aircraft regional-scale flux measurements over complex landscapes of mangroves, desert, and marine ecosystems of Magdalena Bay, Mexico

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Keywords:
aircraft flux, eddy covariance, Magdalena Bay, mangrove, Mexico, NDVI
Abstract

Natural ecosystems are rarely structurally simple or functionally homogeneous. This is true for the complex coastal region of Magdalena Bay, Baja California Sur, Mexico where we studied the spatial variability in ecosystem fluxes from the Pacific coastal ocean, eutrophic lagoon, mangroves, and desert. The Sky Arrow 650TCN Environmental Research Aircraft proved to be an effective tool in characterizing land-atmosphere fluxes of energy, CO$_2$ and water across a heterogeneous landscape at the scale of 1 km. The aircraft was capable of discriminating fluxes from all ecosystem types, as well as between near shore and coastal areas a few kilometers distant. Aircraft-derived average midday CO$_2$ fluxes from the desert showed a slight uptake of -1.32 µmol CO$_2$ m$^{-2}$ s$^{-1}$, the coastal ocean also showed uptake of -3.48 µmol CO$_2$ m$^{-2}$ s$^{-1}$, and the lagoon mangroves showed the highest uptake of -8.11 µmol CO$_2$ m$^{-2}$ s$^{-1}$.

Additional simultaneous measurements of the normalized difference vegetation index (NDVI) allowed simple linear modeling of CO$_2$ flux as a function of NDVI for the mangroves of the Magdalena Bay region. Aircraft approaches can, therefore, be instrumental in determination of regional CO$_2$ fluxes and can be pivotal in calculating and verifying ecosystem carbon sequestration regionally when coupled with satellite derived products and ecosystem models.
1. Introduction

A ubiquitous problem in Carbon Cycle Science is one of adequate sampling, both in time and in space, to infer processes and annual budgets. This problem is typified by the difficulty in determining representative plot and tower measurements that can be extrapolated to larger landscapes and regions (Desai et al., 2008). Understanding processes and controls on carbon flux to make predictions due to climate change are often done at the ecosystem level (micrometeorological studies), while regional-to-global feedbacks, under both current and future conditions, are often done at larger regional scales embodying synoptic-scale weather patterns (Turner et al., 2004).

Models and satellite-based remote sensing observations are often used to estimate fluxes of mass and energy at larger spatial scales. However, empirical data to verify these models and satellite derived products, and that establish mechanistic linkages among scales of measurement are often lacking (Desai et al., 2008). Ecosystem flux estimates scaled from a single tower-based measurement to the landscape- or regional-scale typically lack sufficient spatial density to adequately represent the region’s heterogeneity in structure and process (see e.g., Zulueta et al., 2011). Even ecosystems considered homogeneous (e.g., grasslands and tundra) can have a large degree of functional variability over short distances (meters to kilometers), making the assimilation of plot level and tower measurements with regional and global estimates challenging (Desai et al., 2010). Key to understanding the regional patterns and controls on mass and energy fluxes in terrestrial and aquatic ecosystems is the development of new methodologies that link and constrain the multiple scales of interest (Dolman et al., 2009).

Typically, when assessing carbon flux in terrestrial ecosystems, predominant ecosystem types are selected and monitored because they are assumed to be representative of the regional
flux. However, multiple factors cause spatial variability in the surface fluxes. These include variations in edaphic factors, topography, hydrology, vegetation composition and function, land use, disturbance history, etc. These variations are often not measured or monitored and their relative importance is often not fully understood or explored. This is particularly true in areas where ecosystems are remote, such as the Arctic and tropics, or logistically difficult to adequately measure, such as complex coastal ecosystems.

Assessing spatial variability in surface fluxes in complex ecosystems is important to understand the role they play in local and regional carbon balance; specifically in systems that provide important ecosystem services and have high economic value, e.g., the subtropical mangrove forests along the Baja California, Mexico coastline. One such system is the mangrove forests along Magdalena Bay, the largest expanse of mangrove forests along the Pacific coast of Baja California, Mexico. Magdalena Bay, the surrounding mangrove areas and coastal estuaries, sustain a variety of resident and migratory waterfowl (see Zarate-Ovando et al., 2006), are a feeding and developmental area for five of the world’s seven sea turtle species (see Cliffton et al., 1995) are one of the major winter breeding lagoons for the North Pacific gray whale, *Eschrichtius robustus* (see Urban et al., 2003), and are also a key regional fishing and recreational area. Magdalena Bay has an extensive lagoon area with mangrove forests surrounded by barrier islands and the Pacific Ocean to the west, a central lagoon, and desert eastward of the lagoon to the landward side.

Mangrove forests and subtropical coastal ecosystems are disappearing at a rapid rate. Over the past two decades, over 35% of the world’s mangrove forests have been lost and converted to other land use (Valiela et al., 2001). Despite the mangroves’ high economic value, ecosystem services (Costanza et al., 1997; Reid et al., 2006; Nagelkerken et al., 2008), and role
in global carbon cycling (Bouillon et al., 2008), they continue to be severely disrupted by
tourism developments, exploitation, and aquaculture (Valiela et al., 2001; Alongi 2002; Duke et al., 2007). In May 2004, the Mexican government repealed a law protecting mangroves, which opened the way for extensive coastal developments exacerbating the severe pressure from land use, development, and shrimp farming (Paez-Osuna et al., 2003; Glenn et al., 2006). Though a new mangrove protection law was passed in February 2007, growth and interest in nautical and eco-tourism in Mexico continue to rise. Plans are currently underway to expand and modernize thirteen existing marinas, construction of eleven new commercial ports and coastal resorts along the coast of Baja California and the Sea of Cortez (e.g., the Escalera Nautica project). Continued exploitation and development along these coastlines could threaten previously undeveloped, unprotected mangrove forests and coastal estuaries including those within Magdalena Bay.

The Magdalena Bay area provides a case study for these mangrove forests and also allows comparisons of ecosystem functioning of vastly different ecosystems with the same synoptic-scale climate patterns, and contrasting micro-climate, over relatively short distances. In this study, we use an aircraft-based eddy covariance system to examine the spatial variability in fluxes across various landscapes for the complex Pacific coastal ecosystem in northern Magdalena Bay, Baja California Sur (BCS), Mexico. Since direct measurements and regional estimates of mangrove forest net productivity are limited in this area, an overall goal of this research was to estimate the midday net ecosystem CO$_2$ flux of the mangrove forest area of the Magdalena Bay region with aircraft and satellite measurements. The questions we address are:

(i) Do aircraft-based flux measurements have sufficient sampling resolution to detect differences in process rates among adjacent coastal ecosystems (desert, mangrove, lagoon, and ocean) only a
few kilometers apart? (ii) How do the magnitude and direction of the fluxes differ among the several ecosystems of the area?
2. Materials and Methods

a. Site Description

These data presented here were collected between 24 July and 28 July 2004, in the vicinity of Puerto Aldofo Lopez Mateos (N 25.192450°, W 112.115783°) (hereafter referred to as Lopez Mateos; Fig. 1). This is a small fishing town located in BCS, Mexico, along the northern extent of the Magdalena Bay lagoon. This northern area is protected from the Pacific Ocean by the Isla Magdalena, a sandy barrier island which parallels the mainland coastline, and has three distinct canals to the ocean. The closest canal is the Boca la Soledad, and is 9 km to the northwest of the town.

The area is in a subdivision of the Sonoran Desert classified as the Magdalena Region (Shreve & Wiggins 1964) and the regional terrestrial vegetation is sparse and generally described as desert scrub (Wiggins 1980; Turner & Brown 1982). The precipitation regime in the Magdalena Region is seasonally variable with an annual precipitation of 125 mm with the majority of the precipitation occurring in the fall and winter (89 mm) with the summer and spring having the lowest amounts, 31 mm and 5 mm respectively (Hastings & Turner 1965).

The desert surrounding Lopez Mateos is generally sand with scattered rocky outcroppings. The larger dominant plant species are drought and salt tolerant and include cacti such as *Stenocereus thurberi, Opuntia cholla*, and numerous large individuals (>5 m tall) of the columnar cactus, *Pachycereus pringlei*, and some trees such as *Bursera microphylla, Bursera hindsiana, Simondsia chinensis*, and *Larrea tridentata*. *Ambrosia magdalanae* is a common herbaceous plant. The coastal margins adjacent the desert and mangrove lagoon include perennial species of *Isocoma menziesii, Monanthochloe littoralis*, and *Suaeda monquinii* while the salt marshes along the lagoon edges are occupied by halophytes including *Allenrolfea*
occidentalis, Batis maritime, Salicornia bigelovii, and Sesuvium portulacastrum. The sand
dunes have sparse coverage of Abronia maritima and Lotus bryantii. Although it is part of the
Sonoran Desert, the Magdalena Region contains endemic and expansive mangrove lagoons in
protected coves of Magdalena Bay on the western coast.

Magdalena Bay is a subtropical lagoon system with mangrove forests lining the inside
interior coastline of the protected bay and natural shallow channels (Alvarez-Borrego et al.,
1975). This eutrophic coastal lagoon has higher salinity than the open ocean due to low annual
precipitation, high rate of evaporation, and minimal freshwater inputs from the land (Alvarez-
Borrego et al., 1975). The dominant mangrove species are the red (Rhizophora mangle), white
(Laguncularia racemosa), and to a lesser extent black (Avicennia germinans) mangrove trees.
The Magdalena Bay region spans between 25.73° N and 24.27° N and 111.32° W and 112.35° W
and covers an area of approximately 1409 km² of which the mangrove forest is estimated to
cover 361.23 km² (determined in this study). The average tree height and stem diameter were
3.15 m and 4.09 cm, respectively, with tree densities within the mangroves estimated at 2569
trees ha⁻¹, with a basal area of 3.21 m² ha⁻¹ (Chavez 2006).

Detailed descriptions of Magdalena Bay and surrounding environments can be found in
Alvarez-Borrego et al. (1975) with additional information on the physical and biological
characteristics of the bay region summarized in Bizzarro (2008).

b. Meteorological conditions during flux flights (0930-1830 MST)

All flux flights were conducted under Visual Flight Rules (VFR) conditions from 0930 to
1830 Mountain Standard Time (MST). Clear skies prevailed throughout the campaign with an
average photosynthetically active radiation (PAR) of 1365 µmol m⁻² s⁻¹ and reaching a maximum
of 2224 µmol m\(^{-2}\) s\(^{-1}\) at about 1300 MST. The average air temperatures varied across the different ecosystems with the average across all the flight transects at 24.3 °C with the highest average temperatures observed over the desert (26.8 °C) and the lowest over the ocean (22.6 °C) with the mangroves in between (24.3 °C). The highest recorded temperature was during the midday over the desert and was 31.7 °C.

The wind direction was typically onshore, west-east flow during the day with stable atmospheric conditions occurring at night and early morning. The average wind direction was 282 ± 18.6° with an average speed of 5.7 m s\(^{-1}\) across all transects. The horizontal wind speeds varied across the different ecosystems with highest average speeds over the ocean (6.7 m s\(^{-1}\)) relating to having the lowest friction velocity \((u_*)\) and surface roughness \((z_0)\), 0.26 m s\(^{-1}\) and 0.15 m, respectively. The \(u_*\) and \(z_0\) were similar between the desert and mangrove ecosystems, though the desert had the highest values. Additional turbulence parameters during the flux flights are shown in Table 1.

c. SDSU Sky Arrow 650TCN Environmental Research Aircraft

The San Diego State University (SDSU) Sky Arrow 650TCN Environmental Research Aircraft (ERA; Fig. 2) (hereafter referred to as Sky Arrow) was used to measure fluxes of CO\(_2\), latent \((\lambda E)\) and sensible \((H)\) heat, and momentum. This custom designed aircraft platform is ideal for atmospheric turbulence measurements due to the narrow stream wise profile, high wing design, slow flight speed, and aft mounted engine enabling it to fly in the upper surface layer and lower convective boundary layer. The SDSU Sky Arrow (registration number N272SA, serial number cn002) was the first of its kind and received type certification by the US Federal Aviation Administration (FAA) in July, 1999. It is an all-composite aircraft with custom
designed mounting ports in the floor of the fuselage, and mounting hard-points for instrumentation in the nose and horizontal stabilizer (see Fig. 2). Further details and specifications of the SDSU Sky Arrow are listed in Table 2.

The slim airframe and aircraft configuration allows for an unobstructed nose and the placement of the turbulence sensors to be within the mean streamline with minimal distortion from the airframe (Wyngaard 1988), propeller (Kalogiros & Wang 2002a), and up- and side-wash generated by the wings (Crawford et al., 1996a; Kalogiros & Wang 2002a; Garman et al., 2008). This aircraft incorporated instrumentation for eddy covariance measurements and low-level remote sensing (Dumas et al., 2001). We used the National Oceanic and Atmospheric Administration (NOAA) Air Resources Laboratory developed Mobile Flux Platform (MFP) for eddy covariance measurements (Crawford et al., 1990; Hall et al., 2006) which incorporates a nose mounted Best Air Turbulence (BAT) probe and pressure sphere (Crawford & Dobosy 1992; Hacker & Crawford 1999). A fast response open-path infrared gas analyzer (IRGA) (LI-7500, LI-COR Inc., Lincoln, NE, USA) also located on the nose provided simultaneous measurements of CO₂ and water vapor where density corrections due to the mass transfer of heat and water vapor from one averaging period to the next (Webb et al., 1980) were applied. Dew point temperature was measured with a dew point hygrometer (DewTrak 200, EdgeTech, Marlborough, MA, USA) from the underside of aircraft.

Aircraft attitude was measured using nose and airframe mounted accelerometers (ICS3022, Measurement Specialties, Hampton, CA, USA) and a vector attitude Global Positioning System (GPS); Trimble Advanced Navigation System (TANS Vector, Trimble Navigation Ltd., Sunnyvale, CA, USA) at 50 and 10 Hz respectively. Blending of the two signals achieved an attitude sampling frequency of 50 Hz (Eckman et al., 1999) with a ±0.05
degree accuracy. Aircraft position was measured at 10 Hz with a 12-channel L1 frequency GPS (Model 3151, NovAtel Inc., Calgary, Alberta, Canada) and differentially corrected in post-processing (Waypoint GrafNav, NovAtel Inc.) with position data from a stationary GPS base station of the same type as on the aircraft.

Simultaneous measurements of incoming and reflected radiation were also measured. Up- and down-welling photosynthetically active radiation (PAR) were made with two silicon quantum sensors (LI-190SB, LI-COR, Inc.) and net radiation was measured with a Fritschen-type net radiometer (Q*7.1, REBS Inc., Seattle, WA, USA). The net radiometer was not actively aspirated, because once in flight the airflow around the sensor met or exceeded its requirements for aspiration. The PAR and net radiation sensors were located on the port side of the aircraft’s horizontal stabilizer (see Fig. 2).

Low-level, remotely-sensed surface temperature was measured with an infrared temperature sensor (4000.4GH, Everest Interscience, Tucson, AZ, USA), while a laser altimeter was used to measure aircraft altitude above ground level (LD90-3300HR, Riegl, Horn, Austria). Hyperspectral reflectance measurements from 304 to 1134 nm (255 bands, 3.2 nm bins) were made using a dual channel spectrometer (UniSpec-DC, PPSystems, Amesbury, MA, USA) with a upward facing cosine incident light receptor, and a 10° field of view downward looking lens resulting in an average sampling footprint diameter of 1.48 m at the 8.48 m average flight height above the ground. Detector integration times varied with light intensity and ranged from 9 to 20 ms resulting in sampling frequencies from 5-10 Hz. Ten integration periods were internally averaged before file storage to the data acquisition system. The average ground speed of the aircraft was 38.2 m s\(^{-1}\) so each stored measurement was therefore integrated over an average length of 5.06 m resulting in effective “pixel” resolution of 1.48 m x 5.06 m. A fluoropolymer-
based solid thermoplastic calibration panel (Spectralon SRT-99, Labsphere, Sutton, NH, USA) was measured before and after each flight to correct for variations in solar radiation. The spectral reflectance were interpolated into one nm bands and the normalized difference vegetation index (NDVI) (Tucker 1979) was calculated using the red (620-670 nm) and near infrared (841-876 nm) wavelengths which correspond to the Terra satellite Moderate Resolution Imaging Spectroradiometer (MODIS) bands 1 ($b_1$) and 2 ($b_2$) respectively and calculated as:

$$NDVI = \frac{(b_2 - b_1)}{(b_2 + b_1)}.$$ (1)

Atmospheric turbulence and wind velocity relative to the aircraft were measured with the nose mounted BAT probe housing a 9-hole pressure sphere used to measure static pressure, and to convert micro-scale pressure fields into known velocities of the three dimensional ($u$, $v$, and $w$) winds and their high frequency fluctuations (Brown et al., 1983; Crawford & Dobosy 1992; Hacker & Crawford 1999; Garman et al., 2006). The fast response temperature fluctuations were measured within the nose hemisphere with a 0.13 mm diameter microbead thermistor (VECO 32A402A, YSI, Yellow Springs, OH, USA) at the hemisphere’s stagnation point ($Tp_1$) and a secondary microbead thermistor located within a “fast flow” port at the sphere edge ($Tp_2$), while the mean air temperature was measured with a Thermilinear network (44212, YSI Inc.) (Crawford & Dobosy 1992).

Calibration of the aircraft’s wind vector system was conducted on 28 July 2004 in the morning between 0730-0900 MST, over the ocean, at elevations from 1250-1575 m above sea level (ASL). Conditions during this time were ideal for a calibration flight as the boundary layer was relatively low to the ground and the mixing layer was not deep (i.e., approximately 340 m ASL) with winds aloft being smooth and consistent. In-flight calibrations of the wind vector system were necessary for the instrument installation positions, aircraft flight performance, and
aerodynamics of the aircraft. These calibrations are used to estimate flow distortion, calculate corrections for wind measurements, and ensure synchronization of the data acquisition and measurement instrumentation (Lenschow 1986; Scott et al., 1990; Bögel & Baumann 1991; Vellinga et al., 2013). Modified calibration maneuvers included a constant altitude wind box, standard rate turns (3° s⁻¹), pilot-induced pitch and yaw oscillations, sideslip, and acceleration/deceleration maneuvers (Telford et al., 1977; Lenschow 1986; Bögel & Baumann 1991; Tjernström & Friere 1991; Leise & Masters 1993; Khelif et al., 1999; Williams & Marcotte 2000; Kalogiros & Wang 2002b; Lenschow et al., 2007; Vellinga et al., 2013). A detailed description of calibration procedures for a similarly instrumented and configured Sky Arrow ERA is presented in Vellinga et al. (2013).

A factory calibration of the IRGA was used at the beginning of the measurement campaign, a field calibration was done on 27 July 2004, and one at the end of the campaign and followed AmeriFlux protocols (http://public.ornl.gov/ameriflux/sop.shtml). A CO₂ free ultra-zero air was used as the zero set point for both the CO₂ and water vapor, NIST-traceable primary gas standards were used to span the CO₂, and a portable dew point generator (LI-610, LI-COR, Inc.) was used to span the water vapor. Following the field calibration, the IRGA was allowed to equilibrate with the ambient environmental conditions with several passes over the flight transects throughout the remainder of the day while other meteorological measurements were made.

The SDSU Sky Arrow ERA infrastructure is further described in Dumas et al. (2001) with additional details described in Zulueta et al. (2011). Two “sister” aircraft, one used in the Regional Assessment and Modelling of the Carbon Balance of Europe (RECAB) research project and one operated by the University of Alabama, have the same airframe but with slightly
modified and updated MFP instrumentation packages are also described in Gioli et al. (2006) and Hall et al. (2006) respectively.

d. Aircraft calculations

The high sampling frequency from the aircraft’s instrumentation package allows for the fluxes of mass, momentum, and energy using the eddy covariance (EC) technique (Swinbank 1951; Baldocchi et al., 1988; Loescher et al., 2006b, 2006b). This direct measure of the surface fluxes is expressed accordingly:

\[ F_{\phi} = \rho \overline{w' \phi'} = \rho \left( \overline{w} - \overline{w} \right) \left( \overline{\phi} - \overline{\phi} \right), \]  

where \( F \) is the turbulent scalar flux, \( \phi \) is the flux scalar of interest, \( \rho \) is the mean dry air density, \( w \) represents the vertical wind fluctuation, primes denote turbulent fluctuations and overbars denote ensemble averages (aircraft fluxes embody both temporal and spatial averaging).

Eddy covariance measurements from moving platforms such as aircraft are similar to that of stationary ground-based towers with the exception of the adiabatic heating correction for temperature and the measurement of the wind vectors themselves (Lenschow 1986; Leise & Masters 1993). Aircraft carry the wind measurement sensors through turbulent structures within the atmosphere. Due to the angle of attack and continuous motion (pitch, roll, yaw) of an aircraft while in flight, measurements of the velocity of the instruments (\( V_p \)), must be subtracted from the relative wind velocity (\( V_a \)) in order to determine the winds relative to the Earth’s surface (\( V \)):

\[ V = V_a - V_p. \]  

The velocity vector (\( V_a \)) was computed from pressure differences observed from the nose hemisphere (Brown et al., 1983; Crawford & Dobosy 1992; Eckman 1999) while (\( V_p \)) is obtained from a blending of both high frequency accelerometers and low frequency TANS Vector GPS
signals (Eckman et al., 1999). Calculation of these wind vectors was performed in post-processing using a NOAA designed C-program called, makepod, which incorporates the algorithms and techniques described in Leise and Master (1993), Crawford and Dobosy (1992), Eckman (1999), and Eckman et al. (1999). Calculated wind vectors were then converted and saved to the NetCDF format (Rew et al., 1997).

Tower-based EC calculations typically use an ensemble or temporal average (Baldocchi et al., 1988), however since aircraft move through the turbulent eddies, changes in aircraft speed must also be considered. Aircraft speed and vertical wind velocity are correlated and can lead to biases if only the temporal covariances are used (Crawford et al., 1993a). Wind updrafts result in aircraft acceleration as the pilot must decrease the angle of attack to maintain constant altitude, resulting in fewer data samples, while wind downdrafts result in a deceleration as the pilot must increase the angle of attack to maintain constant altitude, resulting in more data sampled.

Aircraft based EC calculations therefore require a spatial averaging technique for all variables used in the EC calculations as described in Crawford et al. (1993a) and is defined as:

$$\left[ \phi \right] = \frac{1}{ST} \sum_i \phi_i S_i \Delta t,$$

where square brackets indicate the spatial average, $\phi$ represents the variables in the covariance calculations, $S$ is the aircraft speed, subscript $i$ indicates instantaneous values, $\Delta t$ is the time increment between measurements, and overbars represent the average over the time, $T$, of the calculation segment.

Along with aircraft speed, the measurement height is also a determining factor in the length of the spatial average used for the EC computations. Turbulent eddies increase in size from the surface and longer lengths are therefore required to adequately capture all flux-carrying frequencies (wavelengths). A spatial averaging length would ideally be long enough for
adequate turbulence sampling and short enough to differentiate the surface spatial heterogeneity (LeMone et al., 2003). An ogive technique (Desjardins et al., 1989; Frihe et al., 1991; Oncley et al., 1996) can be used to determine the minimum time or length required to capture all flux-carrying frequencies and therefore an optimal spatial averaging length for the aircraft measurements. Determination of the optimal spatial average is done using an ogive function of the integrated cospectrum (Co) of the vertical wind velocity (w) and the scalar (ϕ) of interest, $C_{ow\phi}$ (Desjardins et al., 1989; Oncley et al., 1996) as,

$$Og_{ow\phi}(f_0) = \int_{0}^{f_0} C_{ow\phi}(f)df. \tag{5}$$

High-pass filtering or detrending is not done in order to include all measured fluctuations. The ogive is a cumulative covariance graph over all sampled frequencies up to the full length of the flight transect and shows the cumulative contribution of eddies of increasing size to the total flux. Computations are done over entire data windows from the shortest (few meters) to the longest (entire transect lengths). The cumulative total is equal to the covariance over the sampling time, and ogive curves that approach an asymptote at the low frequencies suggest that all flux-carrying turbulent scales are constrained within the sampling distance. The optimal averaging length is derived from the frequency at which the ogive curve approaches a constant value (Fig. 3).

The airflow past the microbead thermistor at the nose probe’s stagnation point, $T_{p1}$, is aspirated at 5 m s$^{-1}$ (Crawford & Dobosy 1992; Hacker & Crawford 1999), while $T_{p2}$ within the fast flow port is subject to the full airspeed (average of 37.0 m s$^{-1}$) of the aircraft while in flight. Initial spectral analysis of the $T_{p1}$ and $T_{p2}$ temperature signals highlighted larger high frequency signal attenuation for $T_{p1}$, probably related to the reduced airflow over the microbead thermistor at the stagnation point. The $T_{p2}$ microbead thermistor was therefore used for calculations of $H$.
(see Appendix A). Corrections for the dynamic heating of the temperature probes were applied (Crawford & Dobosy 1992). Adjustments to $H$ to account for the assumed thermodynamic expansion of air from evaporative processes consistent with the assumptions used in the Webb et al. (1980) derivation (Paw U et al., 2000) were also applied.

Resolving the frequencies carrying the largest amount of flux is often limited by the inadequate dynamic response time of the EC sensors resulting in an attenuation of the measured covariances (Moore 1986; Horst 1997; Massman 2000). This high frequency attenuation results in some flux loss and can be corrected by calculating transfer functions based on atmospheric stability and theoretical cospectra (Moore 1986; Horst 1997, 2000; Massman 2000, 2001). The flat terrain neutral cospectra described in Kaimal and Finnigan (1994) with data from the 1968 Air Force Cambridge Research Laboratory (AFCRL) Kansas experiments (Kaimal et al., 1972) were used as the theoretical cospectra. To estimate the high frequency attenuation of first-order instruments with a characteristic time constant, $\tau_c$, we used the simplified formula described in Horst (1997):

$$\frac{\langle w'\phi' \rangle_m}{\langle w'\phi' \rangle} = \frac{1}{1 + \left(2\pi n_m \tau_c \bar{u} / z\right)^{\alpha}}$$

(6)

where for $z/L \leq 0$, $\alpha = 7/8$ and $n_m = 0.085$ and for $z/L > 0$, $\alpha = 1$ and $n_m = 2.0 - 1.915/(1 + 0.5z/L)$. Here $\langle w'\phi' \rangle_m$ is the measured covariance between the scalar, $\phi$, the $\langle w'\phi' \rangle$ is the expected covariance, $z$ denotes the aircraft flight height above the surface (average height 8.48 m), $L$ is the Obukhov length (Obukhov 1946; Monin & Obukhov 1954; Obukhov 1971; Foken 2006), and $\bar{u}$ is the mean airspeed (average 37.0 m s$^{-1}$). The determined $\tau_c$ for $Tp_2$ was 0.025 s (see Appendix A) with overall spectral corrections in the range of 6.2% to 12% of the raw fluxes, while the
IRGA $\tau_c$ was 0.008 s with overall spectral corrections in the range of 4.5% to 6.3% of the raw fluxes.

The aircraft-based fluxes were calculated using the 1 km spatial averaging block (approximately 26 s average transit time) determined with Eq. (5) and associated ogive curves (Fig. 3). A 1 km overlapped moving window with an incremental step of 100 m was adopted, and data were assigned and averaged on 1 km contiguous segments aligned according to the differentially corrected aircraft GPS position, resulting in an average flux value for each 1 km spatial segment per flight pass. Each spatially aligned 1 km flux segment was then averaged over all the flight passes resulting in an average flux for each 1 km segment over the entire flight campaign. The associated uncertainties, reported as error bars in the corresponding figures, are computed as the standard error of the mean (SEM) over all flight passes ($n=29$).

Here, we use the micrometeorological convention with negative flux values representing uptake into the ecosystem from the atmosphere. The aircraft-based EC calculations and all footprint analyses were done using MATLAB (Release 2006b, The Mathworks, Natick, MA, USA).

e. Portable tower-based eddy covariance

Portable tower-based ecosystem flux measurements were also done using the EC technique (Swinbank 1951; Baldocchi et al., 1988; Loescher et al., 2006b, 2006b) with a portable tower system. The portable tower was located at the east edge of a large mangrove stand (N 25.25983°, W 112.07738°) along the DIA, and the southern end of the MNG transects (see Fig. 1) between 24 July and 27 July 2004. This was a large contiguous mangrove stand with
an upwind fetch of 750 m to the northwest, and 800 m to the west before reaching the lagoon water. The mean canopy height of the mangrove at this stand was 3.5 m.

Measurements of the three-dimensional winds and virtual temperature were done with an ultrasonic anemometer (WindMaster Pro, Gill Instruments, Lymington, UK), while CO₂ and water vapor were also made using the same model of IRGA (LI-7500, LI-COR, Inc.) as on the aircraft. Incoming PAR (LI-190SB, LI-COR, Inc.), net radiation, \( R_n \), with an aspirated Fritschen-type net radiometer (Q*7.1, REBS Inc.), air temperature and relative humidity, \( T_{air} \) and \( RH \) respectively (HMP45c, Vaisala, Helsinki, Finland), ground heat flux, \( G \), (HFT3, REBS Inc.), and soil temperature (Type-T thermocouples, Omega Engineering, Stamford, CT, USA) were also measured. An updated version of the program described in McMillen (1986, 1988), i.e., WinFlux (San Diego State University) was used to sample the portable tower-based turbulence parameters at 10 Hz with a 400-s time constant for the running mean and digital recursive filter used to estimate the turbulent fluctuations (Eq. 2). The slow response meteorological sensors were sampled at 10-s intervals and stored as half-hourly averages in a datalogger (23X, Campbell Scientific, Logan, UT). Portable tower IRGA calibrations were done in concert with the aircraft IRGA calibration and followed AmeriFlux protocols (see above).

The sonic anemometer and the IRGA were located at a height of 4.2 m above the ground. The location of the net radiometer, air temperature, and \( RH \) sensors were over the mangrove stand while ground heat flux plates were placed 2 cm below the soil surface underneath the mangrove canopy and co-located with the soil temperature probes with measurements depths of 5, 15, and 20 cm below the soil surface.

Fluxes of CO₂, \( H \), and \( \lambda E \) were calculated as half-hourly averages and a 2D coordinate rotation was used to estimate the control volume and the vertical wind velocities perpendicular to
mean streamline (McMillen 1988; Kaimal & Finnigan 1994). The fluxes were corrected for high frequency losses in the measurement system due to inadequate scalar sensor dynamic response (Moore 1986), lateral sensor separation (Kristensen & Jensen 1979), sonic anemometer and IRGA line averaging (Gurvich 1962; Silverman 1968; Kristensen & Fitzjarrald 1984), IRGA volume averaging (Andreas 1981), as well as for low frequency losses from the running mean recursive filter (400-s), and the half-hourly block averaging (Moore 1986; McMillen 1988; Kaimal et al., 1989). We used the approach of Horst (1997, 2000) and Massman (2000, 2001) for calculating transfer functions based on atmospheric stability and the theoretical cospectral curves of Kaimal et al. (1972) described in Kaimal and Finnigan (1994). We used the equivalent time constants of first order filters presented in Table 1 of Massman (2000) with the corrected equations in Table 1 described in Massman (2001). Corrections for concurrent density fluctuations of heat and water vapor were done according to Webb et al. (1980).

Quality control included statistical checks for outliers of the CO₂ and water vapor measurements, and the \( u, v, \) and \( w \) wind velocity components based on six standard deviations from their 30-min mean values and were removed before the flux calculations. Wind directions were filtered so only winds coming from the mangrove stand to the west were considered. Fluxes were discarded when \( u_* \) was less than 0.25 m s\(^{-1}\). This \( u_* \)-threshold (Goulden et al., 1996; Gu et al., 2005) was determined when above a particular \( u_* \) the effect on the net ecosystem exchange (NEE) was unchanged and was similar to the \( u_* \)-threshold of 0.21 m s\(^{-1}\) used by Barr et al. (2010) for a mangrove ecosystem in the Everglades National Park, FL, USA. This \( u_* \)-threshold was invariably associated with the nighttime and early morning low turbulence conditions, and resulted in the rejection of 46% of the total fluxes and nearly 91% of the fluxes between 2200 and 0800 MST. Data gaps of 30-min or less were linearly interpolated (Falge et
al., 2001), and gaps >30-min were filled using the online EC gap filling and flux-partitioning tool located at http://www.bgc-jena.mpg.de/~MDIwork/eddyproc/ that incorporates techniques described in Falge et al. (2001) and enhanced algorithms that consider the temporal autocorrelation of fluxes and their co-variation with meteorological variables as described in Reichstein et al. (2005). The energy balance and the degree of closure between the sum of the half-hourly $H$ and $\lambda E$, $(H+\lambda E)$, and the sum of $R_n$ and $G$, $(R_n-G)$, was used to determine system performance and the quality of the portable tower EC measurements (McMillen 1988; Aubinet et al., 2000).

f. Aircraft flight transects

Four flight transects were established to measure fluxes over the major ecosystems surrounding Lopez Mateos (Fig. 1) and are identified as CST, DIA, MNG, and TNG with details of each transect found in Table 3. Transect DIA was a northeast-southwest orientated transect starting with the desert ecosystem from the east, and extended over the mangrove and lagoon, a sandy barrier island, and ocean to the west. Flight paths along the DIA transect were directly over the portable EC tower upwind footprint located along the edge of a large mangrove stand (see Fig.1). Transect TNG was east-west oriented and similar to DIA with desert to the east and ocean to the west, except that this transect also passed through the mouth of the Boca la Soledad. Transect CST was north-south along the coastline, made 7.3 m above the wave break zone, while the north-south MNG transect was made down the middle of the lagoon and flown over the mosaic of mangroves and surrounding lagoon waters. The DIA and the MNG transects intersect over the same mangrove stand where the portable EC tower was located (see Fig. 1) with the aircraft passing within 10 m and 20 m horizontal separation at the nearest point from the portable
EC tower, respectively. The aircraft followed the terrain as close to the ground as possible at an average measurement height of 8.48 m above all the surfaces with an average ground speed of 38.2 m s\(^{-1}\). Multiple repeated passes were done to increase sampling frequency and to reduce the random flux error (Lumley & Panofsky 1964; Lenschow et al., 1994; Mann & Lenschow 1994; Mahrt 1998). All transects were flown as a continuous ‘track’ (Fig. 1) with repeated passes flown in reciprocal directions. A complete reciprocal track took approximately one hour, and most flights consisted of multiple passes with the total flight duration limited by the aircraft’s fuel capacity (approximately 3.5 hours). A total of 29 full tracks were flown during this campaign between the hours of 0530 and 1930 MST, with the flux measurement flights flown between 0930 and 1930 MST.

**g. Flux Footprints**

To relate fluxes measured by tower and aircraft to their sink/source area on the ground, a footprint model was used to describe the contribution of the surface area to the measurement at a particular location (Leclerc & Thurtell 1990; Schuepp et al., 1990; Kljun et al., 2002; Schmid 2002; Kljun et al., 2004). The flux footprint varies in size and depends on measurement height above the surface, wind vector, surface roughness, and atmospheric stability (Leclerc & Thurtell 1990). Here we used the footprint model of Kljun et al. (2004) which is a parameterization of a Lagrangian stochastic footprint model (LPDM-B, Kljun et al., 2002). The model described in Kljun et al. (2004) is a crosswind integrated footprint model that allows for rapid calculations with input parameters easily derived from common turbulence measurements. This footprint model requires the measurement height \((z)\), boundary layer height \((h)\), friction velocity \((u_*)\), standard deviation of the vertical wind \((\sigma_v)\), roughness length \((z_o)\), and the Obukhov length \((L)\).
The model is valid across a wide range of boundary-layer stabilities and measurement heights with the overall conditions of \(-200 \leq z/L \leq 1\), \(u_* \geq 0.2\) m s\(^{-1}\), and \(1\) m < \(z < h\) (Kljun et al., 2004).

The footprint parameters for the portable tower EC system were calculated for the averaging period of the flux calculations (30-min), while the aircraft parameters were calculated from the spatial average of each measurement segment (1 km) along the flight paths. The \(z_o\) can be derived from the logarithmic wind profile under neutral conditions from:

\[
    u(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_o} \right),
\]

with \(u(z)\) defined as the wind velocity at the measurement height \((z)\), and the von Kármán constant \(\kappa = 0.4\).

Flux footprints were calculated for each 1 km spatial block along each transect and evaluated to determine whether their surface sink/source area were either desert, mangrove, or ocean water. Respective ecosystems along each transect were then averaged to give the footprint estimations for each ecosystem. To compare the aircraft footprint to the portable tower footprint, the mangrove sections of the DIA and MNG transects that intersected the mangrove area measured by the portable tower were isolated from the surrounding ecosystems.

\textit{h. Determining Regional Mangrove Coverage and NDVI}

The NDVI is widely acquired from satellite- and aircraft-based platforms and recognized as an effective ecosystem-level indicator of plant health, primary productivity, and canopy capture of PAR (Tucker 1979; Sellers 1985; Box \textit{et al.}, 1989; Goward \textit{et al.}, 1991; Myneni \textit{et al.}, 1997; Vermote \& Saleous 2006). Determination of regional scale NDVI for the mangroves
within the entire Magdalena Bay was done using the Terra satellite MODIS Surface Reflectance 8-Day L3 Global SIN Grid V005 250 m resolution data product (MOD09Q1, http://modis-sr.ltdri.org/) managed by the MODIS Land Surface Reflectance Science Computing Facility (LSR SCF) (Vermote et al., 1997; Vermote et al., 2002) and distributed by the Land Processes Distributed Active Archive Center (LP DAAC), located at the U.S. Geological Survey (USGS) Earth Resources Observation and Science (EROS) Center (http://lpdaac.usgs.gov).

Each MODIS MOD09Q1 pixel has the best possible observation during an 8-day period and provides MODIS $b_1$ and $b_2$ surface reflectance that are corrected for atmospheric, aerosol interference and high cirrus clouds (Vermote et al., 1997; Vermote et al., 2002; Vermote & Saleous 2006; Vermote & Kotchenova 2008). The five minute by 2300 km swath width MODIS image was subset spatially to approximately 350 X 850 pixels or 80 X 200 km to include only the Magdalena Bay area (Fig. 4). The most temporally appropriate data set acquired for the Magdalena Bay region was between 27 July and 03 August 2004. The selected MODIS image was MOD09q1.A2004209.h07v06.005.201005422248.hdf.

The NDVI was calculated using the raw digital numbers in $b_1$ and $b_2$ from the LP DAAC MODIS product and scaled appropriately so all values are within the 16-bit signed integer image space. For our analysis and visualization of MODIS data, both MultiSpec (version 3.25.10, http://cobweb.ecn.purdue.edu/~biehl/MultiSpec/) (Biehl & Landgrebe 2002) and ERDAS IMAGINE (2010, ERDAS Inc., Norcross, GA, USA) were used. A built-in IMAGINE function for NDVI calculation was used. The NDVI was added, for visualization purposes, as a third band in the image, and an analysis was made to determine if any generalizations regarding the ranges of NDVI within the known cover types could be made in the Magdalena Bay spatial subset. The known cover types of; ocean, bay water, desert, mangroves plus water in mixed
pixels, and homogeneous mangrove were identified by inspection of the three band image. An image displaying false-color-IR was effective in determining vegetated areas (Fig. 4a). In this case the MODIS band $b_2$ is displayed in the red and band $b_1$ in both green and blue channels.

To characterize NDVI values representative of the five cover types, multiple rectangular samples, including only the desired cover type, were selected in MultiSpec and the NDVI values stored in a spreadsheet. To extract the NDVI values for all mangrove pixels, a series of rectangular selections were made along the entire length of the extant mangroves in the Magdalena Bay area, and each selection of pixels saved for further analysis. The extractions were made to exclude as many non-mangrove pixels as possible since all selected pixels had to be screened for appropriate NDVI range and the out-of-range pixels excluded prior to summation and further processing. The total number of pixels within the mangrove NDVI range of 0.3 to 0.8 was summed and multiplied by the area of each pixel to determine the total mangrove ecosystem area.

Since these values were from individually-selected, single pixels of known cover type, we chose to broaden our analysis of the mangrove-only pixels. For further analysis of the mangrove areas, we sampled all pixels within the mangroves. This more extensive sampling included some full-canopy mangrove samples and various mixtures of canopy and water, or canopy and land, water, and even bare soil. The individual pixel values were extracted from the IMAGINE-derived NDVI image and non-vegetated pixels (NDVI $\leq 0.0$) were removed as well as any pixels that had NDVI values greater than any observed in the pixel-by-pixel analysis (NDVI $> 0.79$). Based on the spherical projection of the MODIS MOD09Q1 image, each pixel was $231.66$ m X $231.66$ m or $53666.4$ m$^2$. 
3. Results

a. Aircraft system performance

The specified series of in-flight calibration maneuvers were used to determine the empirical values of calibration constants of the wind computational model (Telford et al., 1977; Lenschow 1986; Bögel & Baumann 1991; Tjernström & Friehe 1991; Leise & Masters 1993; Khelif et al., 1999; Williams & Marcotte 2000; Kalogiros & Wang 2002b; Lenschow et al., 2007; Vellinga et al., 2013). Complex flight maneuvers such as constant altitude standard rate turns (3° s⁻¹) and wind boxes were used to evaluate the results of the calibration of the aircraft’s wind vector system and post-processing routines (Telford & Wagner 1974; Telford et al., 1977; Vellinga et al., 2013). A constant altitude standard rate, 450°, counter clockwise turn flown during the calibration flight on 28 July 2004 at 1560 m ASL was assessed to verify the quality of the calibration. During an aircraft turning maneuver or circular flight path, both $V_p$ and $V_a$ are continuously changing and errors in either of these measurements manifest in errors of $V$. Since no biases were observed in either $V_p$ or $V_a$, the resulting velocity of $V$ was nearly constant throughout the entire 450° turn. The measured velocity of $V$ during this turning maneuver was $3.45 \pm 1.05$ m s⁻¹ with a measured wind direction of $264 \pm 2.38°$ true.

The ogive plots for $C_o(wCO_2)$ of all the flight legs over the DIA and MNG transects show a convergence and asymptotic shape of the ogive functions at the low frequencies suggesting that all flux-carrying turbulence scales are captured within the length of the transects, and can be used to robustly determine the spatial averaging scales (Fig. 3). An averaging length of 1 km was used as the optimal averaging length for the subsequent EC calculations. Ogive functions for the CST and TNG transects also showed similar shapes and convergence towards a spatial averaging length of 1 km (data not shown).
b. Flux Footprints

The results of the footprint analysis for the aircraft and portable tower are presented in Table 1. Overall, the average maximum footprint contribution area (xmax) and the 90% flux contribution distance (x90%) were largest over the ocean, 110.7 m and 303.2 m, respectively. As expected, the z0 was lowest over the ocean and highest over the desert which had a xmax and x90% of 94.8 m and 259.6 m, respectively. The footprint over the mangroves were a mix of continuous mangroves stands and open lagoon water and had an average xmax of 73.9 m and x90% of 217.1 m. The aircraft footprint, xmax and x90%, for the mangrove area where the portable tower was located were five times larger than the footprint of the portable tower mainly due to the differences in measurement height between the portable tower and aircraft, 4.2 m and 7.98 m respectively.

c. Aircraft Measured Spatial Variability of Fluxes and NDVI

The DIA and TNG transects cross over the various ecosystems from the desert in the east, the mangrove and lagoon system in the middle, and the ocean to the west (see Fig. 1). The dashed lines in Fig. 5 demark boundaries of the different ecosystems along the DIA and TNG transects. The DIA transect crosses over a sand covered barrier island between the mangrove lagoon and the ocean, while the TNG transect crosses through the mouth of the Boca la Soledad. The CST and MNG transects were oriented north-south with the CST over the near shore coastline just over the wave break zone, and the MNG down the middle of the mangrove and lagoon area (see Fig. 1). Differences in the fluxes of CO₂, H, and λE are clearly observed from the different ecosystem types/source areas in the DIA and TNG transects (Fig. 5).
The mangrove ecosystem had the largest CO$_2$ uptake with an average flux of -8.11 µmol CO$_2$ m$^{-2}$ s$^{-1}$ with a maximum rate of -14.7 µmol CO$_2$ m$^{-2}$ s$^{-1}$ over a large mangrove stand located between the desert and lagoon (the DIA transect and the southern end of the MNG transect). The average CO$_2$ flux of the desert ecosystem was small with average uptake of -1.32 µmol CO$_2$ m$^{-2}$ s$^{-1}$ while the coastline and near shore ocean had uptake rates of -3.48 µmol CO$_2$ m$^{-2}$ s$^{-1}$.

Average $H$ flux was highest over the desert 282 W m$^{-2}$ while the ocean had the lowest $H$ flux 50.4 W m$^{-2}$. The MNG transect had the largest range in $H$, from 59.7 W m$^{-2}$ to 249 W m$^{-2}$, which was dependent on whether the flight path was over a mangrove stand (high $H$) or lagoon water (low $H$). Small mangrove stands were interspersed within the lagoon area, which resulted in the pattern of $H$ seen in Fig. 5(h). Latent heat was highest over the mangroves (88.0 W m$^{-2}$) lowest over the desert (17.6 W m$^{-2}$), with the near shore and ocean consistently at 44.3 W m$^{-2}$.

We were able to detect differences in NDVI among the various ecosystems and surface features with the ability to resolve even narrow ecosystem borders (see Fig. 1 and Fig. 6a). Borders between ocean and lagoon waters were also distinguishable. The high resolution hyperspectral measurements were averaged to 1 km spatial resolution (Fig. 6b) to match the spatial resolution of the aircraft fluxes (Fig. 6c) for comparisons between NDVI and NEE (Fig. 7).

The NDVI over the mangroves were the highest and most variable, 0.05-0.7, while the desert was consistently low, 0.09-0.15. The range of NDVI within the desert was small, and there was no relationship found between CO$_2$ flux and NDVI while CO$_2$ flux over the mangroves showed increased uptake with higher NDVI (Fig. 7). The highest NDVI recorded was over the large homogeneous mangrove stand where the portable tower was located (see Fig. 1), while low
NDVIs were typically found from a mixed measurement of mangrove and lagoon water, or areas dominated by sand.

d. Portable tower-based mangrove flux measurements

The average diurnal pattern of CO₂ flux, energy balance, PAR, and T_air for the large mangrove stand between 24 July and 27 July 2004 is shown in Fig. 8. Low turbulence and stable atmospheric conditions prevented calculations of fluxes from 0230 to 0730 MST, however, by 0800 the mangrove stand showed strong CO₂ uptake coincident with a PAR of about 800 µmol m⁻² s⁻¹. An average maximum uptake rate of -12.3 µmol m⁻² s⁻¹ persisted from 0830 to 1330 with a peak uptake of 13.5 µmol CO₂ m⁻² s⁻¹. There was a steady decline in CO₂ uptake from 1400 to 1730 and by 1800 the mangrove ecosystem switched from uptake to CO₂ efflux though the PAR was still relatively high (about 800 µmol m⁻² s⁻¹). Nighttime CO₂ flux was consistent and averaged 3.97 µmol CO₂ m⁻² s⁻¹.

The energy balance closure for the portable EC tower was 75% with the slope of the regression between the sum of the half-hourly $H+\lambda E$ and $R_n-G$ at 0.88 with an $r^2 = 0.91$, indicating a small under estimation in $H+\lambda E$. We had good energy balance closure when compared with other FLUXNET (Baldocchi et al., 2001) sites (Wilson et al., 2002). The $R_n$ was typically symmetrical from the peak of 699 W m⁻² occurring at midday (1300 MST). The $\lambda E$ had a similar pattern to $R_n$ with a coincident peak time frame and a maximum of 229 W m⁻². The $H$ typically lagged $R_n$ and $\lambda E$ and peaked between 1330 and 1400 at 435 W m⁻². Ground heat flux was not particularly large even during the midday with a peak of 48.2 W m⁻². Values of $R_n$, $H$, and $G$ were typically negative at night and early morning from 1930 to 0700 MST. The pattern of incoming PAR was also symmetrical from the midday peak of 2064 µmol m⁻² s⁻¹ with
nighttime darkness from about 1930 to 0600 MST, coinciding with local sunset (1917 MST) and sunrise (0553 MST), respectively (data source: http://www.esrl.noaa.gov/gmd/grad/solcalc/). Air temperature was lowest just before sunrise and rose rapidly with maximum temperatures, about 26.3 °C, extending from 1030 to 1630 MST, followed by a nearly linear decline from 1930 MST to around midnight.

*f. Portable tower and Aircraft Intercomparison*

To compare the portable tower- and aircraft-based EC measurements, we isolated the 1 km spatial averaging block of the aircraft along the DIA and MNG transects within the fetch of the mangrove stand measured by the portable tower. This was done for all the passes of the DIA and MNG transects. One hour bins were created with each bin centered on the hour and all the flight passes were placed into corresponding bins depending on their time of passage past the portable tower. These were then averaged and the SEMs were determined. All the portable tower data were treated similarly by binning matching times together over the entire campaign. Each hourly portable tower bin corresponding to the aircraft measurements are compared. The intercomparison between the tower and aircraft are shown in Fig. 9. There was good agreement in $R_n$ between the aircraft and tower (Fig. 9b) with a slope of 1.21 and an $r^2 = 0.96$, showing that the aircraft $R_n$ over this mangrove area was slightly larger compared to the ground based measurements. Compared to the tower, the aircraft underestimated the CO$_2$ flux (Fig. 9a) by 25% with a slope of 0.75 and an $r^2 = 0.74$. There was an underestimation of the tower $H$ by the aircraft with a slope of 0.36 and an $r^2 = 0.55$ (Fig. 9c), while the variance in the aircraft $\lambda E$ could not be explained by variations in the tower $\lambda E$ (Fig. 9d).
g. Aircraft-derived Temporal Patterns

The multiple flights throughout this campaign occurred between sunrise and sunset with the concentration of the flux flights occurring between 0930 and 1830 MST. Aggregating the times, locations, and separating out the various ecosystems for the entire campaign, it was possible to use the aircraft to obtain a temporal flux pattern through the day among the studied ecosystems (Fig. 10). The aircraft-derived temporal patterns of CO₂ flux over the mangrove areas (Fig. 10e) were consistent with that from the portable tower in magnitude, direction, and pattern (see Fig. 8a). Though both \( H \) and \( \lambda E \) estimates were less than those found by the portable tower (Fig. 9c and 9d, i.e. lower magnitudes), the overall diurnal pattern remained consistent between the tower and the aircraft.

h. MODIS Regional Mangrove Coverage and NDVI

The NDVI values for the five cover types are presented in Table 4. With the larger amount of sampled mangrove areas, the range of NDVI of mangrove-dominated pixels was 0.300 to 0.800. Based on this NDVI range, there were 6731 pixels making up 361.23 km² of mangrove-dominated areas in the area of analysis. Using the MODIS-derived NDVI and NEE relationship derived from the aircraft flights (Fig. 7) with all the NDVI mangrove pixels (Fig. 4b), we calculated the NEE for each pixel and summed them for a total mangrove ecosystem CO₂ flux. Based on the number of mangrove-identified MODIS pixels, NEE ranged from -8.0 \( \mu \text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1} \) with an NDVI of 0.300, to -10.7 \( \mu \text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1} \) for the highest NDVI value of 0.800. Fully scaled to the number of mangrove-identified MODIS pixels and weighted by NDVI, the mid-day average CO₂ flux was -9.2 \( \mu \text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1} \), and the sum of all the pixels
over 361.23 km$^2$; the areal mid-day NEE was -526.27 t CO$_2$ hr$^{-1}$ throughout the Magdalena Bay region.
4. Discussion

a. Ecosystem fluxes

In the Magdalena Bay region the distinct borders among ecosystems over small spatial distances (see Fig. 1), and the contrast among the desert, mangrove and lagoon ecosystem fluxes of CO$_2$ is dependent on the time of year. During the summer months, desert ecosystems of Baja California Sur (BCS) have low productivity or are small sources of CO$_2$ to the atmosphere due to relatively small amounts of biomass, high temperatures, solar radiation, and low water availability (Hastings et al., 2005; Bell et al., 2012). However, desert ecosystems are not completely devoid of photosynthetic activity and drought resistant, desert evergreen species such as *L. tridentata* and *S. chinensis*, bark photosynthetic species such as *Cercidium microphyllum* and *Fouquieria splendens*, and succulent species such as *P. pringlei*, *S. thurberi*, and *O. cholla* do persist and can be productive even throughout the summer months. Though photosynthetic activity is reduced by drought conditions and high temperatures, and by drought induced loss of leaves from drought deciduous species and annual grasses, CO$_2$ uptake has been shown to occur in a similar desert scrub community outside of La Paz, BCS, Mexico, particularly in the early- and mid-morning hours (Hastings et al., 2005; Bell et al., 2012). Desert ecosystems are also reported to have significant CO$_2$ uptake annually as well as during the summer months (see e.g., Wohlfahrt et al., 2008). Photosynthesis may often offset ecosystem respiration, however during extended summer droughts, desert ecosystems have been shown to be an annual net source of CO$_2$ to the atmosphere (Hastings et al., 2005; Bell et al., 2012). This net CO$_2$ source during the summer can persist even after a precipitation event where the microbial biota can respond more quickly to small amounts of water from summer precipitation events than can the annuals and drought deciduous species (Huxman et al., 2004; Reynolds et al., 2004; Sponseller 2007).
However, major precipitation events, such as hurricanes and tropical storms, coincide with the maximum productivity of these desert ecosystems (Knapp & Smith 2001; Huxman et al., 2004; Hastings et al., 2005; Bell et al., 2012) due to the co-occurrence of moderated temperatures and increased water availability, however these events occur mainly during the fall and winter months.

The mangroves during the summer months are at their peak of productivity (Osborne 2000; Chavez 2006) with peak measured NEE during the middle of the day. Though the mangrove trees have an apparent abundance of water, they are under water limitation due to the highly saline lagoon waters and have developed physiological coping mechanisms which gives them high water use efficiency (Alongi 2009). The combination of high solar radiation, warm temperatures, and abundant nutrient availability results in the high productivity of these mangrove ecosystems (Alongi 2009).

Here, aircraft-based EC measurements i) demonstrate a net CO$_2$ uptake for a large homogeneous mangrove stand with validation of both direction and magnitude of the flux from the portable tower-based measurements, and ii) differentiate different sources areas, i.e., open water, mangrove, lagoon and desert ecosystems. Though we did not specifically measure the fluxes from the lagoon itself (tower-based or surface-layer based), a subsequent study measured the differences in the CO$_2$ partial pressure ($\Delta pCO_2$) between the atmosphere and the lagoon surface waters (just north of Lopez Mateos) with a shower equilibrator (Broecker & Takahashi 1966) and a non-dispersive IRGA (LI-840, LI-COR Inc.) showed the lagoon to be a source of CO$_2$ to the atmosphere (Ikawa 2012). This corresponds with previous reports that lagoon waters surrounding mangrove forests are small but net CO$_2$ sources (Borges et al., 2003). Since the aircraft integrates fluxes from mangrove and lagoon over the 1 km spatial average, the
integration of the mangrove and lagoon flux signals will show variable and typically smaller 
NEE rates when compared to the mangrove stands with little or no standing water. However, 
due to the very high productivity of the mangroves, daytime NEE exceeds ecosystem respiration, 
the mangrove-lagoon complex was still a significant net CO₂ uptake. Barr et al. (2010) reported 
maximum daytime uptake rates of -20 to -25 μmol CO₂ m⁻² s⁻¹ of a large stature mangrove forest 
in the western Everglades National Park, nearly two times the maximum uptake rates reported 
here.

The near shore ocean CO₂ uptake is likely due to several factors that include nutrient 
outflow from the bay and coastal upwelling of deep, nutrient rich waters. This study is 
consistent with other studies of near shore ocean uptake of CO₂ (Hales et al., 2005; Cai et al., 
2006). The aircraft measurements showed an uptake of CO₂ along the coastline and if the flight 
transect was extended farther into the Pacific, we would have been able to determine the extent 
of the near shore uptake. It is also important to note that increased CO₂ uptake at the shore break 
may be due to increased air-sea exchange through bubble entrainment (Farmer et al., 1993; 
Asher et al., 1996; Zhang et al., 2006; McNeil & D'Asaro 2007). This airborne technique can be 
used to elucidate the ecosystem processes at the terrestrial-coastal-ocean interface, which have 
been receiving increasing attention in both advances in science and in policy (Vargas et al., 
2012).

b. Tower and Aircraft Intercomparison

Proper planning of tower sites and flight paths ensure favorable comparisons between 
tower and aircraft measurements over relatively homogenous sampling areas (e.g., Zulueta et al., 
2011). However, even the most carefully planned sites and campaigns have limitations for direct
tower and aircraft intercomparisons. Since aircraft-based EC measurements integrate over large spatial areas and are in constant transition across the landscape, footprint mismatches and different averaging temporal and spatial scales make direct comparisons between tower and aircraft challenging to interpret. Conceptually, the aircraft footprint is a moving and continuously integrating swath of the upwind landscape with approximate dimensions equal to the footprint length times the distance of each measurement block, and therefore several times larger than the tower footprint even if the aircraft was flown at the same height as the tower. Moreover, because the aircraft transits across the landscape, its footprint for each averaging block are a ‘snapshot’ of the upwind source area while the tower footprint is sampling continually over a typically better defined and constrained area. The temporal and spatial scales from these two approaches still can be linked to make comparisons.

Intercomparisons between aircraft- and tower-based EC measurements have been made in previous studies over a range of different ecosystems (Desjardins et al., 1989, 1992; Kelly et al., 1992; Crawford et al., 1993b, 1996b; Oechel et al., 1998; Gioli et al., 2004; Isaac et al., 2004; Zulueta et al., 2011) and the results presented here are consistent with those studies. An early study using data from the 1987 First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE) campaign (Kelly et al., 1992) found aircraft measurements of $H$ to be 20-50% lower than ground-based measurements, and aircraft measurements of $\lambda_E$ was lower than ground-based measurements at high $\lambda_E$ and greater at low $\lambda_E$ values. Desjardins et al. (1992) in an analysis of the National Research Council (NRC) Twin Otter data from the 1987 and 1989 FIFE campaigns found $H$ and $\lambda_E$ in agreement at the University of Nebraska-Lincoln (UNL) site for both years, but at the Argonne National Laboratory (ANL) site, the aircraft underestimated $H$ by about 40% in both years and
overestimated $\lambda E$ by about 14% in 1989. A later study by Crawford et al. (1996a) found aircraft measurements underestimated $H$ by 10-20% and overestimated $\lambda E$ by 25%, while Desjardins et al. (1997) also found that aircraft measurements underestimated $H$ and overestimated $\lambda E$, but the sum of $H+\lambda E$ however were comparable to the ground-based measurements. In general, these earlier aircraft-tower intercomparisons found that aircraft measurements of $H$ were less than surface-based measurements by 10-50% with mixed results for $\lambda E$. Fluxes of CO$_2$ weren’t always measured in these studies, however, a study with this particular Sky Arrow and instrumentation package and a pair of portable EC towers over a homogeneous upwind surface area showed a near 1:1 relationship between the aircraft- and tower-based CO$_2$ fluxes with slight underestimations of $H$ and $\lambda E$ (Zulueta et al., 2011). The Sky Arrow platform resolved different surface processes and ecosystem types across the arctic landscape providing the observations that facilitated spatial scaling approaches (Zulueta et al., 2011). With this study and others found elsewhere, we have high confidence in our ability to denote ecosystem boundaries and hence also in determining our source areas. The future of linking these approaches can be used for more complex spatial data assimilation approaches for larger areal estimates, and to be used to identify terrestrial-ocean interfaces, and be used for targets of opportunity, e.g., flooding, fires, land use changes.

Flux measurement sites are typically selected within predominant or ecologically important ecosystems within a region. Careful attention is put towards representativeness of a site for a certain area and usually within a structurally and functionally homogeneous ecosystem as possible. However, even ecosystems considered homogeneous, such as desert, mangroves, and arctic tundra show a large degree of spatial variability in structure and function over short distances of kilometers or less (Walker et al., 1994; Shaver et al., 1996; Hinkel et al., 2001;
Vourlitis et al., 2003; Riedel et al., 2005). In areas where an ecosystem is heterogeneous or where multiple ecosystems converge such as the complex coastal ecosystem of Magdalena Bay, selecting a site with a representative source area is clearly difficult, and a bias in site selection can lead to biases in scaling flux estimates to the region if only tower information is used (Kelly et al., 1992). The ability to differentiate various landscape features from aircraft-based measurements in terms of their surface fluxes have been demonstrated here as well as across a range of different ecosystems from grasslands (Desjardins et al., 1992; Kelly et al., 1992; Gioli et al., 2004), broadleaf, evergreen, and boreal forests (Crawford et al., 1996b; Gioli et al., 2004), agricultural fields (Desjardins et al., 1989; Gioli et al., 2004; Isaac et al., 2004), arctic tundra (Brooks et al., 1996; Oechel et al., 1998; Zulueta et al., 2011), and subtropical coastal environments (Crawford et al., 1993b). Whereas towers are limited in spatial area coverage, aircraft have the ability to measure across large expanses of landscape and in this case aircraft flux measurements may be more accurate an assessment of the landscape or regional fluxes than extrapolating solely from tower-based measurements. However, aircraft are limited in temporal coverage and measurements tend to be biased as flight regulations and safety limit aircraft flights to daylight hours, or ‘good’ weather conditions. The aircraft- and tower-based flux measurements are therefore complementary and combined provide for both a temporal and spatial assessment of surface fluxes and a means for improved landscape and regional scaling.

c. Towards Regional Scale Flux Estimates

Measurements of regional scale surface fluxes (at scales of 100s of km²) and validation of remotely sensed satellite data products and model outputs are necessary to understand the processes which determine a region's influence on local and global carbon budgets and climate
(Dolman et al., 2009; Desai et al., 2010). Numerous measurements of landscape scale EC fluxes are actively underway globally through collaborative efforts of FLUXNET (Baldocchi et al., 2001). However, despite the hundreds of towers within the FLUXNET community, the majority are located within the Northern hemisphere and temperate climates (Williams et al., 2009) and even these are not randomly nor uniformly distributed. Their locations were not decided in a coordinated effort to facilitate the spatial scaling of fluxes. Therefore, the spatial density of representative flux towers is still relatively sparse and the range of spatial variability inadequately sampled to scale with confidence from ecosystem-to-landscape-to-region-to-continent. This is particularly true in harsh environments, and/or remote areas such as semi-arid, arid, desert, tropical, and arctic ecosystems. In many of these areas, infrastructure is limiting and challenging logistics make flux measurements not easily obtainable. The mismatch between the chamber- and tower scales of measurement, and a mismatch between ecosystem-level scales and modeled grid-size estimates of fluxes, along with the inability to determine mechanistic linkages among these scales has remained a major impediment to validating model algorithms and outputs as well as quantifying regional carbon fluxes with confidence.

Nonetheless, regional estimation of fluxes, including those in remote areas, can be improved by the aircraft methodology presented here. This methodology is an effective bridge from plot to ecosystem to regional scales and for use in extrapolation, interpolation, and model and satellite verification (Desjardins et al., 1997; Ogunjemiyo et al., 1997; Oechel et al., 2000; Beringer et al., 2011b) and can be used to establish baseline measurements that can be coupled with satellite derived data products.

The region may have surface and landscape features with much finer spatial scales than the minimum resolvable flux scale of the aircraft (1 km), however, the integrated nature and
continuous spatial coverage provide by aircraft are valuable in assessing the spatial variability across much larger spatial areas than towers alone. The spatial variability of fluxes among these complex coastal ecosystems outlines the importance of the need for measurements that can sufficiently cover the regional spatial variability. Validation of satellite data and model outputs in this ecosystem would be problematic, if not biased, if based on extant and typical tower flux values and locations. A common method of validation of modeled or satellite derived data are from point measurements from EC tower sites (Sasai et al., 2005; Turner et al., 2005; Heinsch et al., 2006; Sims et al., 2006; Sasai et al., 2007), and an intensive validation of the MODIS data products have been undertaken with the goal to improve the MODIS algorithms (Cohen et al., 2003; Turner et al., 2005). Even though tower-based measurements are temporally explicit, their spatial coverage is quite small compared to MODIS products, resulting in significant data gaps to scaling from tower- to regional- and global-scale estimates.

There is increased need to scale ecological processes and their abiotic drivers to larger and larger spatial scales for use in ecology, as well as for the use of policymakers (NRC 2001, 2003). Advancing scaling strategies are central to new emerging ecological networks (Marshall et al., 2008; Peters et al., 2008; Schimel et al., 2011), and multidisciplinary campaigns like CarboEurope Regional Experiment Strategy (CERES) (Dolman et al., 2006; Dolman et al., 2009) and Savanna Patterns of Energy and Carbon Integrated Across the Landscape (SPECIAL) (Beringer et al., 2011a; Beringer et al., 2011b) have been initiated using similar technologies and methodologies presented here. Use of light aircraft for near-surface EC flux estimates increases the spatial sampling density from tower-based footprints to larger areas, and used in conjunction with towers, remotely sensed products, and ecosystem models enables a better understanding in the patterns and controls on regional fluxes. Even at micro-meteorological ideal tower sites
Aircraft based measurements, while limited in temporal scale (made during a few key times per season) provide important, spatially integrated information to assess questions of spatial homogeneity in the source area footprints. This includes, for example, analyses and data presentation of physiological responses, including NEE, to diurnal environmental conditions, water stress, and other environmental controls and opens the way for additional analyses such as light response curves for the diversity of land surface types encountered along continuous flight lines of 10s to 100s of km. Diurnal and seasonal patterns of NEE, H, λE, and energy balance across large inaccessible areas can be determined with repeated aircraft flights and such an approach can alleviate the need for multiple towers at these, often remote, areas.

d. Rapidly evolving aircraft technologies

Aircraft-based flux measurement technologies continue to rapidly evolve. Since the certification of the SDSU Sky Arrow, there have been numerous improvements and advancements in the electronics within the MFP (see e.g., Hall et al., 2006), development of fast ultrasensitive temperature sensors (see e.g., French et al., 2001) not requiring the complex corrections described here, implementation of advanced integrated inertial measurement units and GPS (IMU-GPS) electronics (see e.g., Garman et al., 2006; Hall et al., 2006; Vellinga et al., 2010; Vellinga et al., 2013), as well as more durable and robust BAT probes and pressure spheres (see e.g., French et al., 2004; Eckman et al., 2007). New wind tunnel tests (Garman et al., 2006; Dobosy et al., 2013) and calibration techniques and procedures (Garman et al., 2006; Garman et al., 2008; Vellinga et al., 2013) have improved the accuracy of wind vector
calculations, while research into flux disaggregation methods (Ogunjemiyo et al., 2003; Kirby et al., 2008; Hutjes et al., 2010), and surface flux mapping techniques (e.g. Mauder et al., 2008) have improved associating surface fluxes to landscape elements in heterogeneous areas. Analysis strategies like the flight-path segmentation presented in Vellinga et al. (2010) allow for regional scale estimates of heterogeneous terrain from aircraft-based fluxes based on landscape characteristics.

Miniaturization of electronics have also allowed for even smaller airborne platforms capable of flux measurements such as microlight aircraft (see e.g., Metzger et al., 2011; Metzger et al., 2012) and even mini unmanned aerial vehicles (UAVs see e.g. Spiess et al., 2007; van den Kroonenberg et al., 2008; Thomas et al., 2012). The latter potentially addressing the VFR flight limitations of manned aerial vehicles by being able to operate under marginal conditions, at night, or even under severe conditions or extremely remote locations considered either too dangerous or not feasible for manned aerial vehicles. Though the use of UAVs within the national airspace system is extremely limited at this time, aviation policies, procedures and standards are being worked on by the various government agencies in order to accommodate and progress along with the rapidly evolving aircraft technologies.
5. Conclusion

Aircraft based EC fluxes of CO₂, water vapor, energy, momentum as well as low level remote sensing were successfully measured over the structurally and functionally complex coastal region of Magdalena Bay, Baja California Sur, Mexico. Our results demonstrate that the Sky Arrow aircraft was effective in characterizing and distinguishing land-atmosphere fluxes at spatial resolutions of 1 km across different heterogeneous landscapes. The increased sampling density of the aircraft provides fluxes at spatial resolutions that cannot be achieved by fixed location ground-based approaches alone. Aircraft based EC techniques are a valuable tool to estimate regional scale fluxes particularly in areas that are logistically challenging or remote. Though aircraft can provide increased spatial sampling density, its temporal coverage is limited by flight rules and regulations, aircraft flight endurance, weather conditions, as well as cost of deployment and operation. Therefore, aircraft-based measurements provide a strong compliment to tower-based EC and provide insight in the spatial distribution and patterns of fluxes. Combined with satellite-based information (e.g., MODIS) the aircraft approach yields measurements that can effectively be used to instruct, verify, and validate large-scale estimates from ecosystem process models as well as other satellite derived data products, and be instrumental in constraining the uncertainty in regional scale fluxes.
APPENDIX A

Aircraft Temperature Sensor Signal Attenuation and Correction

Inadequate dynamic frequency response of EC sensors result in some flux loss due to high frequency attenuation of the measured variances and covariances (Moore 1986; Horst 1997, 2000; Massman 2000, 2001). Typically, spectral corrections are based on analytical methods requiring the sensor time constant to be known (Moore 1986), or on determining the sensor attenuation transfer function assuming spectral and cospectral similarity with a non-attenuated variable (e.g., sonic temperature). In our case, none of these approaches were applicable to estimate flux loss and correct the H fluxes since the theoretical time constant of temperature sensors is reported in still air, and cannot be assumed valid in actual flight conditions. Since there is no non-attenuated reference variable, we adopted a fully data-driven approach based on isolating flight portions in different atmospheric stability conditions (i.e., neutral and unstable), computed power spectra, and retrieved the time constant by minimizing the mean square error the match between observed and theoretical spectral shapes in the inertial subrange region (e.g., Kaimal & Finnigan 1994). Though the two aircraft microbead thermistors, Tp1 and Tp2, were the same sensor type, they were located in two different physical locations within the BAT probe hemisphere and therefore aspirated differently while in flight. The signal attenuation was expected to be different due to Tp1 being in the center and aspirated at 5 m s\(^{-1}\), while the Tp2 sensor within the fast flow port was aspirated at true airspeeds. Power spectra were analyzed to determine the signal attenuation of each sensor.

Normalized power spectra were first created by identifying two homogeneous surfaces within the study region, ocean and desert sand. During the flights the atmospheric stability over the ocean was always in near neutral conditions, while over the desert sand surface was in
unstable conditions. A total of 32 transects was available over the ocean, and 30 over the desert with four exclusions from each surface due to sporadic low frequency contributions. Power spectra highlighted a larger attenuation in the Tp1 sensor particularly in neutral conditions where smaller eddies (i.e., higher frequencies) are relatively more important (Fig. A1). Therefore, the less attenuated Tp2 sensor was selected for calculations of H.

A first-order gain transfer function, $H_c(f)$, describes the dynamic frequency response of the aircraft temperature sensors, given by

$$H_c(f) = \frac{1}{1+(2\pi f \tau_c)}, \quad (A1)$$

where $\tau_c$ is the time constant, and $f$ is the sampling frequency. The observed attenuated power spectra, $S(f)$, can therefore be expressed as

$$S(f) = \frac{S_0(f)}{H_c(f)}, \quad (A2)$$

where $S_0(f)$ is the real power spectra, and $H_c(f)$ is the transfer function (Equation A1).

The -5/3 power law represents the theoretical spectral shape within the inertial subrange region (Kolmogorov 1941; Obukhov 1941). To constrain the fit in the inertial subrange we used the region between 1-10 Hz, excluding higher frequencies due to possible aliasing and assumed a null attenuation for signals <1 Hz. The $\tau_c$ was estimated as the best fit between corrected and theoretical spectra at 0.022 s and 0.028 s, for neutral and unstable conditions, respectively. The average value of 0.025 s was used for the subsequent data processing. Figure A2 shows the observed and corrected power spectra corresponding to the neutral ($\tau_c=0.022$ s) and unstable ($\tau_c=0.028$ s) conditions, and Figure A3 shows the uncorrected and corrected ($\tau_c=0.025$ s) Tp2 power spectra for a single flight.
Acknowledgements

This work was funded by grants from the National Science Foundation under OISE-0072140, DGE-0139378, and with funding for the initial development of the Sky Arrow under DBI-9604793. We acknowledge the support of W. T. Lawrence by the NOAA funded Center of Excellence in Remote Sensing project, City College of New York. The authors thank the Centro de Investigaciones Biológicas del Noroeste for their Baja CaliforniaSur support, and especially the invaluable help from L. Galvan, A. Michel, L. Miller, A. Mendieta, J. L. Leon de la Luz. We are grateful to N. Kljun for providing us with MATLAB code for her footprint model, and ArcGIS assistance from J. Isles. We also thank TerraMetrics for providing the TruEarth\textsuperscript{®} 15-meter resolution satellite imagery. This paper has benefited by conversations with R. Hutjes, J. Elbers, R. Leuning, R. Desjardins, and S. Verma, as well as comments from H. Loescher, M. Mauder, D. Baldocchi, and several anonymous reviewers. The authors would like to dedicate this work to the late T. L. Crawford of the Field Research Division of the NOAA whose pioneering work and dedication to the use of small research aircraft has been instrumental in the development of the Sky Arrow.
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Table 1. Footprint estimations for each ecosystem type along the aircraft flight transects with the footprint estimation of the portable tower included for comparison. Each flight transect is separated based on their respective desert, mangrove, and ocean sections and combined together to get the average footprint estimations. The $z$ is the measurement height, $u^*$ the friction velocity, $w_s$ is average horizontal wind speed, $\sigma_w$ is the standard deviation of the vertical wind, $z_0$ the roughness length, $x_{max}$ is the peak contribution distance of the footprint function, and $x_{90\%}$ is the upwind distance from the measurement location where 90% of the flux contribution is included within the footprint.

<table>
<thead>
<tr>
<th>Section</th>
<th>$z$ (m)</th>
<th>$u^*$ (m s$^{-1}$)</th>
<th>$w_s$ (m s$^{-1}$)</th>
<th>$\sigma_w$ (m s$^{-1}$)</th>
<th>$z_0$ (m)</th>
<th>$x_{max}$ (m)</th>
<th>$x_{90%}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Desert</td>
<td>10.47 ± 1.66</td>
<td>0.51 ± 0.13</td>
<td>5.35 ± 1.49</td>
<td>0.59 ± 0.09</td>
<td>0.68 ± 0.36</td>
<td>94.79 ± 23.42</td>
<td>259.63 ± 64.15</td>
</tr>
<tr>
<td>Mangrove</td>
<td>8.05 ± 1.22</td>
<td>0.46 ± 0.15</td>
<td>5.14 ± 1.36</td>
<td>0.52 ± 0.12</td>
<td>0.54 ± 0.32</td>
<td>79.26 ± 17.12</td>
<td>217.11 ± 46.91</td>
</tr>
<tr>
<td>Ocean</td>
<td>7.54 ± 1.22</td>
<td>0.26 ± 0.06</td>
<td>6.71 ± 1.19</td>
<td>0.25 ± 0.04</td>
<td>0.15 ± 0.1</td>
<td>110.7 ± 21.74</td>
<td>303.23 ± 59.54</td>
</tr>
<tr>
<td>Mangrove (near Portable Tower)</td>
<td>7.98 ± 0.89</td>
<td>0.47 ± 0.1</td>
<td>5.07 ± 1.26</td>
<td>0.52 ± 0.1</td>
<td>0.52 ± 0.2</td>
<td>78.22 ± 14.63</td>
<td>214.27 ± 40.07</td>
</tr>
<tr>
<td>Portable Tower</td>
<td>4.2</td>
<td>0.44 ± 0.17</td>
<td>2.17 ± 0.89</td>
<td>0.52 ± 0.18</td>
<td>0.57 ± 0.12</td>
<td>38.09 ± 1.6</td>
<td>104.34 ± 4.39</td>
</tr>
</tbody>
</table>
Table 2. Details and specifications for the SDSU Sky Arrow 650TCN ERA.

<p>| | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Powerplant</strong></td>
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</tr>
<tr>
<td>Manufacturer</td>
<td>Bombardier Rotax</td>
</tr>
<tr>
<td>Model</td>
<td>912F</td>
</tr>
<tr>
<td>Power Output</td>
<td>59.6 kW @ 5800 RPM</td>
</tr>
<tr>
<td><strong>Propeller</strong></td>
<td></td>
</tr>
<tr>
<td>Manufacturer</td>
<td>Hoffman</td>
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<tr>
<td>Description</td>
<td>2-blade, fixed pitch</td>
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<tr>
<td><strong>Dimensions</strong></td>
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<tr>
<td>Length</td>
<td>8.15 m†</td>
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<tr>
<td>Height</td>
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<td>Wing Span</td>
<td>9.68 m</td>
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<td><strong>Weights</strong></td>
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<td>Empty Weight</td>
<td>460 kg</td>
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<td>Maximum Takeoff Weight</td>
<td>650 kg</td>
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<td>Usable Load</td>
<td>190 kg</td>
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<td><strong>Capacities</strong></td>
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<td>Usable Fuel</td>
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<tr>
<td>Endurance</td>
<td>4.25 h</td>
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<tr>
<td><strong>Performance</strong></td>
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<tr>
<td>Cruise Speed</td>
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<tr>
<td>Stall Speed</td>
<td>21 m s(^{-1})</td>
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<td>Ceiling</td>
<td>4115 m</td>
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<tr>
<td>Takeoff Distance</td>
<td>240 m</td>
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<tr>
<td>Landing Distance</td>
<td>135 m</td>
</tr>
</tbody>
</table>

† With BAT-probe attached, 7.60 m without
Table 3. Flight line transect details and GPS endpoints for each flight transect. See Fig. 1 for visual representation of each transect over the study area.

<table>
<thead>
<tr>
<th>Transect ID</th>
<th>Endpoint Coordinates 1 (WGS84, Decimal Degrees)</th>
<th>Endpoint Coordinates 2 (WGS84, Decimal Degrees)</th>
<th>Length (km)</th>
<th>Orientation</th>
</tr>
</thead>
<tbody>
<tr>
<td>DIA</td>
<td>Lat: N 25.19060°</td>
<td>Lat: N 25.305484°</td>
<td>24.6</td>
<td>ENE-WSW</td>
</tr>
<tr>
<td></td>
<td>Lon: W 112.20184°</td>
<td>Lon: W 111.989563°</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TNG</td>
<td>Lat: N 25.267507°</td>
<td>Lat: N 25.282829°</td>
<td>19.5</td>
<td>East-West</td>
</tr>
<tr>
<td></td>
<td>Lon: W 111.996474°</td>
<td>Lon: W 112.190790°</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CST</td>
<td>Lat: N 25.311325°</td>
<td>Lat: N 25.397138°</td>
<td>9.3</td>
<td>North-South</td>
</tr>
<tr>
<td></td>
<td>Lon: W 112.132006°</td>
<td>Lon: W 112.115517°</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MNG</td>
<td>Lat: N 25.398526°</td>
<td>Lat: N 25.249593°</td>
<td>16.3</td>
<td>North-South</td>
</tr>
<tr>
<td></td>
<td>Lon: W 112.087637°</td>
<td>Lon: W 112.078126°</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 4. NDVI statistics for five cover types within the Magdalena Bay region calculated from a single MODIS MOD09Q1 8-day composite 250m bands $b_1$ and $b_2$ image. Pixels were selected through visual identification of the cover type and manual sampling of multiple pixels within those cover types using MultiSpec (Biehl & Landgrebe 2002). Clouds, fog, or sunglint precluded the use of many water-dominated pixels.

<table>
<thead>
<tr>
<th>Cover Type</th>
<th># of Discrete Pixels</th>
<th>NDVI</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Average</td>
<td>Min.</td>
<td>Max.</td>
</tr>
<tr>
<td>Ocean</td>
<td>19</td>
<td>-0.727</td>
<td>-1.00</td>
<td>-0.514</td>
</tr>
<tr>
<td>Bay Water</td>
<td>1590</td>
<td>-0.825</td>
<td>-1.00</td>
<td>-0.013</td>
</tr>
<tr>
<td>Desert</td>
<td>3520</td>
<td>0.209</td>
<td>0.151</td>
<td>0.260</td>
</tr>
<tr>
<td>Mangroves + Water</td>
<td>42</td>
<td>0.584</td>
<td>0.469</td>
<td>0.790</td>
</tr>
<tr>
<td>Homogeneous Mangrove</td>
<td>69</td>
<td>0.635</td>
<td>0.463</td>
<td>0.757</td>
</tr>
</tbody>
</table>
Fig. 1  Puerto Aldofo Lopez Mateos in the northern region of Magdalena Bay. The four flight transects are identified as “DIA”, “TNG”, “CST”, and “MNG” (see Table 3 for transect details). Transect NDVI is derived from the onboard hyperspectral instrument with an effective pixel size of 1.78 m x 5.06 m. Inset image (a) shown to emphasize the ability to resolve fine surface details with the onboard hyperspectral instrumentation. The closest approach of the aircraft to the portable EC tower along the DIA and MNG transects were 10 m and 20 m horizontal separation, respectively. Base imagery courtesy of TerraMetrics.
Fig. 2  Photograph of the SDSU Sky Arrow 650TCN ERA while parked at the Lopez Mateos airstrip. Locations of the sensors are shown. The BAT-probe (inset) is on the nose of the aircraft while the four TANS Vector attitude GPS antennas are located on the center of each wing, top of the engine cowling, and top of the horizontal stabilizer. The data acquisition system and computers are located behind the rear seat. Downward looking sensors are located in the two view ports in the bottom of the fuselage of the aircraft.
Fig. 3. Ogive plots for all the flux flights (0930-1830 MST) over the MNG and DIA transects (see Fig. 1). The solid vertical line at 1 km shows that a spatial average of 1 km is appropriate to capture nearly all turbulent scales for the flux calculations. The dashed line is the 10 km spatial scale, and the end of the graph is the total length of each transect.
Fig. 4 Satellite imagery of the entire Magdalena Bay region derived from MODIS MOD09Q1 and aircraft data. False color image (a) using MODIS near-infrared spectral reflectance for red and MODIS red spectral reflectance for both green and blue image colors emphasizes the mangroves, and highlights vegetation density in increasing intensities of red due to the strong reflectance of near-infrared light by foliage. The calculated mangrove NDVI (b) based on the MOD09Q1 data, and mangrove CO$_2$ flux (c) in µmol CO$_2$ m$^{-2}$ s$^{-1}$ derived from the CO$_2$ flux and NDVI relationship determined by the aircraft measurements (see Fig. 7). Both (b) and (c) images use MODIS near-infrared spectral reflectance values as a grey-scale background, with either calculated values of NDVI (b) or CO$_2$ flux (c) for the thematic color code overlay of the vegetated mangrove areas. The pixelation within the figure is the native pixelation of the MODIS image.
Fig. 5 Average fluxes of $CO_2$, $H$, and $\lambda E$, across the DIA, TNG, MNG, and CST aircraft flight transects (see Fig. 1). The dashed vertical lines indicate ecosystem boundaries and the dotted line indicates the sandy barrier island. The closest approach of the aircraft to the portable EC tower along the DIA and MNG transects were 10 m and 20 m horizontal separation, respectively. Shading indicates the SEM.
Fig. 6  CO$_2$ flux and NDVI along the DIA transect. High spatial resolution NDVI measurements (a) from the aircraft can distinguish fine scale differences in surface type, however since the aircraft cannot resolve fluxes less than the spatial averaging block (1 km), the high resolution NDVI is averaged with the same 1 km spatial block (b) as the CO$_2$ flux (c). The map is presented to show the correspondence of the NDVI and CO$_2$ flux to the surface along the DIA transect. Shading in (c) indicates the SEM.
Fig. 7 Comparisons of CO₂ flux and NDVI along the mangrove covered sections along the DIA transect. There is an increase in the CO₂ uptake of the mangroves with increasing NDVI. Vertical and horizontal bars indicate SEMs.
Fig. 8 Portable tower measurements of (a) CO₂ flux, (b) surface energy balance, (c) PAR, and (d) $T_{air}$ (see Fig. 1. for tower location). A comparison of tower- and aircraft-based EC CO₂ flux is presented in (a). Vertical bars indicate SEMs.
Fig. 9 Intercomparison between the portable tower and the aircraft over the same mangrove stand along the DIA and southern end of the MNG transect (see Fig. 1 for tower location). Aircraft data are one hour bins of 1 km spatial blocks corresponding to the time of passage past the portable tower. The portable tower data are binned accordingly by matching times. The solid diagonal line is the 1:1 line, and vertical and horizontal bars are the SEM.
Fig. 10 Aircraft derived temporal patterns of CO$_2$ flux, $H$, $\lambda E$, and $R_n$ along the desert, mangrove, and ocean sections of all flight transects.
**Fig. A1** Normalized power spectra for Tp₁ and Tp₂ sensors in unstable (desert transect) and neutral (sea transect) conditions. Grey lines are spectra for single individual transects, the solid black line is the average spectra, and the dashed line is the -5/3 slope power law.
Fig. A2  Normalized observed and corrected power spectra for the Tp2 sensor in unstable (a) (desert transect) and neutral (b) (sea transect) conditions. Grey lines are spectra for single individual transects, the solid red line is the average spectra, the solid blue line is the corrected spectra, and the dashed line is the -5/3 slope power law.
Fig. A3  Normalized average raw power spectra (solid black line), and time-response corrected ($\tau_c=0.025$ s) spectra (solid gray line) for $T_p2$ for a single flight on 27 July 2004. The dashed line is the -$5/3$ slope power law.
Chapter 2

Aircraft-derived regional scale CO₂ fluxes from vegetated drained thaw-lake basins and interstitial tundra on the Arctic Coastal Plain of Alaska

Published in Global Change Biology

Citation:
Aircraft-derived regional scale CO₂ fluxes from vegetated drained thaw-lake basins and interstitial tundra on the Arctic Coastal Plain of Alaska

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Abstract

The landscape surface of the Barrow Peninsula of Alaska is a mosaic of small ponds, thaw lakes, different aged vegetated drained thaw-lake basins (VDTLBs), and interstitial tundra which have been dynamically formed by both short- and long-term processes. We used a combination of tower- and aircraft-based eddy covariance measurements to characterize the spatial and temporal patterns of CO₂, latent, and sensible heat fluxes along with MODIS NDVI, and were able to scale the aircraft-based CO₂ fluxes to the 1802 km² Barrow Peninsula region. During typical 2006 summer conditions, the midday hourly CO₂ flux over the region was 2.04 × 10⁵ kg CO₂ h⁻¹. The CO₂ fluxes among the interstitial tundra, Ancient, and Old VDTLBs, as well as between the Medium and Young VDTLBs were not significantly different. Combined, the interstitial tundra and Old and Ancient VDTLBs represent 67% of the Barrow Peninsula surface area, accounting for 59% of the regional flux signal. Although the Medium and Young VDTLBs represent 11% of the surface area, they account for a large portion, 35%, of the total regional flux. The remaining 22% of the surface area are lakes and contributed the remaining 6% of the total regional flux. Previous studies treated vegetated areas of the region as a single surface type with measurements from a few study sites; doing so could underestimate the regional flux by 22%. Here, we demonstrate that aircraft-based systems have the ability to cover large spatial scales while measuring the turbulent fluxes across a number of surfaces and combined with ground- and satellite-based measurements provide a valuable tool for both scaling and validation of regional-scale fluxes.

Keywords: aircraft, arctic, eddy covariance, flux, NDVI, normalized difference vegetation index, tundra, vegetated drained thaw-lake basin

Received 7 January 2011 and accepted 4 March 2011

Introduction

The arctic climate is changing and is expected to experience shifts in precipitation patterns and unprecedented increases in temperature (Serreze et al., 2000; New et al., 2001; Comiso, 2003; ACIA, 2005). Predicted changes in terrestrial ecosystems include a temperature rise of 3–7 °C in the summer and 4–9 °C in the winter, hypothesized to decrease periods of snow cover and altering the tundra thermal regime, length of growing seasons, deepening of the soil active layer depth, increasing net primary productivity, and accelerating permafrost thawing and degradation leading to subsequent hydrological changes (Serreze et al., 2000; Comiso, 2003; McDonald et al., 2004; ACIA, 2005; Hinzman et al., 2005; Smith et al., 2005; Euskirchen et al., 2006, 2009).

Increased active layer depths and degrading permafrost in the Arctic have global implications due to vast stores of approximately 495.80 Pg of organic carbon within the first 1 m of soil, and a total of 1672 Pg of stored carbon in the soils of northern permafrost regions (Schuur et al., 2008; Tarnocai et al., 2009). There is a large potential for this region to become a significant positive feedback on atmospheric CO₂ concentration (Oechel et al., 1993) if nutrient dynamics and depth of thaw within this system change, and these vast stores of organic matter become more biologically mobile.
There is also some evidence that heterotrophic respiration will also increase, which may offset any net gains in ecosystem carbon uptake even in the face of increased gross primary productivity (Euskirchen et al., 2009). Although ecosystem acclimation may lead to net carbon sequestration over time (Oechel et al., 2000a), arctic tundra ecosystems may have multiple- and time variant responses to warming depending on their current state (dry vs. wet) (Oberbauer et al., 2007), or in the case of the northern coastal plain of Alaska, an area's succession state (Hinkel et al., 2003; Bockheim et al., 2004; Eisner et al., 2005; Zona et al., 2010).

Although considered homogeneous, the arctic tundra exhibits a large degree of spatial variability in structure and function over short distances (Walker et al., 1994; Shaver et al., 1996; Hinkel et al., 2001; Vourlitis et al., 2003; Engstrom et al., 2005; Riedel et al., 2005) as well as over much larger spatial scales (Oechel et al., 2000b; Vourlitis et al., 2003). The terrestrial landscape surface of the Barrow Peninsula of Alaska is a mosaic of small ponds, thaw lakes, different aged vegetated drained thaw-lake basins (VDTLBs), and interstitial tundra (Hussey & Michelson, 1966; Hinkel et al., 2003, 2005; Frohn et al., 2005) (Fig. 1). The Barrow Peninsula is essentially a mixed chronosequence of different aged land features in various states of succession brought about by a long-term process called the thaw-lake cycle (Britton, 1957; Billings & Peterson, 1980; Jorgenson & Shur, 2007). This process of formation, development, and drainage of these thaw-lakes results in the current landscape surface coverage of 72% thaw-lakes and VDTLBs, and the remaining 28% as interstitial tundra not affected by the thaw-lake cycle (Hinkel et al., 2003). The VDTLBs along the Barrow Peninsula are classified by their ages since having been drained, from young (<50 years), medium (50–300 years), old (300–2000 years), and ancient (2000–5500 years) (Hinkel et al., 2003) with some interstitial tundra areas being >5000 years old (Eisner et al., 2005). This chronosequence of VDTLBs and mosaic of interstitial tundra is important in the region’s carbon balance as the different aged VDTLBs have different organic layer thickness, accumulation rates (Chapin et al., 1980; Hinkel et al., 2003; Bockheim et al., 2004), and carbon exchange dynamics (Harazono et al., 2003; Zona et al., 2010) than the interstitial tundra areas (Chapin et al., 1980; Kwon et al., 2006).

Therefore, the spatial structure of fluxes across the Barrow Peninsula are no doubt heterogeneous, however, the focus of long-term ecosystem flux studies in the region has been on the upland, drier, interstitial tundra areas (e.g., Coyne & Kelley, 1975; Oechel et al., 2000a; Kwon et al., 2006). Although studies of carbon fluxes over Alaskan arctic ponds and lakes have been done (Coyne & Kelley, 1974; Kling et al., 1991, 1992; Eugster et al., 2003), relatively few have focused on thaw lakes and VDTLBs (see e.g., Phelps et al., 1998; Harazono et al., 2003; Zona et al., 2010), and even less on measurements that integrate the large-scale variability of the larger arctic landscape or region (see e.g. Brooks et al., 1996; Oechel et al., 1998a).

The lack of representative spatial measurements in the Arctic typically results in a generalization or oversimplification of the terrestrial land surface used in regional and global climate models (Clein et al., 2000; Sitch et al., 2007) as well as model parameterizations constrained by inputs from only a limited amount of research locations (Sitch et al., 2007). The spatial patterns and variability of the landscape is therefore lost when ground measurements are scaled to the region resulting in unaccounted for uncertainties in the regional scale estimates. Satellite remote sensing-based observations can be used to represent the spatial variability of the earth’s surface with spectral reflectance measurements, derived vegetation indices, surface temperature, etc., and used as drivers of models that describe ecosystem carbon dynamics. However, there remains a lack of detailed spatial measurements with sufficient sampling density of the surface fluxes, processes and drivers to fully represent the spatial heterogeneity of the arctic landscape. Empirical data to verify and constrain these satellite-based products, and model outputs that establish mechanistic linkages among scales of measurement are also often lacking (Desai et al., 2008), and observations of regional carbon exchange have been limited (Desai et al., 2008; Dolman et al., 2009; Riley et al., 2009). The large degree of functional variability across a heterogeneous landscape can therefore make the assimilation of plot level and tower measurements with regional and global estimates challenging (Desai et al., 2010). A key to understanding the regional patterns and controls on surface fluxes are methodologies that link and constrain these multiple scales of interest (Dolman et al., 2009).

The Barrow Peninsula provides an opportunity to compare the ecosystem fluxes across a heterogeneous landscape featuring a mixed chronosequence of thaw-lakes, VDTLBs, and interstitial tundra. Using a combination of portable tower- and aircraft-based eddy covariance (EC) measurements, we sampled fluxes of CO2, H2O, and sensible and latent heat fluxes across the coastal plain of Alaska incorporating thaw-lakes, different aged VDTLBs, and interstitial tundra. Using the spatial information provided by Hinkel et al. (2003) and satellite derived NDVI measurements from the Moderate Resolution Imaging Spectroradiometer (MODIS) on the Terra satellite, we scale the fluxes of CO2 to the Barrow Peninsula region.© 2011 Blackwell Publishing Ltd, Global Change Biology, 17, 2781–2802
The objectives of this research are to (i) determine the spatial pattern of CO2 fluxes across the arctic landscape incorporating different aged VDTLBs and interstitial tundra, (ii) determine the temporal pattern of CO2 and energy fluxes of two different aged VDTLBs and interstitial tundra, and (iii) assess the regional scale flux using a combination of tower-, aircraft-, and satellite-based measurements. We also address the question: are aircraft-based flux measurements suitable to accurately characterize the spatial fluxes along the heterogeneous landscape of the Arctic Coastal Plain?

Materials and methods

Study area

The study area is on the Arctic Coastal Plain of Alaska near the city of Barrow (Fig. 1) and measurements were made between July 27 and September 4, 2006. The maritime climate results in short, moist, and cool summers with the long-term average summer temperature (1976–2007; June–August) at 2.4 °C and with the average temperatures for June, July, and August at 0.6 °C, 3.3 °C, and 3.2 °C, respectively (source of data: National Oceanic and Atmospheric Administration Global Monitoring Division, ftp://ftp.cmdl.noaa.gov/met/hourlymet/brw/rtd/). Average precipitation for June–August is 63.4 mm (Kwon et al., 2006) with the majority of precipitation as rain or mist.

The surface features along the Barrow Peninsula, an area approximately 1800 km² (Fig. 1 and Table 3), are the result of a long term landscape process called the thaw-lake cycle (Britton, 1957; Billings & Peterson, 1980; Jorgenson & Shur, 2007). This cycle begins when surface depressions form due to melting ground ice forming low-centered polygons which then accumulate water and form small ponds. The ponds exacerbate melting and ground subsidence and eventually coalesce into larger ponds and small lakes. The lakes expand and incorporate other nearby lakes and ponds which is then lengthened by wave and wind erosion resulting in elliptical
Table 1  Aircraft flight transects and dates of measurement along the Barrow Peninsula of Alaska

<table>
<thead>
<tr>
<th>Transect ID</th>
<th>Endpoint Coordinates 1 (WGS84, decimal degrees)</th>
<th>Endpoint Coordinates 2 (WGS84, decimal degrees)</th>
<th>Length (km)</th>
<th>Orientation</th>
<th># of transects</th>
<th>Measurement dates (year 2006)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Short</td>
<td>71.248757°, –156.503089°</td>
<td>71.157431°, –156.412729°</td>
<td>6.7</td>
<td>N–S</td>
<td>110</td>
<td>8/10–8/12, 8/23–9/2</td>
</tr>
<tr>
<td>Long</td>
<td>71.265211°, –156.515821°</td>
<td>71.037703°, –156.292318°</td>
<td>16.8</td>
<td>N–S</td>
<td>42</td>
<td>8/10–8/12, 8/23–9/2</td>
</tr>
<tr>
<td>PER</td>
<td>71.098210°, –156.475549°</td>
<td>71.026754°, –156.352254°</td>
<td>5.5</td>
<td>NE–SW</td>
<td>35</td>
<td>8/10–8/12, 8/23–9/2</td>
</tr>
</tbody>
</table>

Table 2  Locations of the portable tower measurement sites

<table>
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<tr>
<th>Site name</th>
<th>Basin age* (years since drainage)</th>
<th>Measurement dates (year 2006)</th>
<th>Site coordinates (WGS84, decimal degrees)</th>
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</thead>
<tbody>
<tr>
<td>Site 1</td>
<td>Young</td>
<td>7/27–8/13</td>
<td>71.220861°, –156.480306°</td>
</tr>
<tr>
<td>Site 2</td>
<td>Medium</td>
<td>7/27–8/13, 8/22–8/28</td>
<td>71.179280°, –156.434550°</td>
</tr>
<tr>
<td>Site 3</td>
<td>&gt;5000</td>
<td>8/22–8/28</td>
<td>71.051670°, –156.393331°</td>
</tr>
</tbody>
</table>

Tower heights at all sites were 2.56 m AGL.

*Basin ages from Hinkel et al. (2003).

Table 3  Information on the surface type, VDTLB ages, area coverage, CO2 flux, and NDVI for the Barrow Peninsula

<table>
<thead>
<tr>
<th>Surface type</th>
<th>Age* (years since drainage)</th>
<th>Surface area (km2)</th>
<th>Surface area (%)</th>
<th>NDVI – MODIS</th>
<th>NDVI – aircraft</th>
<th>CO2 flux (mg CO2 m-2 s-1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Young</td>
<td>0–50</td>
<td>57</td>
<td>3.2</td>
<td>0.541 ± 0.102</td>
<td>0.518 ± 0.0642</td>
<td>−0.177 ± 0.0418</td>
</tr>
<tr>
<td>Medium</td>
<td>50–300</td>
<td>141</td>
<td>7.8</td>
<td>0.53 ± 0.0915</td>
<td>0.568 ± 0.0682</td>
<td>−0.152 ± 0.0261</td>
</tr>
<tr>
<td>Old</td>
<td>300–2000</td>
<td>443</td>
<td>24.6</td>
<td>0.512 ± 0.0864</td>
<td>0.495 ± 0.0453</td>
<td>−0.105 ± 0.0191</td>
</tr>
<tr>
<td>Ancient</td>
<td>3000–5500</td>
<td>97</td>
<td>5.4</td>
<td>0.509 ± 0.0838</td>
<td>0.493 ± 0.0421</td>
<td>−0.0966 ± 0.0426</td>
</tr>
<tr>
<td>Interstitial</td>
<td>&gt;5000</td>
<td>669</td>
<td>37.1</td>
<td>0.506 ± 0.113</td>
<td>0.54 ± 0.0553</td>
<td>−0.102 ± 0.0548</td>
</tr>
<tr>
<td>Lakes</td>
<td>–</td>
<td>395</td>
<td>21.9</td>
<td>–</td>
<td>–</td>
<td>−0.00954 ± 0.00497</td>
</tr>
</tbody>
</table>

Total area
1802

Domain boundaries are shown in Fig. 1. The CO2 fluxes and NDVI values are shown as mean value ± the standard deviation of the mean, while the area CO2 flux is reported for the total area of each respective surface type. The MODIS NDVI data set was acquired between July 20 and 28, 2006 (see Fig. 7).

*Basin ages from Hinkel et al. (2003).

Lakes oriented perpendicular to the prevailing wind (Carson & Hussey, 1962). The lakes eventually drain through erosion, stream piracy, or bank overflow (Mackay, 1988) and low-centered polygons and ponds may reform (Billings & Peterson, 1980; Jorgenson & Shur, 2007). This long-term process controls much of the landscape features of the Barrow Peninsula with ~22% thaw-lakes, ~50% VDTLBs, and ~28% interstitial tundra independent of the thaw-lake cycle (Hinkel et al., 2003).

The vegetation surrounding the Barrow area is dominated by wet sedge tundra characterized by mixed communities of Carex aquatilis, Eriophorum angustifolium, Eriophorum scheuchzeri, Dupontia fisheri, the sub-shrub, Salix rotundifolia (Webber, 1978) as well as an abundance of mosses and lichens. The area has low relief topography consisting of high- and low-centered polygonal tundra (Brown et al., 1980). The vegetation communities within the VDTLBs are in a continuous succession as the basin ages and its surface and water status conditions change. Recently drained basins start off barren and begin a predictable succession pattern with younger aged VDTLBs having D. fischeri, E. scheuchzeri, and Arctophila fulva in the wet areas while medium aged VDTLBs also have the former plant species and also include C. aquatilis and E. angustifolium. As the basin ages and polygonization occurs, the old basins have low centered polygons with wet centers having C. aquatilis, and some Sphagnum mosses, small ponds with A. fulva, and rims having some moss and lichen coverage. Within ancient basins are small ponds and lakes with C. aquatilis and polygon rims with Cassiope tetragona and grasses (Bliss & Peterson, 1992; Hinkel et al., 2003).

Flight transects and tower locations

Three aircraft flight line transects were established across the Barrow Peninsula and three locations selected for portable tower EC measurements between July 27 and September 4,
AIRCRAFT-DERIVED REGIONAL SCALE CO₂ FLUXES

2006. A detail of the flight transects and portable tower locations are in Tables 1 and 2, respectively, with a map of the flight transects and tower locations shown in Fig. 1. The Short transect followed the same flight path of the Long transect for the first 6.7 km and was flown with greater frequency to increase the number of aircraft passes over the portable tower locations at Site 1 and Site 2. The third aircraft transect and Site 3 was located in a remnant of the original landscape that was not reworked by thaw-lake processes and informally identified as the Peterson Erosional Remnant (PER) (Bockheim et al., 2004; Eisner et al., 2005). This area is one of the largest areas of interstitial tundra on the Barrow Peninsula.

Tower-based EC

Tower-based ecosystem flux measurements were done using the EC technique (Swindank, 1951; Baldocchi et al., 1988; Loescher et al., 2006) with two identically configured portable tower systems deployed at the three selected sites (see Fig. 1 and Table 2). The portable towers had fixed measurement heights of 2.56 m above ground level (AGL) at all sites. Three-dimensional winds and virtual temperature were measured with a sonic anemometer (CSAT3, Campbell Scientific Inc., Logan, UT, USA), while CO₂ and H₂O vapor measurements were made using a fast response open-path infrared gas analyzer (IRGA) (LI-7500, LI-COR Inc., Lincoln, NE, USA). Incoming and reflected photosynthetic photon flux density (PPFD) (LI-190SB, LI-COR Inc.), and net radiation, Rn (NR-LITE, Kipp & Zonen, Delft, the Netherlands) was measured within the upwind sampling area at 1.4 m above the ground. Ground heat flux, G, (HFT3, REBS Inc., Seattle, WA, USA) was measured at 2 cm below the ground surface, while soil temperatures were measured at surface, 5 and 10 cm below the ground (Type-T thermocouples, Omega Engineering, Stamford, CT, USA). Air temperature, T_air, and relative humidity, RH, (HMP45c, Vaisala, Helsinki, Finland) were also measured at 2.56 m in a radiation shield.

The turbulence and meteorological measurements were sampled and recorded at 20 Hz on a solid state datalogger (CR3000, Campbell Scientific Inc.) with a 2GB CompactFlash for raw data storage. The 20 Hz raw turbulence data were processed using EDIRe (version 1.5.0.9, University of Edinburgh; http://www.geos.ed.ac.uk/abs/research/micromet/EdiRe/) and included spike detection and removal algorithms (Højstrup, 1993; Vickers & Mahrt, 1997) before being processed as 30-minute block averages. A 2-d coordinate rotation was used to align the vertical wind velocities normal to mean streamline (McMillen, 1988; Kaimal & Finnigan, 1994). Corrections for time delays, sensor frequency response and sensor separation were done (Moore, 1986), as well as corrections for the effects of air density fluctuations of sensible (H) and latent heat (LE) fluxes according to Webb et al. (1980). Processed data were quality assessed and checked according to tools described in Foken & Wichura (1996) and Thomas & Foken (2002) with additional data being rejected using instrument data diagnostic flags, and periods of rain, snow, fog, or if wind direction passed through the tower structure. The cospectrum comparisons (Kaimal & Finnigan, 1994), the energy balance and the degree of closure between the sum of the half-hourly 

\[ H + LE \]

and the combination of \( R_n \) and \( G \) (\( R_n - G \)), were used to assess system performance and the quality of the portable tower EC measurements (McMillen, 1988; Aubinet et al., 2000). Data gaps of 30-min or less were linearly interpolated (Falge et al., 2001), while gaps >30-min were filled using the online gap-filling tool located at http://gaia.agraria.unitus.it/database/eddyproc/ which incorporates techniques described in Falge et al. (2001), and enhancements to the algorithms described in Reichstein et al. (2005) that consider the temporal autocorrelation of fluxes and their co-variation with meteorological variables. If we had missing values for \( R_n \), \( T_{air} \), and \( RH \), surrogate data were obtained from the adjacent portable tower or the permanent EC tower (http://www.fs.forsyth.nrcan.gc.ca/database/eddyproc/ which incorporates techniques described in Falge et al. (2001)) located approximately 6 km to the east southeast of Barrow and described in Turner et al. (2005).

Airborne measurements

All aircraft data were collected using the San Diego State University (SDSU) Sky Arrow 650TCN Environmental Research Aircraft (hereafter referred to as Sky Arrow). The aircraft was flown as close to the surface as possible and average flight characteristics for all the flights was a height of 6.3 m AGL with an average ground speed of 40.5 m s⁻¹. The Sky Arrow is equipped with a Mobile Flux Platform (MFP) (Crawford et al., 1990) that includes a nose mounted Best Aircraft Turbulence (BAT) pressure sphere (Crawford & Dobosy, 1992; Hacker & Crawford, 1999) for turbulence measurements, a fast response open-path infrared gas analyzer (IRGA) (LI-7500, LI-COR Inc.), and a GPS based vector attitude system (TANS Vector, Trimble Navigation Ltd., Sunnyvale, CA, USA). Turbulence variables, aircraft attitude, and CO₂ and H₂O vapor fluctuations were measured at 50 Hz, while aircraft position was measured at 10 Hz using a 12-channel L1 frequency GPS system (OEM3, NovAtel Inc., Calgary, AB, Canada) and differentially corrected in post-processing (Waypoint GrafNav, NovAtel) using two GPS reference base stations (OEM3, NovAtel Inc., and NetRS, Trimble Navigation Ltd.).

Simultaneous measurements of incoming and reflected PPFD were made with two quantum sensors (LI-190SB, LI-COR Inc.), and net radiation was measured with a Fritsch-type net radiometer (Q*7.1, REBS Inc.). The net radiometer was not actively aspirated since once in flight, the airflow around the sensor exceeds its requirements for aspiration. Surface temperature was measured with an infrared temperature sensor (4000.4GL, Everest Interscience, Tucson, AZ, USA) with a 4° field of view (FOV) lens and aircraft altitude above the ground was measured with a laser altimeter (LD90-3300HR, Riegl, Horn, Austria). Hyperspectral reflectance measurements from 304 to 1134 nm (255 bands, 3.2 nm bins) were made using a dual channel spectrometer (UniSpec-DC, PPSystems, Amesbury, MA, USA) with an upward facing cosine incident light receptor, and a 10° FOV downward looking lens connected to the detector with fiber optics. Detector integration time was 25 ms and three samples were internally averaged.
before storage to the data acquisition system that resulted in sampling frequencies of 5–7 Hz. The 10° FOV lens and average flight height of 6.3 m AGL resulted in a sampling diameter of 1.1 m, and with an average ground speed of 40.5 m s⁻¹, each measurement was therefore integrated over 3.0 m, resulting in an effective ‘pixel’ resolution of 1.1 m × 3.0 m. The spectral reflectances were interpolated into 1 nm bands and the normalized difference vegetation index (NDVI) (Tucker, 1979) was calculated using the red (620–670 nm) and near infrared (841–876 nm) wavelengths which correspond to the MODIS bands 1 (b₁) and 2 (b₂), respectively. The NDVI is recognized as an effective ecosystem-level indicator of plant health, primary productivity, and canopy ability to capture photosynthetically active radiation (Tucker, 1979; Sellers, 1985; Box et al., 1989; Woodward et al., 1991; Myneni et al., 1997; Vermote & Saleous, 2006) and widely acquired from satellite- and aircraft-based platforms.

**Aircraft data processing**

EC measurements from aircraft platforms are similar to those of stationary ground-based towers with the exception of the compression correction for temperature, the measurement of the wind vectors relative to the surface (Crawford & Dobosy, 1992; Leise & Masters, 1993; Eckman et al., 1999), corrections for sideward and upwash generated from the aircraft’s fuselage and wings, respectively, (Crawford et al., 1996a), and a spatial instead of a temporal average (Crawford et al., 1993a). Determination of the optimal spatial averaging length was done using an ogive function of the integrated cospectrum of the vertical velocity and the scalar of interest (Desjardins et al., 1989; Oncley et al., 1990). The spatial average for this study was determined to be 1 km using the ogive function over all flights and transects. Additional details and description of the flux computation procedures and processing of these aircraft data can be found in Crawford & Dobosy (1992), Crawford et al. (1993a), Eckman et al. (1999), Gioli et al. (2004), and in the references therein.

**Flux footprints**

Relating the relative flux contribution of the surface to the measurement location for both aircraft- and tower-based measurements requires a footprint model described with a function depending on the measurement height above the ground, wind vector, surface roughness, and atmospheric stability (Schmid, 2002; Kljun et al., 2004). Here, we used the crosswind integrated footprint model described in Kljun et al. (2004) for both the tower- and aircraft-based measurements. This model was derived from footprint estimates of a three-dimensional Lagrangian footprint model (LPDM-B; Kljun et al., 2002) which employed a stochastic backward particle trajectory for improved model accuracy and provides simplified parameterizations allowing for rapid calculations with input parameters derived from turbulence measurements normally made by EC systems. This footprint model requires the measurement height (zm), boundary layer height (h), friction velocity (u*), standard deviation of the vertical wind (σz), roughness length (z₀), and the Obukhov length (L) (Obukhov, 1946; Monin & Obukhov, 1954; Obukhov, 1971; Foken, 2006). The model is valid across a wide range of boundary-layer stabilities and measurement heights with the overall conditions of −200 ≤ zm/L ≤ 1, u* ≥ 0.2 m s⁻¹, and 1 m ≤ zm ≤ h (Kljun et al., 2004).

The parameters for the tower-based EC system were calculated for the averaging period of the flux calculation (30-min) while the aircraft parameters were calculated from the spatial average of each measurement segment (1 km) along the flight paths. The aircraft was flown as close to the ground, tower location, and tower measurement height as possible to facilitate similar source areas for comparable measurements. Since the aircraft incorporates a spatial average for flux calculations, the footprint is conceptually an integrated swath of the landscape upwind with approximate dimensions equal to the footprint length times the distance of each measurement segment (1 km). The footprint area of the aircraft is therefore several times larger than that of the tower. Comparisons between aircraft- and tower-based measurements can be difficult to interpret when the upwind footprint areas span different land surface types; however, here the upwind areas used for the comparisons are constrained only within the surface features of interest. Here, we use the micrometeorological convention with negative flux values representing uptake (e.g., a flux going into the ecosystem from the atmosphere).

**Instrument calibrations**

To ensure synchronization of the aircraft data acquisition and measurement instrumentation (Scott et al., 1990; Bögel & Baumann, 1991), a calibration of the aircraft’s wind vector system was conducted from 00:00 to 02:00 (HH:MM) Alaska Standard Time (AST) on July 9, 2006. The calibration flight was done from 1.5 and 10 km off the coast of Barrow, over the ice covered Beaufort Sea at elevations from 1250 to 1325 m above sea level (ASL). Flight conditions were ideal as it was clear with steady and consistent winds aloft. Calibration maneuvers included a constant altitude wind box, standard rate turns, varied pitch and yaw oscillations, sideslip, and acceleration/deceleration maneuvers (Bögel & Baumann, 1991; Leise & Masters, 1993).

The hyperspectral detector on the aircraft was calibrated before and after each flight to correct for variations in solar radiation using a pressed Teflon calibration panel (Spectralon SRT-99, Labsphere, Sutton, NH, USA). Calibration of the IRGA for both the portable towers and the aircraft were done once a week and followed AmeriFlux protocols (http://public.ornl.gov/ameriflux/sop.shtml). A CO₂-free ultra-zero air was used as the zero set point for both the CO₂ and water vapor, WMO-traceable primary gas standards were used to span the CO₂, and a portable dew point generator (LI-610, LI-COR Inc.) was used to span the water vapor.

**Tundra areas and satellite-based NDVI**

A geographical information system (GIS; ArcGIS 9.3, ESRI, Redlands, CA, USA) was used to map the Barrow Peninsula study area, and the spatial coverage of the VDTLBS and interstitial tundra areas (Table 3). The GIS data layers were obtained (K. Hinkel, personal communication) and based on
the data used in Hinkel et al. (2003). Intersect analysis using ARCGIS software was used to correlate the fluxes, surface types and ages, and NDVI parameters.

Determination of the regional NDVI for the Barrow Peninsula was done using the MODIS Terra Surface Reflectance 8-Day Composite L3 Global SIN Grid V005 250 m resolution data product (MOD09Q1, http://modis-sr.ltdri.org/) managed by the MODIS Land Surface Reflectance Science Computing Facility (LSR SCF) (Vermote et al., 1997, 2002) and distributed by the Land Processes Distributed Active Archive Center (LP DAAC), located at the U.S. Geological Survey (USGS) Earth Resources Observation and Science (EROS) Center (http://lpdaac.usgs.gov). The NASA Warehouse Inventory Search Tool (WIST; https://wist.echo.nasa.gov) was used for final data selection and download.

The MODIS 8-day composite MOD09 products utilize the most ‘cloud free’ observations for single ca. 250 m pixels during an 8-day acquisition period. The surface reflectance is corrected for high cirrus clouds, aerosol interference, and other atmospheric parameters to give a ground-level value in the red (620–670 nm) and near-infrared (841–876 nm) spectral regions that are the MODIS bands b1 and b2 (Vermote et al., 1997, 2002; Vermote & Saleous, 2006; Vermote & Kotchenova, 2008). The swath data was obtained in a sinusoidal projection and converted to a UTM grid using the MODIS Reprojection Tool – Swath software, MRTSWATH (v2.1; https://lpdaac.usgs.gov/lpdaac/tools/modis_reprojection_tool_swath) and following re-projection, each pixel was 232 m × 232 m. There were three 8-day composite MODIS Terra data sets that were appropriate for the time frame of the measurement campaign, day of year (DOY) 193, 201 and 209, however, due to significant cloud coverage in two of the data sets only one provided a clear view of the entire Barrow Peninsula with a majority of cloud-free pixels. This DOY 201 data set was acquired between July 20 and 28, 2006.

The raw digital values in b1 and b2 were used to calculate the NDVI using MultSpec (v3.25.10; http://cobweb.ecn.purdue. edu/~biehl/MultSpec/) (Biehl & Landgrebe, 2002) scaled appropriately so all values are within the 16-bit signed integer image space. The calculation was NDVI = (b2–b1)/(b2 + b1). Comparisons between aircraft- and satellite-based NDVI were done by averaging the high-resolution aircraft measurements for each corresponding satellite pixel along the flight transects.

### Statistical analysis

Aircraft-based CO₂ flux and NDVI measurements along the flight line transect were analyzed with the age classifications for the VDTLBs and interstitial tundra. Comparison between groups was made using a one-way ANOVA with normality assessed using a Shapiro–Wilk test, and the Holm–Sidak method for multiple pair-wise comparisons where \( P < 0.05 \) was considered significant.

Least-squares linear regression was used to compare the energy balance closure of the portable EC towers half-hourly sum of \( H + LE \), and the combination of \( R_w - G \), as well as comparisons between the aircraft- and portable tower-based EC measurements, and the comparison between the aircraft- and MODIS-based NDVI measurements. All statistical analyses were performed using SIGMAPLOT (11, Systat Software Inc., San Jose, CA, USA).

### Results

#### Tower and aircraft performance

The flux footprint parameters and results of the footprint analysis for the aircraft and portable towers are presented in Table 4. Each portable tower had similar parameters to input in to the footprint analysis, exhibiting similar surface microtopography and surface roughness \( (z_o) \). Overall, the footprint estimates \( (x_{90\%}) \) over the three portable tower sites averaged no larger than 135 m with the footprint peak contribution \( (x_{\text{max}}) \) between 37 and 49 m upwind of the towers. The aircraft measurement height was typically 2.5 times higher and the footprint estimates \( (x_{90\%}) \) and \( (x_{\text{max}}) \) was approximately 2.8 times larger than those of the tower. As expected, the standard deviations of the footprint estimates for the aircraft are much larger when compared with those of the towers, however, the aircraft had a spatial average over an integrated area much larger than the tower footprint areas.

The \( R_w \) ranged between \(-59.2 \) and \( 378.1 \) \( \text{W m}^{-2} \), \( H \) from \(-21.2 \) to \( 234.5 \) \( \text{W m}^{-2} \), \( LE \) from \(-9.9 \) to

### Table 4 Results of the footprint analysis for the aircraft- and portable tower-based EC measurements

<table>
<thead>
<tr>
<th>Location</th>
<th>( x ) (m)</th>
<th>( u^* ) (m s(^{-1}))</th>
<th>( w_s ) (m s(^{-1}))</th>
<th>( \sigma_w ) (m s(^{-1}))</th>
<th>( z_o ) (m)</th>
<th>( x_{\text{max}} ) (m)</th>
<th>( x_{90%} ) (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site 1</td>
<td>2.56</td>
<td>0.33 ± 0.09</td>
<td>4.04 ± 1.15</td>
<td>0.37 ± 0.10</td>
<td>0.04</td>
<td>44.6 ± 3.0</td>
<td>122.1 ± 8.3</td>
</tr>
<tr>
<td>Site 2</td>
<td>2.56</td>
<td>0.33 ± 0.08</td>
<td>3.71 ± 0.98</td>
<td>0.38 ± 0.10</td>
<td>0.03</td>
<td>41.4 ± 3.3</td>
<td>113.3 ± 9.0</td>
</tr>
<tr>
<td>Site 3</td>
<td>2.56</td>
<td>0.32 ± 0.06</td>
<td>4.14 ± 0.65</td>
<td>0.34 ± 0.05</td>
<td>0.02</td>
<td>48.0 ± 1.2</td>
<td>131.5 ± 3.2</td>
</tr>
<tr>
<td>Long</td>
<td>6.38 ± 0.91</td>
<td>0.31 ± 0.06</td>
<td>5.01 ± 3.39</td>
<td>0.31 ± 0.06</td>
<td>0.04</td>
<td>130.0 ± 47.0</td>
<td>343.4 ± 63.2</td>
</tr>
<tr>
<td>Short</td>
<td>6.26 ± 0.78</td>
<td>0.36 ± 0.08</td>
<td>5.22 ± 1.25</td>
<td>0.36 ± 0.07</td>
<td>0.02</td>
<td>119.8 ± 18.8</td>
<td>328.2 ± 51.5</td>
</tr>
<tr>
<td>PER</td>
<td>6.25 ± 0.46</td>
<td>0.32 ± 0.06</td>
<td>6.26 ± 0.66</td>
<td>0.32 ± 0.05</td>
<td>0.02</td>
<td>128.8 ± 17.8</td>
<td>352.8 ± 48.7</td>
</tr>
</tbody>
</table>

Results are shown as means ± the standard deviations of the mean.\( x \) is the measurement height, \( u^* \) is the friction velocity, \( w_s \) is the wind speed, \( \sigma_w \) is the standard deviation of the vertical wind (w), \( z_o \) is the surface roughness, \( x_{\text{max}} \) is the distance of peak contribution, and \( x_{90\%} \) is the distance of 90% of the footprint contribution (Kljun et al., 2004).

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168.8 W m$^{-2}$, and $G$ from −9.7 to 39.5 W m$^{-2}$. The energy balance closure between the sum of the half-hourly $H + \lambda E$, and $R_n-G$ (McMillen, 1988; Aubinet et al., 2000) for both portable EC towers is shown in Fig. 2. A least-square linear regression was fitted to the portable tower data resulting in an $r^2$ of 0.91 and 0.92, and slopes of 0.87 and 0.83 for portable Towers A and B, respectively, and when the data from the two towers were combined resulted in an $r^2$ of 0.91 and a slope of 0.85. We interpret these results as indicating the portable EC towers had good system performance.

There was also good agreement between the aircraft- and tower-based CO$_2$ flux measurements with the comparison being nearly 1:1 (Fig. 3), while the aircraft slightly overestimated the ground-based measurements of $R_n$ and slightly underestimated the ground-based surface measurements of $H$ and $\lambda E$.

Temporal and Spatial Patterns of CO$_2$ Flux

The heterogeneous nature of the landscape on the Barrow Peninsula is expressed in the mixed chronosequence of VDTLBs, thaw lakes, small ponds, and interstitial tundra areas (Fig. 1). The spatial pattern of the CO$_2$ fluxes across the Long transect for the period between August 10 and 11, 2010 (Fig. 4) incorporates various VDTLB ages, small ponds, and interstitial tundra areas into its footprint.

The young VDTLB at Site 1 is relatively small compared with the medium aged VDTLB at Site 2, however, it had the largest net uptake of $-0.24$ mg CO$_2$ m$^{-2}$ s$^{-1}$ with an average uptake of $-0.16$ mg CO$_2$ m$^{-2}$ s$^{-1}$ when the fluxes are spatially integrated across the basin. The medium aged VDTLB at Site 2 had larger uptake along the northern portion than the southern portion of the basin at $-0.20$ and $-0.12$ mg CO$_2$ m$^{-2}$ s$^{-1}$, respectively, with an overall average uptake of $-0.15$ mg CO$_2$ m$^{-2}$ s$^{-1}$ when the fluxes were spatially integrated across the basin. The interstitial tundra areas are interspersed throughout this landscape and even though fluxes from small patches (<100 m) cannot be discerned by the aircraft measurements, the range of the CO$_2$ fluxes were between $-0.11$ and $-0.19$ mg CO$_2$ m$^{-2}$ s$^{-1}$, which also corresponded to the PER interstitial tundra region with fluxes ranging from $-0.13$ to $-0.17$ mg CO$_2$ m$^{-2}$ s$^{-1}$ (Fig. 5). One Ancient VDTLB was measured along the Long transect and showed a net CO$_2$ uptake from $-0.11$ and $-0.13$ mg CO$_2$ m$^{-2}$ s$^{-1}$ (Fig. 4) and was similar to another Ancient VDTLB along the PER transect with uptake from $-0.10$ and $-0.12$ mg CO$_2$ m$^{-2}$ s$^{-1}$ (Fig. 5). The Old VDTLBs had the lowest CO$_2$ flux and ranged from $-0.06$ and $-0.14$ mg CO$_2$ m$^{-2}$ s$^{-1}$. The northern extent of the PER transect was Lake Sungovoak (see Fig. 1) and fluxes approached zero as the aircraft entered lake area (Fig. 5, see Table 3). Larger and deeper thaw lakes are expected to either have near neutral CO$_2$ flux or be a slight source to the atmosphere (Kling et al., 1991, 1992), and since only a small and shallow section of the lake was measured, it is difficult to quantify with certainty whether this type of lake was CO$_2$ neutral or a net source to the atmosphere.

The diurnal patterns of CO$_2$ flux across the three measurement sites as measured by the portable towers and aircraft are shown in Fig. 6. The diurnal patterns between July 27 and August 13, showed Sites 1 and 2 were a source of approximately $0.4$ mg CO$_2$ m$^{-2}$ s$^{-1}$ between 2200 and 0600 with peak uptake occurring between 1100 and 1330 that were $-0.25$ and $-0.31$ mg CO$_2$ m$^{-2}$ s$^{-1}$ for Sites 1 and 2, respectively. Although Site 2 had larger peak uptake, Site 1 showed larger uptake rates when integrated over the full diurnal timeframe. Later in the season between August 23 and August 28, the magnitude of the fluxes decreased and the diurnal pattern was less pronounced as well as a shift in the daily maximum flux rate from ~1100 to ~1300 (Fig. 6). Comparing the EC data against other towers in the Barrow area showed similar diurnal patterns, however, with smaller magnitudes in the diurnal patterns of CO$_2$ flux (data not shown).

Spatial NDVI

The spatial pattern of NDVI from the MODIS, DOY 201, 8-day composite is shown in Fig. 7, and the spatial areas of each identified NDVI range class in Table 5. It should be noted that the total spatial area is $32$ km$^2$ larger than the spatial area reported in Table 3. This is due to the MODIS pixels along the domain borders (see Fig. 1).
extending past the borders of the Hinkel et al. (2003) dataset, and the entire MODIS pixel area was included in the area calculations. An analysis of the reflectance data from the aircraft spectrometer, satellite imagery, and satellite NDVI showed that pixels with NDVI < 0.37 had a high degree of standing water and contaminated the NDVI vegetation signal that resulted in significantly decreased NDVI values. The NDVI range for the vegetated areas was between 0.37 and 0.70, while large water bodies, thaw lakes, and ocean were identified by NDVI < 0.01.

The spatial pattern of NDVI from the aircraft measurements and the associated MODIS, DOY 201, 8-day composite pixels along the Long and PER transects are shown in Fig. 8. The range of NDVI along the aircraft transects for both the aircraft and MODIS NDVI was between 0.37 and 0.70 being indicative of vegetated surfaces. The standard deviations of the aircraft NDVI measurements within this range were larger than the MODIS measurements with standard deviations of 0.061 and 0.049, respectively. The NDVI values from 0.1 to 0.37 tended to be a mixture of vegetation, standing water or bare soil with NDVI values in the lower part of this range having a larger percentage of standing water and/or bare soil, and NDVI values between 0.1 and –1 are standing open water. The areas predominated with vegetated cover and with little to no standing water (NDVI > 0.37) were used for further analysis and validation of the MODIS NDVI.

Comparisons between the aircraft- and satellite-based NDVI measurements (NDVI > 0.37) along the Long and

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Fig. 3 Intercomparisons between the aircraft- and tower-based flux measurements of CO$_2$, $R_n$, $H$, and, $\lambda E$. Gray bars are standard deviations of the mean and the diagonal line indicates unity.
PER flight transects show a linear relationship with the slope of the regression of 0.6 and an $r^2 = 0.45$ (Fig. 9). This indicates that the aircraft provides a suitable validation of the MODIS NDVI measurements of the vegetated areas and provide a representative measure of the surface NDVI. The spatial NDVI estimated here can therefore be used with confidence to scale the aircraft fluxes to the Barrow Peninsula region.

Fig. 4 Average aircraft-based CO$_2$ flux measurements along the Long transect between July 27 and August 13, 2006. Color coding defines the ages since drainage of the different aged VDTLBs (see Fig. 1) as well as the interstitial tundra areas. Squares indicate locations where the aircraft's flux footprint was completely within the indicated surface type.

Fig. 5 Average aircraft-based CO$_2$ flux measurements along the PER transect between July 27 and August 13, 2006. Color coding defines the ages since drainage of the different aged VDTLBs (see Fig. 1) as well as the interstitial tundra areas. Squares indicate locations where the aircraft's flux footprint was completely within the indicated surface type. Gray bars are standard deviations of the mean.
Comparisons of surface type with CO₂ flux and NDVI

We were able to detect a high degree of spatial variability among the various types of ecosystems observed by the aircraft. Comparisons of the aircraft-derived CO₂ fluxes from these various aged VDTLBs, thaw lakes, and interstitial tundra were also found to be significantly different, with the exception of those comparisons with Ancient and Old VDTLBs, and interstitial tundra regions, and the Medium and Young VDTLBs (Table 6). The CO₂ fluxes from the Medium and Young aged VDTLBs were also not significantly different.

Fig. 6 The diurnal patterns of CO₂ flux across the three measurement sites (see Fig. 1) as measured by the portable EC towers and aircraft between July 27 and August 13, 2006 (left side), and between August 23 and August 28, 2006 (right side). Gray bars indicate standard deviations from the mean.
Comparisons of daily net ecosystem exchange by Zona et al. (2010) also showed no significant differences between Medium and Young aged VDTLBs, however they did show a significant difference between the Old and Medium and Young aged VDTLBs, which is in contrast to the results reported here.

The aircraft-based NDVI measurements also showed there were significant differences among the land surface types with the exception of the Ancient and Old VDTLBs (Table 6). The satellite-based NDVI measurement showed similar results as the aircraft-based measurements, and similarly did not show significantly different NDVI between the Ancient VDTLB and the interstitial tundra. The highest NDVI from the aircraft-based NDVI measurements was in the Medium aged VDTLBs and showed no pattern of NDVI with VDTLB age, however, the satellite-based NDVI showed the Young aged VDTLBs as having the highest values and decreased with VDTLB age and showed the interstitial tundra having the lowest NDVI (Table 3).

Regionally scaled estimates

By partitioning the associated aircraft fluxes into their respective source areas (land surface types) from the Long and PER transects, weighting the fluxes based on the fractions of the landscape from the basin coverage (Table 3), and determining the functional relationship between NDVI and CO2 flux, a simple regional flux was derived. The aircraft CO2 fluxes along the Long and PER transects and the MODIS, DOY 201, 8-day composite (NDVI > 0.37) were used to determine the functional relationship (Fig. 10). A quadratic function best described the relationship between CO2 flux and MODIS NDVI with an $r^2 = 0.50$ (Fig. 10). Since NDVI < 0.37 were shown to be standing water or large lakes and assumed to be a source of CO2 to the atmosphere, we used an average CO2 flux value reported in the literature for arctic lakes, and estimate a CO2 flux of 0.0102 mg CO2 m$^{-2}$ s$^{-1}$ (Coyne & Kelley, 1974; Kling et al., 1991, 1999).

Table 5  NDVI range classes and spatial coverage from the MODIS MOD09Q1, DOY 201, 8-day composite for the domain shown in Fig. 1 and Fig. 7

<table>
<thead>
<tr>
<th>NDVI</th>
<th># Pixels</th>
<th>% of total image</th>
<th>Surface area (km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;0.1</td>
<td>6,793</td>
<td>19.8</td>
<td>365.6</td>
</tr>
<tr>
<td>0.1-0.2</td>
<td>503</td>
<td>1.5</td>
<td>27.1</td>
</tr>
<tr>
<td>0.2-0.3</td>
<td>767</td>
<td>2.2</td>
<td>41.3</td>
</tr>
<tr>
<td>0.3-0.37</td>
<td>750</td>
<td>2.2</td>
<td>40.4</td>
</tr>
<tr>
<td>0.37-0.4</td>
<td>509</td>
<td>1.5</td>
<td>27.4</td>
</tr>
<tr>
<td>0.4-0.5</td>
<td>4,281</td>
<td>12.5</td>
<td>230.4</td>
</tr>
<tr>
<td>0.5-0.6</td>
<td>19,412</td>
<td>56.6</td>
<td>1044.8</td>
</tr>
<tr>
<td>0.6-0.7</td>
<td>1250</td>
<td>3.6</td>
<td>67.3</td>
</tr>
<tr>
<td>&gt;0.7</td>
<td>10</td>
<td>0.0</td>
<td>0.5</td>
</tr>
<tr>
<td>Total pixels</td>
<td>34,275</td>
<td>Total area</td>
<td>1844.8</td>
</tr>
</tbody>
</table>
The CO₂ flux for each MODIS NDVI pixel was calculated, then binned according to land surface type, multiplied by the corresponding surface, and summed to estimate a regional flux. Because the ANOVA showed that the CO₂ fluxes were not significant among the interstitial tundra, Ancient, and Old VDTLBs (see Table 6), we pooled these estimates together. The CO₂ fluxes of the Medium and Young VDTLBs were also not significantly different and those data were pooled together as well. The scaled estimates for the region are shown in Table 7. The area average midday hourly CO₂ flux for the 1802 km² study region (see Fig. 1) was $-2.04 \times 10^5$ kg CO₂ h⁻¹ for representative midsummer conditions.
The combined interstitial tundra, Ancient and Old VDTLBs accounted for nearly 59% of the total area flux and were representative of 67% of the surface area. The Young and Medium VDTLBs represented 11% of the total surface area but had higher flux rates resulting in 35% of the total area flux. The remaining 22% of the surface area were represented by the large thaw lakes and standing water areas which contributed to 6% of the total flux. Using the combined average CO₂ flux of the interstitial tundra, and Ancient and Old (ITAO) VDTLBs, $-0.101 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$, and subtracting this estimate from the aircraft measured CO₂ flux along the Long transect demonstrates the relative spatial heterogeneity of the surface fluxes along the Barrow Peninsula (Fig. 11).

**Discussion**

**Tower and aircraft performance**

The energy balance closure for the portable EC systems was similar to that found in wet-sedge tundra (Kwon et al., 2006) and VDTLBs within the Barrow peninsula (Zona et al., 2010), as well as other wet-sedge tundra ecosystems along the Alaskan Arctic Coastal Plain (Vourlitis & Oechel, 1997). Compared with other EC systems within FLUXNET (Baldocchi et al., 2001), the energy balance closure found here from the portable EC systems are considered very good (Wilson et al., 2002).

Energy balance is commonly unclosed for EC systems (Mahrt, 1998; Twine et al., 2000) and the imbalance of energy closure here is suspected to be due to uncertainties associated with (1) the difficulty of measuring small flux densities of $\mathcal{H}$ and $\mathcal{L}E$ during near-neutral early morning and stable night-time conditions, (2) mismatches in the spatial characterization between the flux footprint and the localized spatial area of the $R_n$ and $G$ measurements, (3) the differences in temperature and water vapor due to microtopography resulting in horizontal advection, (4) configuration of the sonic anemometer (J. Frank, personal communication), and/or (5) unmeasured storage of energy particularly in inundated and water saturated (H. Loescher, personal communication).

Intercomparisons between tower and aircraft EC measurements along the Arctic Coastal Plain of Alaska have been made previously with good success (Brooks et al., 1996; Crawford et al., 1996b; Oechel et al., 1998a). The underestimation of $\mathcal{H}$ and $\mathcal{L}E$ are consistent with

![Fig. 9 Comparisons between the aircraft- and satellite-based NDVI measurements (NDVI > 0.37) along the Long and PER flight transects. Gray bars indicate standard deviations from the mean and the diagonal line indicates unity.](image)

<table>
<thead>
<tr>
<th>Table 6</th>
<th>Pairwise comparisons following a one-way ANOVA of CO₂ flux, NDVI (aircraft- and MODIS-based), and the different aged VDTLBs, interstitial tundra areas, and thaw lakes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pairwise comparison</td>
<td>CO₂ flux</td>
</tr>
<tr>
<td>Ancient vs. Young</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Ancient vs. Medium</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Ancient vs. Old</td>
<td>NS</td>
</tr>
<tr>
<td>Ancient vs. Interstitial</td>
<td>NS</td>
</tr>
<tr>
<td>Interstitial vs. Young</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Interstitial vs. Medium</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Interstitial vs. Old</td>
<td>NS</td>
</tr>
<tr>
<td>Old vs. Medium</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Old vs. Young</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Medium vs. Young</td>
<td>NS</td>
</tr>
<tr>
<td>Lake vs. Ancient</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Lake vs. Old</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Lake vs. Medium</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Lake vs. Young</td>
<td>$P&lt;0.001$</td>
</tr>
<tr>
<td>Lake vs. Interstitial</td>
<td>$P&lt;0.001$</td>
</tr>
</tbody>
</table>

Critical value for the ANOVA was 0.05 and NS = not significant.
results from previous comparisons between aircraft and towers in arctic tundra (Brooks et al., 1996; Crawford et al., 1996b; Oechel et al., 1998a) as well as other ecosystems ranging from grasslands (Desjardins et al., 1992; Kelly et al., 1992; Gioli et al., 2004), forested ecosystems (Crawford et al., 1996b; Gioli et al., 2004), agricultural fields (Desjardins et al., 1989; Gioli et al., 2004; Isaac et al., 2004), to subtropical coastal environments (Crawford et al., 1993b). These previous studies reported that $H$ measurements from aircraft underestimated the ground-based tower fluxes from 10% to 50%, while aircraft measurements of $\lambda E$ did not follow a consistent pattern, neither did the under- or overestimation of tower-based observations. Sources of error between aircraft and towers are discussed in Mann & Lenschow (1994) and Mahrt (1998), and though these previous studies used different suites of instrumentation, software packages and were subjected to different sources of potential bias, the most commonly discussed sources of error were the (i) loss of low frequency flux contributions due to short flight transects, (ii) high frequency flux loss due to undersampling, (iii) footprint mismatches, and (iv) vertical flux divergence with height (Desjardins et al., 1989, 1992; Kelly et al., 1992; Crawford et al., 1993b, 1996b; Oechel et al., 1998a; Gioli et al., 2004; Isaac et al., 2004). If we discount the systematic sources (issues i–iii above), argues that a universal scaling law may apply (Brutsaert, 1982) linking surface flux measurements to lower boundary layer estimates. But larger datasets spanning a range of source/sink strengths and atmospheric conditions would be needed to test this notion.

In this study, we minimized these sources or error by flying the Sky Arrow as close to the surface as possible, and having multiple repeated flights over the same spatial area, though footprint mismatches will persist due to the spatial integration of the aircraft measurements. Although the footprint of the aircraft is larger than that of the tower, comparisons of the fluxes were similar since it was possible to have entire aircraft footprint areas within the land surface type being measured by the portable EC systems. However, there is also a mismatch between the sampling area of the fluxes and the downward looking remotely sensed measurements (i.e., NDVI measurements). The land surface type underneath the aircraft at the time of the measurement was not necessarily the same as the upwind footprint area. This was particularly true when the aircraft was at the margins of a land surface type (e.g., edges of VDTLBs) or when transitioning between surface types. The different sampling areas between the remotely sensed measurements and the fluxes partly account for the uncounted variability in the NDVI and CO$_2$ flux comparisons (see Fig. 10).

Another footprint mismatch was between the aircraft- and satellite-based NDVI measurements. The high spatial sampling frequency and resolution from the aircraft instrumentation provided greater detail in the spatial patterns of NDVI and demonstrated the high spatial variability of tundra types that was not apparent in the satellite-based measurements. The higher dynamic range of NDVI values from the aircraft measurements was in part due to the increased resolution in spatial sampling and increased spectrum reflectance due to

![Fig. 10](functionalRelationship.png)

**Fig. 10** Functional relationship between the aircraft-based CO$_2$ fluxes along the Long and PER transects and the MODIS, DOY 201, 8-day composite (NDVI > 0.37). The diagonal line is the quadratic function that best describes the relationship and the bars indicate the standard deviations from the mean.

<table>
<thead>
<tr>
<th>Surface type</th>
<th>Surface area (km$^2$)</th>
<th>Surface area (%)</th>
<th>Area CO$_2$ flux (kg CO$_2$ h$^{-1}$)</th>
<th>Area CO$_2$ flux (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Young + Medium</td>
<td>198</td>
<td>11.0</td>
<td>$-8.17 \times 10^4$</td>
<td>35</td>
</tr>
<tr>
<td>Interstitial + Ancient + Old</td>
<td>1209</td>
<td>67.1</td>
<td>$-1.37 \times 10^5$</td>
<td>59</td>
</tr>
<tr>
<td>Lakes</td>
<td>395</td>
<td>21.9</td>
<td>$1.45 \times 10^4$</td>
<td>6</td>
</tr>
<tr>
<td>Total area</td>
<td>1802 km$^2$</td>
<td></td>
<td>Total area hourly CO$_2$ flux</td>
<td>$-2.04 \times 10^5$ kg CO$_2$ h$^{-1}$</td>
</tr>
</tbody>
</table>

The area CO$_2$ flux (kg CO$_2$ h$^{-1}$) is calculated for each respective surface area, while the area CO$_2$ flux (%) is based on the total absolute magnitude of the area CO$_2$ flux and the net total area hourly CO$_2$ flux.
reduced atmospheric attenuation between the surface and the instrument. Case in point, the small surface features, including small vegetation patches, ponds, thaw lakes, and standing water areas were not visible in the satellite pixels.

The standing water greatly decreased the NDVI signal due to poor reflectance of both the red and near-infrared spectrum that resulted in very low and negative NDVI values. Small ponds and thaw lakes were easily distinguished along the flight transects with the high spatial sampling resolution of the aircraft, however, the lower resolution MODIS pixel was only able to differentiate the larger water bodies. Depending on the alignment of the MODIS pixel and the inclusion of large

**Fig. 11** Detailed image of the *Long* flight transect showing the visible surface features and the various aged VDTLBs and interstitial tundra areas. The color coding defines the ages since drainage of the different aged VDTLBs (see Fig. 1) as well as the interstitial tundra areas. The black line indicates the combined average CO$_2$ flux of the interstitial tundra, and Ancient and Old (ITAO) VDTLBs, $-0.101$ mg CO$_2$ m$^{-2}$ s$^{-1}$, and is displayed along the *Long* transect flight path. The white bars are the deviations from the ITAO mean with bars to the left of the line indicating a larger uptake than the ITAO mean and bars to the right of the line indicating a smaller uptake than the ITAO mean. The black arrow shows the mean prevailing wind direction and therefore the upwind area contributing to the CO$_2$ flux.
water bodies within each pixel, there appeared to be a misalignment in the areas identified as water in the MODIS NDVI when compared with the aircraft measurements (see Fig. 8). Owing to the disparity between the pixel resolutions, these offsets between the two measurements were therefore expected. The physical differences between how the aircraft- and satellite-based NDVI were derived was mentioned previously, but the differences observed here are interpreted as due to the higher sampling and spatial resolution of the aircraft compared with the satellite-based pixel.

Scaling methodology

The scaling of surface measurements to larger landscapes or the region is problematic due to inaccuracies of quantifying the spatial heterogeneity of the surface fluxes. Yet, we have high confidence in our ability to calculate and delineate footprints from specific cover types. This was due to very little differences in canopy height and roughness elements among the VDTLBs, compared with other non-tundra ecosystems (e.g., forests, shrubs, and desert). Larger structural discontinuities would cause large uncertainties in resolving source areas from airborne measurements. Hence, the differences found here among cover types, we attribute in our ability to resolve source areas and estimate their associated functional response. This ability may not be as defined over more complex landscapes containing different ecosystem types with large structural differences (i.e., edge effects, clearings in forests, urban environments), and should be explored further.

Ecosystem and atmospheric models typically treat arctic tundra as a single surface type and utilize limited surface measurements for model parameterization. For example there are currently two long-term EC towers within the northern Arctic Coastal Plain of Alaska (Turner et al., 2005; Kwon et al., 2006) which are both located on the Barrow Peninsula. They have been used to represent wet sedge tundra ecosystems in regional scaling exercises even though these towers are within interstitial tundra areas that represent approximately ~37% of the Barrow Peninsula surface area. Because there is no significant difference in the CO2 fluxes between the interstitial, and the Old and Ancient VDTLBs, these towers represent approximately ~67% of the Barrow Peninsula surface area, accounting for ~59% of the regional flux signal (see Table 7). Utilizing just these tower data and extrapolating them to the vegetated areas of the whole Barrow Peninsula would underestimate the CO2 flux by ~22% since the higher flux rates of the Young and Medium aged VDTLBs were not accounted for. Although the Young and Medium VDTLBs represent 11% of the surface area of the Barrow Peninsula, they account a large portion, ~35%, of the total regional flux and should be considered in regional estimates.

Even though this study measured the thaw lake (Lake Sungovoak) as having a small uptake rate of ~0.009 mg CO2 m⁻² s⁻¹, we discount this estimate as having contamination from other nearby ecosystems because the aircraft was only able to sample a few hundred meters into the lake footprint (see Fig. 5). Although the flux rate was close to zero, the uptake rate was several times larger than that reported by Kling et al. (1991, 1992) (~0.0004 and ~0.003 mg CO2 m⁻² s⁻¹) for two lakes as having a small CO2 uptake. These were the only two lakes out of 25 lakes studied that demonstrated uptake, all the remaining arctic Alaska lakes were all CO2 sources (0.0006–0.03046 mg CO2 m⁻² s⁻¹ with the overall average CO2 flux of 0.01065 mg CO2 m⁻² s⁻¹ (Kling et al., 1991, 1992). In other studies, a lake and pond measured in the Barrow area was reported to be a CO2 source, ranging from 0.007 to 0.023 mg CO2 m⁻² s⁻¹ (Coyne & Kelley, 1974), over Toolik Lake, AK a source of 0.005 mg CO2 m⁻² s⁻¹ (Eugster et al., 2003), while a boreal lake showed a source from 0.009 to 0.018 mg CO2 m⁻² s⁻¹ (Vesala et al., 2006). The regional scaling reported here used the average CO2 flux rate reported in the literature for the arctic lakes and assumed to be an accurate and conservative estimate than by assuming these large lakes have neutral CO2 flux.

Annual ice-free periods are short, ~60–90 days. But because a larger fraction of fresh water arctic bodies have shown to be a source of CO2 while a smaller fraction as a sink, argues that processes that control the source/sink status are not clear, e.g., limited surface renewal (Kratz et al., 2005; MacIntyre et al., 2006), depth and surface heating, plant and microbe community, age of sediment, etc. Although the CO2 fluxes from lakes and ponds are relatively small, they still can contribute significantly to the regional flux due to their large total area coverage over arctic Alaska (Kling et al., 1991, 1992).

The scaled regional estimates for the Barrow Peninsula are calculated from the aircraft-based flux measurements and should be considered a midday peak average CO2 flux. Although using a full diurnal measurement would yield a daily integrated CO2 flux value, these data are still valuable and validate modeled and scaled estimates.

Landscape surface

There were systematic differences in both NDVI and surface CO2 flux that correlated with the various land-
landscape features. The presence of continuous permafrost underlying the region (Brown et al., 1980) and the annual freezing and thawing of the upper soil layers result in the patterned ground of ice-wedge polygons, thaw lakes, and VDTLBs (Hussey & Michelson, 1966; Tieszen, 1978; Billings & Peterson, 1980; Brown et al., 1980). The microtopographic differences along the ice-wedge polygons lead to variations in soil moisture while larger elevation differences between the interstitial tundra and VDTLBs cause lateral redistribution of water into the lower lying VDTLBs (Engstrom et al., 2005). The surface hydrology is important in controlling ecosystem CO$_2$ flux and energy balance (Hinzman & Kane, 1992; McFadden et al., 1998; Oechel et al., 1998b; Chapin et al., 2000, 2005) as well as structuring vegetation communities (Walker et al., 2006). The surface moisture affects the NDVI in several ways including directly by the presence of standing water (see Fig. 8) and indirectly by affecting the surface vegetation cover (Engstrom et al., 2008; Zona et al., 2010). Hence, we attribute the differences in NDVI and CO$_2$ fluxes between the surface features on the Barrow Peninsula to a combination of surface soil water content and vegetation cover.

The vegetation within the Young- and Medium-aged VDTLBs are more productive than the older surface types (Billings & Peterson, 1980; Bliss & Peterson, 1992) (see also Table 3) with the wettest areas having the most productive vegetation (Zona et al., 2010). Young VDTLBs have a thin soil organic layer, little standing dead leaf material, and have had insufficient time for the formation of polygonal ground and ponds (Billings & Peterson, 1980; Hinkel et al., 2003; Bockheim et al., 2004). They are thought to have high nutrient availability as redistribution of water from higher to lower elevations accumulate nutrients (Kummerow et al., 1987) in the lakes which become available to the plants once the lake drains. Long-term accumulation rates are estimated to be the highest for Young VDTLBs averaging 967 g CO$_2$ m$^{-2}$ yr$^{-1}$ (Bockheim et al., 2004). Medium-aged VDTLBs have typically drier surfaces due to ground heave and flat-centered polygonal tundra forms, and though soil respiration may increase due to aeration, net CO$_2$ uptake persists as plant productivity is still relatively high though long-term accumulation rates decrease (Bockheim et al., 2004).

The long-term carbon accumulation rates of the Old and Ancient VDTLBs, and interstitial have been reported to decrease with age and is assumed that older ecosystems have lower productivity (Billings & Peterson, 1980; Bockheim et al., 2004), however, we were unable to differentiate CO$_2$ flux differences between the older surface types (Table 6). This is probably due to the formation of patterned ground and from the formation of low centered polygons and merged ponds that increase the soil water content and raise the water table. Plant productivity may remain high as organic matter is mineralized and become available while the soil respiration is depressed due to saturated soils and increased water table, a major control on soil respiration (Billings et al., 1982; Oberbauer et al., 1991; Oechel et al., 1998b; Olivas et al., 2010).

Considering that water is a major driver on ecosystem CO$_2$ flux (Oechel et al., 1998b), changes in water availability and alterations in surface hydrology can greatly affect the structure and function of arctic ecosystems (Hinzman & Kane, 1992), particularly in a warming and drying climate (Oechel et al., 2000a; Serreze et al., 2000; Hinzman et al., 2005). Increased permafrost depth will result in more thermokarst (Davis, 2001) and surface subsidence (Jorgenson et al., 2006; Schuur et al., 2008) which can impact the surface drainage of the VDTLBs and potentially eliminate the formation of ice-wedges and polygonal tundra. Conversion from wetland vegetation to upland vegetation types may occur as the ground dries bringing changes to soil properties, which may exacerbate CO$_2$ release due to increased organic matter decomposition. In addition, thawing permafrost may also enlarge existing thaw lakes and stimulate CO$_2$ sequestration as organic matter will slowly accumulate within wetter ecosystems due to suppressed soil respiration. Given these uncertainties and the potential for large scale changes in the VDTLBs and the Arctic as a whole, the research presented here provides a means to assess regional spatial patterns of CO$_2$ fluxes, vegetation coverage and status, and energy balance.

Limitations and improvements of Aircraft measurements of surface processes

A key problem facing ecologists, modelers, and policymakers alike is how best to scale surface processes and their controls to larger regional or continental areas (Peters et al., 2008; Schimel et al., 2009). All approaches have their limitations and sources of uncertainty. The challenge is how to best link multiple types of data at complementary temporal and spatial scales to aggregate our understanding and constrain these uncertainties. Aircraft measurements provide a key spatial link between ground-based estimates (e.g. biometry, chambers, tower-based EC, inventories) to the lower boundary-layer, and from the lower boundary-layer to larger remotely sensed products (AVIRIS, MODIS, and other satellite-based data).

Aircraft operations are often limited by regulations prohibiting flight under severe weather or unsafe flight conditions (e.g. fog, icing), or in the case of the low level

(<10 m AGL) flux flights, conditions with no visual ground reference (e.g. night). During the arctic summer, however, nighttime conditions do not limit aircraft operations. Instead, on the Barrow Peninsula, weather conditions can limit visibility, particularly in the early morning when fog persists. These limitations introduce a bias to the aircraft-based flux measurements to relatively ‘good’ conditions and times, and therefore these aircraft-based CO2 fluxes should be interpreted and used as an average daytime flux. The aircraft measurements during this campaign, however, occurred around mid-morning to evening (1000 to 2000 AST) with occasional flights extending towards midnight. The correspondence of aircraft to the tower EC measurements is nearly 1:1 (see Fig. 3), hence we have high confidence in using these aircraft data in characterizing the fluxes across this landscape. The temporal pattern measured by the aircraft also shows good correspondence with the temporal pattern of the tower EC systems (see Fig. 6), also suggesting that the aircraft-based data is robust to provide diurnal measurements if flights can be flown throughout the entire day. Pairing aircraft- and tower-based EC measurements provides a mechanistic way to link both the spatial and temporal patterns across heterogeneous landscapes, and can be used to improve the scaling of surface measurements to the region.

The aircraft flux footprint is a spatial integration of the upwind surface area, which can conceptually be visualized as a rectangular swath with the length equal to the spatial averaging block (here it is 1 km) times the distance of the footprint function (see Table 4). During this campaign the flight height was kept as close to the surface as possible to ameliorate systematic underestimations of surface fluxes due to flux divergence with height (Mahrt, 1998), and to have high confidence in characterizing the fluxes with the land surface type. Increasing the flight height AGL increases the sampling footprint (integrating over larger areas) but at the expense of spatial resolution. Because of our ease of flying fixed low heights and because of the assumed large spatial heterogeneity in land cover types, our observational design was to optimize our ability to detect small differences in flux among these land cover types (increased resolution) by flying fixed, low-level transects. In this way, we were able to link, with high confidence, to the MODIS products. Other aircraft based measurements may have science requirements that support grid patterning (Ogunjemiyo et al., 1997), vertical transects, or other boundary layer characterizations.

Even though our approach was a simple NDVI parameterized inventory model, its success may be due in part in our ability to link surface flux estimates with NDVI-parameterized footprints of representative ecosystems. 24/7 solar angles and $R_n$ (<400 W m$^{-2}$) were small compared with other lower latitude ecosystems, and most all the measurements were made in stable, neutral or very weakly convective conditions. The ability to have strong independent flux estimates among ecosystem types may be quite different under conditions with high $R_n$ and strongly convective atmospheres, and also prohibit flying aircrafts close to the ground.

Since aircraft can quickly transition in situ turbulent structures, they can provide a more accurate representation of the spatial fluxes (and their variances) even across heterogeneous landscapes. Numerous techniques to link aircraft measured fluxes to surface heterogeneity are rapidly evolving (e.g. Ogunjemiyo et al., 1997; LeMone et al., 2007; Kirby et al., 2008; Alfieri et al., 2009; Hutjes et al., 2010) and combined with tower-based EC flux measurements increase the reliability of both the spatial and temporal flux patterns that is critical to improvements of scaling surface fluxes to the region. Future work might also include using enhanced MODIS data products incorporating sophisticated corrections for atmosphere and bidirectional reflectance (Huete et al., 2002; Román et al., 2009), use of higher resolution satellite-based imagery such as IKONOS (Dial et al., 2003), QuickBird, or the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) sensor (Abrams, 2000), as well as using other advanced vegetation indices such as the Enhanced Vegetation Index (EVI) and the Land Surface Water Index (LWSI).

Conclusion

Aircraft-based EC systems have the ability to cover large spatial scales while measuring the turbulent fluxes across a number of surface and land use types (e.g. Desjardins et al., 1989; Desjardins et al., 1992; Kelly et al., 1992; Crawford et al., 1993b, 1996b; Brooks et al., 1996; Oechel et al., 1998a; Issac et al., 2004; Gioli et al., 2006; LeMone et al., 2007; Mauder et al., 2008; Alfieri et al., 2009; Hutjes et al., 2010) and combined with ground- and satellite-based measurements provide a valuable tool for both scaling and validation of regional-scale fluxes.

Here, we demonstrated that aircraft-based measurements; (i) provide an upscale link to satellite based NDVI, (ii) partition landscape scale carbon source and sink status for a range of vegetation classes, and (iii) used in tandem, can be used to estimate regional rates of carbon exchange. We used a simple inventory scaling approach to estimate a spatially averaged, mean, midsummer regional flux rate of $-2.04 \times 10^5$ kg CO$_2$ h$^{-1}$ over the 1802 km$^2$ Barrow Peninsula study region ($-0.032$ mg CO$_2$ m$^{-2}$ s$^{-1}$) that used ground-, aircraft- and satellite-based measurements.
Acknowledgements

This work was funded by grants from the National Science Foundation under award OPP-0436177, and the Department of Energy Western Regional Center of the National Institute for Climatic Change Research under award DE-FC02-06ER64159. Additional support for W.T. Lawrence by the NOAA funded Center of Excellence in Remote Sensing project is also acknowledged. The authors would like to thank the Barrow Arctic Science Consortium and Polar Field Services for logistics support while in Alaska. We are thankful for the field support from C. Laskowski, J. Verfaillie, C. Sturtevant, and R. Bryan, as well as GIS assistance from J. Iles. We are grateful to K. Hinkel for providing us with the GIS data regarding the drained thaw-lake basins on the Barrow Peninsula, N. Kijon for providing us with MATLAB code for her footprint model, UNAVCO for the additional GPS base station data in Barrow, and TerraMetrics for providing the TruEarth® 15-meter resolution imagery. This paper has benefited by conversations with R. Desjardins, as well as comments on the manuscript from the subject editor and three anonymous reviewers.

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Conclusion

An important problem facing scientists today and in the future, is how best to scale surface processes and their controls to larger regional or continental scales (Peters et al., 2008; Schimel et al., 2011), and how to link surface observations and data assimilation approaches at complementary temporal and spatial scales to aggregate and constrain the uncertainties in scaled and modeled estimates (Desai et al., 2010). Measurements of regional-scale surface fluxes and validation of remotely sensed satellite data products and model outputs are therefore necessary to understand the processes that determine a region’s influence on local and global carbon budgets and climate (Dolman et al., 2009; Desai et al., 2010).

SERA-based EC flux platforms have demonstrated the ability to differentiate various landscapes in terms of their surface fluxes across a range different ecosystem types and land surfaces (e.g., Crawford et al., 1993; Brooks et al., 1996; Crawford et al., 1996; Oechel et al., 1998; Gioli et al., 2004; Gioli et al., 2006; Migletta et al., 2007; Hutjes et al., 2010), and here we demonstrated that the SDSU Sky Arrow SERA was effective in characterizing and distinguishing land-atmosphere fluxes at spatial resolutions of 1 km across vastly different heterogeneous landscapes. Combined with tower-based EC, satellite-derived data products (e.g., NDVI), and spatial information (e.g., land surface types or ages), we were able to use a simple inventory scaling approach to estimate regional-scale CO₂ fluxes.

Here, we demonstrated that SERA-based measurements (i) provide an upscale linkage to satellite-based NDVI, (ii) partition landscape scale CO₂ fluxes for a range of ecosystem types, and (iii) when used in conjunction, can be used to estimate regional-scale CO₂ fluxes. The SERA-based methodology presented here, provides a mechanistic linkage to both the spatial and temporal patterns across heterogeneous landscapes which can be used to instruct, verify, validate, and constrain the scaling of surface observations and regional model outputs.
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