Mass and Energy Exchange of a Plantation Forest in Scotland Using Micrometeorological Methods

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University of Edinburgh
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Declaration

This thesis has been composed by myself from the results of my own work, except where stated otherwise, and has not been submitted in any other application for a degree.
Dedication

I dedicate this thesis to my Mother and Father, who have blessed me with an admiration of nature. Their patience, encouragement and love have sheltered me in the hardest of times and inspired me with the freedom to err and thrive at all times.
Acknowledgements

Much gratitude is due to my advisors John Moncrieff and Paul Jarvis. Their support and tolerance have allowed the development of this thesis and the ideas contained within its pages. I wish also to thank my examiners Ray Leuning and the internal reviewer for their selfless examination of this extensive manuscript.

Many thanks go to Keith McNaughton who maintained that fine line between interest and boredom in difficult topics, and whose darkening of my door was equally matched by mine of his. Kate Heal was most helpful in her assistance in determining site DBH and provision of throughfall and stream flow data. Matt Williams and Belinda Medlyn are thanked for provision of model results for comparison with measured data and Colin Legg for attempting to answer my odd statistics questions. John Grace, Maurizio Mencuccini, and Keith Smith were most useful for sharing the occasional beer, meal, and random discussion.

Thanks go to Allan Kelly, John Massheder, and Ford Cropley for listening to my questions about programming and for giving useful answers which helped to move EdiRe along to a successful conclusion. Also of invaluable help were the EdiRe users Allesandro Cescatti, Colin Lloyd, Hank Loescher, Sune Moller, Ed Swiatek, and Georg Wolfhart who helped find and solve many of the problems encountered during development.

The efforts of Steve Scott in getting the Griffin site up and running and construction of the Edibox and the power system were invaluable and the contributions of Lisa Wingate’s on site field work and analysis helped to fill many gaps in the Griffin results. I would also like to thank her for keeping me awake on the trips back and forth to Griffin. Also, Franz Conen was a most helpful and friendly companion and contributed his soil carbon measurement results. On the more theoretical end of the task, I would like to thank John Finnigan, Bart Kruijt, Yadvinder Mahli, Patrick Meir, and Mark Rayment for their help in answering questions about the myriad of theoretical and analytical details associated with the results from this experiment. I would also like to thank Jamie Trembath for keeping our field sites running while I wrote up this experiment and for being a clever fellow.
Others have contributed their time in listening to my complaints and have been good and understanding friends when the need arose. I would like to thank Allen Young, Sandra Patino, Mike Clearwater, and Caroline Nichol for helping to share the mental and emotional load of doing a PhD. There are many others in the department who I also thank for their cheerful friendliness: Jordi Martinez-Valialta, Emiliano Pegararo, Ana Ray, Fiona Carswell, Gail Jackson, Sophie Hale, Maree Lucas, and Peter Levy always had smiles on their faces and a nice word to say, so thank you.

There were many staff at the university who were most helpful throughout the course of my study. In particular I would like to thank Sheila Wilson, Connie Fox, Graham Walker, and Derek Scott for their organizational contributions and Bob Astles, Andy Gray, and Malcolm Ritchie and the numerous servitors for the more practical aspects of research within the confines of the Darwin building. Also, the fine workmanship of Alex, Dave, Graeme, George, and Peter from the workshop were crucial in creating some of the equipment used in the field and in keeping some of the vehicles running.

Last, I would like to thank my many flatmates: Jane Norton, James Hepher, Sara Marrow, Mark Goodwill, Ingo, Miga, Hazel, Gareth, and that other guy who’s name I forget. You mostly a pleasure and at worst interesting, but I will always remember you – even the last guy.

Tillhill Forestry Ltd. made the Griffin research site available for our use and I wish to thank them for making this research possible.

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Abstract

This thesis presents the energy, water, and carbon budgets of a Sitka spruce plantation forest in Scotland over the period 1997 to 2001. The site's infrastructure, layout, and methods employed in site operation are described and a detailed analysis of the temporal and environmental dependencies of microclimatological measurements is presented. The site microclimate is observed to be strongly influenced by the site's oceanic climate, and canopy development. Atmospheric structure is observed to affect temporal patterns of microclimatological variables while topography is observed to affect microclimatological and flux measurements. Eddy covariance flux measurement theory and methods are examined and specific inadequacies are addressed. Theoretical aspects of eddy covariance that were examined include signal despiking, coordinate rotation, low frequency flux contributions, as well as corrections for density fluctuations, angle of attack errors, and sonic temperature determination. An analysis of frequency response correction methods was used to determine if superior methods could be identified. Fluxes of momentum were used to verify existing measures of atmospheric turbulence and analysed to identify canopy structure and growth. Sensible heat fluxes were found to have an unexpected negative bias, only a portion of which can be attributed to instrument error. This bias is found to depend upon topography and wind speed but is apparently unrelated to katabatic flow. Large errors in latent heat flux were caused by enhanced tube attenuation and were corrected using improved frequency response corrections. Interannual variability of momentum and sensible heat flux were closely associated with wind speed variability, while interannual variability of net ecosystem exchange was attributable primarily to radiation. The source of variability of latent heat flux was not clearly identifiable. Missing values of latent heat flux were modelled using a canopy conductance model, which incorporated effects of canopy evaporation. Missing values of nocturnal net ecosystem exchange were obtained using a temperature dependent model of ecosystem respiration. This model incorporated effects of friction velocity and topography. Diurnal values of net ecosystem exchange were modelled using a light response model of gross ecosystem exchange. This model included dependencies on temperature, vapour pressure deficit, and cloudiness. A comparison of measured net ecosystem exchange data with the output of two
models suggests that a diurnal version of the nocturnal flux loss problem may exist. The hydrology budget losses are within (~10%) of precipitation gains and energy budget components obtained 90% closure. Annual net ecosystem uptake for Griffin is estimated as 749 g m\(^{-2}\), and residuals of the carbon budget components fall within the range of experimental measurements.
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Symbol List

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Roman Characters

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**Acronyms**

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### Greek Characters

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<td>Density of H₂O</td>
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<td>u</td>
<td>Standard deviation of stream wise velocity</td>
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xxiv
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<th>Units (value)</th>
<th>Equ. (sect.)</th>
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<td>Shear stress</td>
<td>N m(^{-2})</td>
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<td>Time constant</td>
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<td>Equivalent sensor time constant</td>
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<td>Transmissivity of atmospheric water vapour</td>
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<td>B.59</td>
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<td>ν</td>
<td></td>
<td>Viscosity of air</td>
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<td>Atmospheric scattering coefficient</td>
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<td>Albedo model light scattering coefficient</td>
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<tr>
<td>ξ</td>
<td></td>
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<td>ξ</td>
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<td>Latitude</td>
<td>degrees</td>
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<tr>
<td>Ψ</td>
<td></td>
<td>Stability adjustment function</td>
<td></td>
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Proposition 37
No improvement can take place in the Art of the present generation until all classes, Artists, Manufacturers, and the Public, are better educated in Art, and the existence of general principles is more fully recognised.

Proposition 13
Flowers or other natural objects should not be used as ornaments, but conventional representations founded upon them sufficiently suggestive to convey the intended image to the mind, without destroying the unity of the object they are employed to decorate. *Universally obeyed in the best periods of Art, equally violated when Art declines.*

Proposition 4
True beauty results from that repose which the mind feels when the eye, the intellect, and the affections, are satisfied from the absence of any want.

Owen Jones
*The Grammar of Ornament, 1856*
1 Introduction

The Earth’s climate has experienced large variations over its history (IPCC Working Group 2001; Karl & Trenberth 2003), yet concern has been expressed over the current changes of the Earth’s climate. These concerns arise from the high probability that current climate change has been induced by anthropogenic activities (Karl et al. 2003). At the start of the 21st century, the rise in world temperature has become large and rapid enough to achieve significant correlation with anthropogenic factors - particularly rises in greenhouse gases. The potential effects of rising temperatures and the possible associated changes in climate behaviour (IPCC Working Group 2001) are likely to have significant economic and social effects on humanity (King 2004), as well as potentially disastrous effects for many of Earth’s other species (Walther et al. 2002). How humanity addresses climate change may determine the success of institutions and species alike (Tilman et al. 2001; Dietz et al. 2003; Hasselmann et al. 2003). In order to take the correct steps in addressing climate change, it is important to understand the effects of human actions before they are carried out, but achieving this understanding requires the modelling of a system of infinite complexity.

Although current models do an admirable job of assessing the effect on climate of both human and natural processes (IPCC Working Group 2001), these models can only produce results of a quality no better than the quality of the information upon which they are based (Betts et al. 2001; Wang et al. 2001; Suntharalingam et al. 2003).

The ‘greenhouse gases’ alter the solar and terrestrial radiation absorption of the atmosphere and as a result influence the radiation budget of the Earth. Depending upon their radiation absorption characteristics, their chemical interactions and their position in the atmosphere, these gases may cause either warming or cooling of the Earth (IPCC Working Group 1996). These complex interactions can be quantified as radiative forcings (units of W m$^{-2}$), which describe, for the atmosphere below the tropopause, the increase in net radiation required to produce an atmospheric warming equivalent to that caused by a given change in the atmospheric burden of a greenhouse gas (IPCC Working Group 2001).
There are a number of gases and particulate matter that may cause radiative forcing, including carbon dioxide, methane, nitrous oxides, halocarbons, as well as aerosols, ozone, and water vapour (UNEP 1987; IPCC Working Group 1996; IPCC Working Group 2001). Carbon dioxide has the largest radiative forcing relative to pre-industrial conditions, despite its unit forcing potential being smaller than many other gases. Carbon dioxide’s larger radiative forcing is the result of the rapid increase of atmospheric CO₂ concentration over the past two centuries (Andres et al. 1999). This rapid increase is the result of exploitation of the highly concentrated energy afforded by fossil fuels that has fostered the rapid expansion of human endeavour. This stored energy is itself the result of millennia of carbon sequestration by photosynthesis, which is now being released back to the atmosphere as a result of burning of these fossil fuels (British Petroleum 2003). Photosynthesis is the means by which plants store solar energy and as a result the source by which humans and almost all species on earth derive energy for growth. As a result, the cycling of CO₂ and our understanding of its sources and sinks, and processes that affect these sources and sinks is critical for our understanding of the role CO₂ will play in the future of climate change.

Current estimates indicate that the annual anthropogenic contribution of CO₂ to the atmosphere is about 6.3 Pg (C) a⁻¹. The primary source of this contribution is the exponential increase in fossil fuel burning and cement production since the 18th century and which currently accounts for about 75% of anthropogenic source. The bulk of the remaining anthropogenic sources can be attributed to land use change (IPCC Working Group 2000).

While anthropogenic sources of CO₂ are reasonably well understood, our understanding of the more diverse sinks of CO₂ is less well developed. Primary sinks are associated with either ocean or terrestrial vegetation. No significant anthropogenic or geological CO₂ sinks currently exist. The uptake of CO₂ by the oceans of about 1.7 Pg (C) a⁻¹ occurs by direct solution of CO₂ into ocean water and through buffering reactions of carbonates and bicarbonates. The terrestrial uptake of carbon, of the order of 1.4 Pg (C) a⁻¹, is a result of photosynthesis by vegetation.
This relationship of photosynthesis to some of the primary greenhouse gases gives vegetation a key role in climate modelling. The uptake of carbon by the vegetation is however, a spatially and temporally variable factor (Oncley et al. 1997; Katul et al. 1999; Chen et al. 2003; Leuning et al. 2004;) that can be strongly influenced by human management. For this reason, much effort is being put towards understanding the responses of different ecosystems to climatic and human influences and how these responses influence the carbon cycle of ecosystems.

Early studies of the carbon exchange focused on basic understanding of photosynthesis under field conditions, providing the theory upon which more complete estimates of the carbon cycle could be based (Verhagen et al. 1963; Monteith 1965; Duncan et al. 1967; Yoshida 1972). It was not until the development of suitable measurement devices in the 1960's and 70's that in-situ measurement of CO₂ exchange became a practical endeavour (Baumgart 1969; Lemon et al. 1970; Saugier 1970). Although such larger scale carbon exchange measurements were attempted, measurement systems were generally not capable of long-term spatially averaged carbon exchange measurement (Brown & Rosenberg 1968). Carbon exchange could still be more reliably estimated for vegetation and soil using changes in carbon stocks. Such estimates did not, however, provide knowledge of the temporal variation of carbon exchange on timescales shorter than seasonal, annual, or longer.

With the advent of more robust CO₂ measurement devices (Bingham et al. 1978; Jones et al. 1978; Suyker & Verma 1993), CO₂ exchange processes could be measured over scales of time and space relevant to photosynthesis and respiration. As a result improved knowledge of the processes of ecosystem carbon cycle processes led to improvements in process models of carbon cycle components. These studies were often limited in extent because of the expense, robustness, and portability of the measurement systems. Increased political demand for knowledge of carbon cycle processes at the end of the 20th century (IGBP Terrestrial Carbon Working Group 1998; Schulze et al. 2002) and recent advances in electronics and engineering have given scientists the ability to make temporally and spatially extensive measurements of carbon exchange processes for representative ecosystems. These advances have lead to a more comprehensive view of how the carbon cycle reacts in its natural
environment. Concurrently, experimentation has developed from individual field experiments to large-scale organized experiments.

Efforts at developing comprehensive data sets began with the advent of large multidisciplinary projects such as ABLE, HAPEX, FIFE, BOREAS, LBA etc. (Sellers et al. 1992; Goutorbe et al. 1997; Sellers et al. 1997; Avissar et al. 2002). These projects enabled coincident, detailed research into selected regional ecosystem dynamics over limited time scales. Since the mid 1990s, such research has broadened into loose networks of researchers supplying information about the ecosystem in which they work as part of an organization of networks (Euroflux, Ameriflux, Ozflux, Asiaflux etc.) under the umbrella of a cooperative entity: FLUXNET (Baldocchi et al. 1996; Running et al. 1999; Baldocchi et al. 2001). This network, currently consisting of over 200 sites, shown in figure 1.1, conducting related experimental research, has been developed to provide a more consistent, organized source of experimental data from which necessary information can be drawn for the purposes of process definition for climate modelling. Yet, it can be seen from figure 1.1 that the distribution of flux measurements has distinct political and economic overtones that will limit our knowledge of ecosystem responses to climate.

![Figure 1.1 Locations of FLUXNET experimental sites. Site latitude and longitude were obtained from the FLUXNET web site and include all listed sites in 2004.](image)

It is likely that, similar to the effect of the initial development of laboratory-based equipment had on the advancement of knowledge in the underlying processes of
photosynthesis, current technological and methodological advances will lead to advances in our knowledge of ecosystem-scale carbon cycles and their effect on the Earth's climate.

Given the number of experiments on ecosystem carbon cycle research, it is perhaps surprising that there should be a lack of knowledge of ecosystem carbon cycle processes. The development of this relatively widespread network of flux measurements is, however, a relatively recent phenomenon. An idea of just how recent it is can be seen from its presence in the scientific literature. A search for journal articles that refer to either "CO$_2$ flux" or "carbon dioxide flux" since 1980 and a similar search for articles referring to "annual CO$_2$ flux" are shown in figure 1.2. While such a focused search will not capture all relevant information on this topic it does show the increase in ecosystem level carbon cycle research, starting in the late 1980s with the improvement of CO$_2$ and wind velocity measurement techniques. Given the recent widespread development of ecosystem carbon cycle measurements it is not surprising that standard systems of measurement, quality control, and application of theory are not yet fully developed.

![Figure 1.2 Number of journal articles containing references to either "CO$_2$ flux" or "carbon dioxide flux" (filled circles) and on the right axis the number of articles containing the phrase "annual CO$_2$ flux", (open circles). References were obtained from the ISI Web of Science database.](image)

Despite the rapid increase in ecosystem carbon research many questions still remain (Shackley et al. 1998; Visser et al. 2000; Baldocchi 2001; Wang & Hsieh 2002; 2003; Enquist et al. 2003;). Both spatial and temporal expansion of carbon cycle
measurements are needed to constrain what are currently imprecise measurements. Such objectives may be achieved by expanding the scale of measurements through the use of such techniques as inverse modelling, boundary layer budget flux estimation, aircraft flux measurements and remote sensing. Such techniques will in turn require high precision ground truth values obtained through micrometeorological and stock assessment methods. Further research into the underlying physiology of photosynthesis and respiration is needed to support both measurement and modelling of the carbon cycle. This effort must extend over time periods that capture a wide range of climate variability; such time scales are likely to be on the order of decades (Baldocchi et al. 1996).

To these ends, this thesis provides a set of measurements describing the interaction of a representative forest stand with its environment in an unique climatological region of Europe. The data presented in this thesis represent five years of research carried out at one site, Griffin Forest, in association with the Euroflux network.

The location of Griffin in central Scotland (56° 36' 23.59" N, 3° 47' 48.55" W) on the western edge of Europe and the eastern edge of the Atlantic Ocean, gives Griffin a unique climate relative to other flux sites and establishes the site’s significance. Griffin’s position in the climate continuum for both Europe and the world is indicated on the frequency distributions of selected climate variables in Figure 1.3.

The region in which Griffin lies stands out as having high precipitation, cloud cover, and percent wet days; this combined with moderate temperature and below average temperature range creates its unique environment. It is readily apparent that the proximity of central Scotland to the Atlantic Ocean and the Gulf Stream have a moderating effect on its climate. It can be seen from figure 1.3 that the climate of central Scotland is both humid and temperate, and the same data plotted as a precipitation-temperature climate space diagram figure 1.4 reinforces the unique climate of Griffin and shows its unique position relative to other sites in the Euroflux network. The plentiful moisture and moderate temperatures are likely to be responsible for the relatively large productivity of the Griffin ecosystem.
Figure 1.3 Land surface frequency distribution of selected annual average climate variables for Europe (solid line) and the world (dotted line) obtained from 0.5 deg gridded climate data (New et al. 1999). Descending arrow indicates climate conditions for grid cell enclosing Griffin experiment site. The global, interpolated, observed data presented here were obtained from the University of East Anglia's IPCC data distribution centre and are described in New et al. (1999).

The moderating influence of the Atlantic Ocean promotes an environment that displays less interannual variability than other sites within the Euroflux network. This influence may also moderate the effects of climate change of this site relative to effects in other regions of Europe. However, it is suggested that these climate change effects, when referenced to interannual variability, may be proportional to changes observed in other regions.
With consideration to the importance of research at Griffin and given the capability of a research environment with extensive prior research experience and facilities, a programme of experimental research was set out to achieve the following goals:

- To characterize the forest, soil, microclimatology, and ecosystem exchanges.
- To identify potential errors in the measurement, processing and analysis of variables and fluxes and to suggest possible solutions to those problems.
- To characterize the relationships between relevant environmental variables and exchange processes.
- To establish a justifiable multi-year record of carbon and energy exchange.

The remainder of this thesis purports to address these research goals. Although most of the relevant information is contained within the following eight chapters, some has been seconded to appendices to allow a more compact document. The following list briefly describes the chapters and appendices and the content therein.
• Chapter 2 Experimental site: describes experiment site, characteristics of vegetation and soil and experiment infrastructure.

• Chapter 3 Canopy microclimate measurements: describes methods of collection and analysis of climatological variables.

• Chapter 4 Griffin climatology: describes site climatology for the five-year experimental period, 1997 to 2001.

• Chapter 5 Ecosystem exchange methods: describes methods of ecosystem exchange determination.

• Chapter 6 The effect of time averaging on the estimation of fluxes: describes the development of a temporal averaging transfer function.

• Chapter 7 Comparison of frequency response correction methodologies: describes an analysis of methods of frequency response corrections.

• Chapter 8 Griffin ecosystem exchanges: describes ecosystem exchange results and provides estimates of water, energy, and carbon budgets.

• Chapter 9 Conclusions and recommendations: Concluding comments.

• Appendix A Climatological measurement methods: describes sensor deployment, maintenance, and calibration methods.

• Appendix B Derivations, equations, and models: presents derivations, equations and models employed in the thesis.

• Appendix C Plot measurements: tabulates measurements associated with site sample plots.

• Appendix D Instrumentation and equipment: tabulates instruments and equipment used in the experiment and their associated calibration factors.

• Appendix E Instrumentation time lines: describes instrument installation and maintenance time lines.

• Appendix F Software: tabulates software used in this project and provides description of statistical validity of eddy covariance calculations.

• Appendix G Signal quality control ranges: Provides values of signal default quality control ranges employed.

• Appendix H A correction for sonic temperature errors resulting from flow acceleration and sensor head distortion: Describes development of a correction to sonic temperature measurements for errors caused by sensor probe deformation and air flow acceleration.
2 Experiment site

This chapter introduces the field site used in this study. It considers the geological and biological features of the area that bear upon the measurement methods and results described in later chapters. This chapter also introduces some practical issues associated with fieldwork at the site, and describes the characteristics and variability in of the soil and forest.

2.1 Site natural history

A greatly simplified summary of the geology of Scotland shows that much of the country consists of metamorphic Precambrian and Lower Palaeozoic rock, while more southerly regions are characterized by more recent sedimentary deposits (Read & Watson 1966). The geology of Scotland is similar in origin to much of Scandinavia, Wales and Ireland (West 2002). As a result of easterly glacial flow in the southern Scottish highlands during the Devensian/Weichselian glacial period, soils in this region have developed from glacial till whose origin was the schistic Dalradian stratum of the central highlands (Curtis et al. 1976; West 1968).

The advancement of vegetation into Britain quickly followed the last glacial retreat, approximately 10,000 years B.P.. Based primarily on archaeological investigation of pollen deposits, vegetation is thought to have advanced from sphagnum-dominated tundra to *Betula* and *Salix* dominated communities followed by invasion *Corylus* and *Pinus*. In more southerly parts of Britain, greater amounts of *Quercus*, *Tilia*, and *Ulmus* became the dominant species (Evans 1975; West 1968). Clearance of land by human application of fire may be suggested by the presence of *Corylus* species (Evans 1975), but extensive agricultural clearances in Scotland are not believed to have occurred before the end of the first century AD with the development of a more agricultural and pastoral economy (Anderson 1967). Following the development of agriculture, the amount of forest declined slowly, reaching a minimum of about 5% at the beginning of the 20th century (Best 1959). Preparation for the First World War brought about the realization of the perilous condition of forestry resources in the UK and the Forestry Commission was created to re-develop forest resources. After the Second World War, most land in Scotland was either used as pasture (30%) or was
classified as mire (30%) with forests constituting only 5% of land cover. Ambitious plans to achieve 2 million hectares of forested land in the UK by 2000 (Best 1959) have been achieved; with forest cover now approximately 2.7 million hectares (Forestry Commission 2001a). With the relative recovery of UK forests, the amount of forest cover is approximately 10%, with an even higher percent of land cover in Scotland (15%). In comparison, agriculture accounts for approximately 40% of land cover, of which about 75% is grass or grazing (Mackey et al. 1998). Development of forests has taken place primarily on hilly terrain and other areas less suitable for agriculture. This is reflected in the reduction of land area of mire and pasture land (Mackey et al. 1998). However, the level of forest cover in the UK and Scotland is still lower than most EU countries, which averaged 25% forest cover and lower still than the world average of 31% forest cover (Forestry Commission 1991; Mackey et al. 1998).

![Map of Scotland and Tayside region](image)

**Figure 2-1** Map of Scotland (right) and Tayside region (left) with woodlands over 2 ha in dark grey. Black star indicates location of Griffin field site. Maps taken from Forestry Commission, National Inventory of Woodland and Trees, Scotland – Tayside region inventory report (Smith & Gilbert 2002).

Coniferous species currently have a greater coverage, (60%), than broadleaf species primarily as a result of extensive plantings of conifers since 1940. Because of its characteristics, (MacDonald 1967) the species of choice for these plantings has been Sitka spruce (*Picea sitchensis* (Bong.) Carr.). This is now the dominant UK forest cover species, accounting for approximately 30% of all forest cover. Because of
favourable climate conditions, 76% of UK Sitka plantations are located in Scotland (Forestry Commission 2001b).

In the Tayside region, in which this experiment was situated (figure 2.1), most land in valley bottoms is employed for agricultural purposes, while land on higher ground is employed as rough grazing land. Similar to Scotland as a whole, forest cover in Tayside comprises 12.9% of land cover and Sitka spruce constitutes 29.0% of all forests.

2.2 Site selection

It is obvious that, to provide a representative measure of the ecosystem exchanges associated with forestry in the UK, Sitka spruce plantations must be included. To establish a field site, crop type, climate conditions, soil type, topography, reasonable controlled access to the site, the presence of facilities near the site, as well as a working relationship with forest manager must all be considered. The selected site, shown in figure 2.1, provided a representative site and secure conditions. Practical limitations of power, travel and transport were onerous at time and, although not scientific in nature, such limitations did influence the conduct of the experiment. In the following sections we describe and justify relevant aspects of development of the experiment site.

2.3 Site infrastructure

Ideally, the surface vegetation at an experiment site would consist solely of the vegetation to be studied. Unfortunately, until suitable remote sensing devices are developed, scientists must inevitably modify an environment to obtain its characterization. As an example, for the conduct of this experiment it was necessary to install towers and power generation facilities and the amount of equipment used during the experiment made vehicular access to the site necessary. While development of sampling plots was also necessary, most of the sampling carried out during this experiment was non-destructive. Disturbances caused by infrastructure and human presence at the site was minimized as far as practical.
2.3.1 Access

The site was located approximately 80 miles from the University of Edinburgh, thus making single day visits to the site possible. The majority of this distance was travel on motorway and maintained country roads; only the final three miles was on dirt roads with the final 500 m on non-maintained farm track that bisected the site. At the beginning of the experiment this track was in good repair, though not of acceptable condition for use by 2-wheel drive vehicles. In 1999 the department acquired a 4WD vehicle, which greatly enhanced access to the site and allowed for an extra hour of work to be done at the site. Vehicle access increased erosion to the track, necessitating road repair by the end of 2000. During this repair, drainage channels were installed which crossed the road to prevent water flowing down the track during times of heavy rain and snowmelt. It is believed that these repairs did not affect the biological or hydrological character of the experiment.

2.3.2 Power

Several sensors used in the experiment required a continuous electrical power supply making it necessary to have power available on site. Initial inquiries into providing mains power at the site indicated that the expense was too high to justify its installation for the duration of what was initially a three-year project. It was therefore necessary to evaluate on site power production facilities.

2.3.2.1 Determination of needs

Power requirements for the Griffin site were based on the power usage of the eddy covariance and meteorological sensors and did not include equipment used for periodic calibration or maintenance. Power use specifications were obtained from equipment manuals and are assumed to show normal instrument operation; the power use requirements are given in table 2.1. All equipment was powered from 12 volt batteries, which had their voltage levels maintained by the installed power generation system discussed in the following sections.
Table 2.1 Power consumption estimates for Griffin field experiment. Observed usage is based on typical 7 amp draw at 24 V.

<table>
<thead>
<tr>
<th>System Components</th>
<th>Power, W</th>
<th>Number</th>
<th>Total power, W</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample cell pump</td>
<td>24.0</td>
<td>2</td>
<td>48.0</td>
</tr>
<tr>
<td>Licor 6262</td>
<td>12.0</td>
<td>2</td>
<td>24.0</td>
</tr>
<tr>
<td>Mass flow controller</td>
<td>3.3</td>
<td>1</td>
<td>3.3</td>
</tr>
<tr>
<td>Gill Solent anemometer</td>
<td>1.8</td>
<td>1</td>
<td>1.8</td>
</tr>
<tr>
<td>Solenoid</td>
<td>2</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>Computer</td>
<td>15.0</td>
<td>1</td>
<td>15</td>
</tr>
<tr>
<td>Battery Chargers</td>
<td>5.0</td>
<td>3</td>
<td>15</td>
</tr>
<tr>
<td>Psychrometers</td>
<td>2</td>
<td>3</td>
<td>6</td>
</tr>
<tr>
<td>Loggers/sensors</td>
<td>1</td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>Inverter</td>
<td>17.0</td>
<td>1</td>
<td>17</td>
</tr>
<tr>
<td>Line loss</td>
<td>5</td>
<td></td>
<td>5</td>
</tr>
<tr>
<td>Estimated</td>
<td></td>
<td></td>
<td>138.1</td>
</tr>
<tr>
<td>Observed</td>
<td></td>
<td></td>
<td>168</td>
</tr>
</tbody>
</table>

2.3.2.2 Power generation methods

While several methods of power generation were considered, some methods such as hydroelectric and fuel cells were found to be too expensive after initial consideration. The three methods with power generation potential applicable to this site included solar panels, wind turbines and fossil fuel generators. Each option had good and bad aspects and it was decided to employ a combination of all three.

Storage batteries

Unfortunately, neither of the alternative power systems was capable of producing a continuous power output of the level required by the experiment, and it was desirable to minimize the use of fossil fuel based power. It was therefore necessary to employ a bank of storage batteries to provide a continuous supply of power. An inverter was used to convert the 24 V DC of the battery bank into 240 V AC that was transmitted to the towers and converted to 12 V for use by the instrumentation. The wind, solar, and fossil fuel generators were used to charge the bank of storage batteries. Because of the frequent discharge/recharge cycle of the battery bank, it was necessary to use deep cycle lead acid batteries. These batteries came in several configurations (2V, 6V, 12V) and had capital costs that ranged from £2.50 to £8.50 Ahr⁻¹. It was decided to employ a set of six 12 V batteries connected in series and parallel to act as a set of
three 24 V batteries. The batteries were placed close to the generator to minimize line losses while charging the batteries.

**Fossil fuel power generator**

Fossil fuel generators are relatively inexpensive and provide power on demand but require constant maintenance and contaminate the experiment site by exhausting one of the trace gases being measured. Nevertheless, the unreliability of the wind and solar systems made it necessary to use a generator. Propane was chosen as the fuel source to simplify transport of the fuel to the site. Because the generator was operated intermittently, it was necessary to obtain a generator with autostart capabilities. Because of its potential for contamination by exhaust gases, the generator should have been placed at either a great distance from, or at the base of the flux towers. These options were rejected because of power losses associated with long distance transmission and the potential for contamination under low wind speed conditions, respectively. Instead it was decided to place the generator midway between the two power sinks and to assess potential errors in retrospect. Generator operation was recorded in order to facilitate later analysis.

**Solar panels**

Solar panels are low maintenance, non-polluting, unobtrusive and have an improving economic efficiency in time, making them an obvious choice for generation of power during the summer months. Based on initial power usage estimates of 50 W, a BP representative estimated (personal communication) that an array of 5 panels would supply power to the system for 4 to 6 months of the year. Low solar angles during the winter months meant that at least thirty solar panels would have been required for year-round operation.

Six 75 W solar panels were deployed. The solar panels were located 4 m above the ground in a forest ride to reduce shading by trees. The panels were mounted using a support frame atop two sections of scaffolding. The scaffolding was used to construct an enclosure for the generator and battery bank, protecting them from the elements. This location also reduced line losses in transmission of power to the battery bank. The solar panels were placed facing south with an elevation angle of approximately 60 degrees to maximize the radiation available during the spring and fall seasons.
Wind turbine

Similar to solar panels, wind turbines are low maintenance, non-polluting, and have an improving economic efficiency with time. Data collected at a height of approximately 5 m from July through August 1996 indicated site mean wind speed of approximately 2.3 m s\(^{-1}\). Using the power/wind speed curves for a typical wind turbine and the collected wind speed data it was estimated that approximately 20% of the initial power use estimates could be satisfied by the wind turbine. By raising the turbine higher above the forest canopy and considering the increase in wind speeds of 1 to 2 m s\(^{-1}\) during the winter months it was estimated that greater than 60% of the initial estimate of power requirements could be met using wind power during the winter months.

The turbine was also placed near the battery bank to minimize line losses. The distance of the turbine from any wind sensors was at least 70 m. According to Beyer et al. (Vermeer et al. 2003), after passing a turbine, wind energy recovers to 95% of its upwind level upwind at a distance of 20 times the turbine rotor diameter. Given the rotor diameter of the turbine employed was 2.6 m it is likely that the wind turbine had no significant effect on the experimental measurements of wind.

2.3.2.3 Power production

As a system, the solar and wind generated power supplied 24-volt power to charge the battery bank. The battery bank was monitored so that if its charge level fell below 22.0 Volts the generator would start and charge the batteries until they reached a level of 26.6V.

Power consumption and production values are presented in figure 2.2. The power consumption of the site appears to vary between 140 and 230 W. This variation is likely to be caused by periods of downtime and missing logger data and does not reflect actual variations in the power consumption. Based on intermittent manual inspection, actual consumption appeared to be about 168 W (i.e. 7 A at 24 V).

The power produced by the generator was assumed to be that specified by the manufacturer (1000 W). It is likely that the actual production may have been less
than this amount. In either case it is immediately obvious from figure 2.2 that
generator power was overwhelming important, and inefficient, in satisfying the power
needs of the site. The generator's power contribution exceeded the site's power
requirements in all months.

![Bar graph showing average power consumption by month](image)

**Figure 2-2 Average power consumption by month. Values represent the
average amount of power used in a half hour period.**

The combined contribution of solar and wind power to the site's power requirements
averaged approximately 40 W, though generally higher in spring and summer and less
in the autumn and winter. This contribution was roughly equivalent to the original
power specifications for the eddy covariance system and had the additional power
needs of the profile system not been added these alternative energy sources would
probably have supplied over 50% of the site's power needs. Also, apparent from
figure 2.2 is the success in implementing the dual source alternate energy supply
system. The solar panels supplied the majority of the alternative power from March
through September, while increased wind in the winter months allowed greater wind
turbine power production from October through March. In general, however, the
solar panels produced about 30% more power than did the wind turbine. Some of the
loss in effectiveness of the wind turbine resulted from decreases in mean wind speed
at the height of the wind turbine due to growth of the forest. Forest growth also
affected solar panel power production by shading the panels at low sun angles.

A breakdown of the costs and economic efficiencies of the systems deployed are
presented in table 2.2 for 1997-2001. It is apparent that the power production costs of
the fossil fuel generator were lowest and the wind turbine's were the highest.
However, it must be kept in mind that the power produced by the generator was 37% greater than the actual power required by the site. If this were taken into account the effective cost per kWh of the generator would rise from £0.91 to £1.25 kWh⁻¹.

Table 2.2 Estimated power production costs for individual power generation methods used at the Griffin field site. Total production costs incorporate additional costs for power conversion and transmission. Total usage costs reflect the cost of producing the amount of power actually used.

<table>
<thead>
<tr>
<th>Item</th>
<th>Cost (£)</th>
<th>Est. # Required</th>
<th>Power Produced (kWh)</th>
<th>Cost (£/kWh)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Generator</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Propane generator</td>
<td>1000</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Replacement engine</td>
<td>660</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Autostart, and repair</td>
<td>300</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Propane Regulator</td>
<td>65</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Propane</td>
<td>5300</td>
<td>2/mo @ 40</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oil</td>
<td>48</td>
<td>4/year @ 12</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Subtotal</td>
<td>7373</td>
<td></td>
<td>8098</td>
<td>0.91</td>
</tr>
<tr>
<td><strong>Solar Panels</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Solar Panel</td>
<td>1620</td>
<td>6 @ 270</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mounting structure</td>
<td>400</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cable</td>
<td>30</td>
<td>100 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Subtotal</td>
<td>2050</td>
<td></td>
<td>1000</td>
<td>2.05</td>
</tr>
<tr>
<td><strong>Wind Turbine</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Turbine</td>
<td>1800</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12 m mast</td>
<td>1250</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mast base</td>
<td>1100</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Subtotal</td>
<td>4150</td>
<td></td>
<td>752</td>
<td>5.52</td>
</tr>
<tr>
<td><strong>Associated Items</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inverter Charger</td>
<td>1100</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>DC supplies</td>
<td>300</td>
<td>2 @ 150£</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cable</td>
<td>200</td>
<td>200 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Batteries</td>
<td>1740</td>
<td>6 @ 290£</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Power electronics</td>
<td>850</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Subtotal</td>
<td>4190</td>
<td></td>
<td>9850</td>
<td>1.80</td>
</tr>
<tr>
<td><strong>Total Production</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>17763</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Total Usage</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>17763</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The power production costs as a function of time, figure 2.3, were derived from information in table 2.2. The curves represent the cost of component power production except the 'system used' curve, which also includes costs for power storage and transmission. Note that the generator curves do not include effects of experiment.
contamination or regular system maintenance costs associated with fossil fuel systems. Significant effort was expended on maintenance and generator related quality assessment such that excluding consideration of these factors could bias planning of power implementation of future experiments.

![Figure 2-3 Decline in power production cost estimates as a function of years of usage. The 'Generator (kW usage)' curve represents cost based on power used instead of power produced by the generator.](image)

Based on the available information, it is recommended that future installations at this site, which do not have access to mains power, rely more heavily on solar power. This is especially true if the potential for increases in power requirements is small and the potential for long-term deployment is large. However, if potential contamination can be justified and manpower is not a limiting factor, for experiments of less than 10 years it is more economically efficient to deploy power sources that employ fossil fuel.

### 2.3.3 Towers

Triangular welded lattice towers were used for the deployment of meteorological instruments through and above the canopy. Two of these towers were installed at the site. Each tower consisted of five ten-foot sections and a base plate. The towers were guyed at 10 and 20 foot levels; guys were attached to separate guying stakes. It was necessary to tighten the guy wires after the second field season, and at that time the guy wire turnbuckles were secured to eliminate the possibility for self-loosening.
For a period in 1997, a short mast was placed 10 m southwest of the flux measurement tower. This mast was used for measurement of within canopy temperature fluctuations. A set of aluminum walk up scaffolding towers was installed in the northern end of the experiment site in 1999. These towers were associated with other experimental work being carried out on the site and did not interfere with flux or microclimate measurements.

2.4 Experiment sites

Several measurement sites were used in this experiment, as shown in figure 2.4. These sites can be classified by their purpose: flux, microclimate, and sample plots. The following section describes the placement and purposes of these experiment sites.

![Figure 2-4 Schematic of site showing tower, plot and road locations.](image)

2.4.1 Sample Plots

Forest sample plots were required to characterize the site and acted as focal points for ancillary experiments carried out over the course of the main experiment. To characterize the site properly, it was necessary to determine the plot size, the number of plots required and an appropriate spatial distribution of plots.
Initial considerations of randomised sample locations were dismissed after considering the impracticalities of locating, and relocating trees or plots in the dense near-surface canopy structure of the forest. Instead, the row structure of the plantation enabled determination by a systematic multistage sampling procedure. Additional clustered sample plots were placed in the vicinity of the towers. In this scheme, 49 sample plots were laid out in a 7 by 7 grid. Each plot was separated from its nearest neighbours by 100 m so that the grid covered a 700 by 700 m area. This design covered the majority of the experiment site.

The level of accuracy obtained with this sampling plan was obtained by the following method. The expected coefficient of variation, $D_e$, was determined from the planned sample size, $n$, and pilot measurements of the sample mean, $\bar{X}$, and standard deviation, $\sigma_X$:

$$D_e = \frac{\sigma_X}{\bar{X} \cdot \sqrt{n}}$$  (2.1)

It was decided to base the initial measurements on a 10 by 10 m sample plot size. From the known planting spacing of 2 m, this plot size would correspond to between 16 and 25 trees per plot. Pilot measurements of nine 0.01 ha plots were collected in the vicinity of the flux tower (see section 2.4.3). Using this data, the accuracy of using a set of 49 plots to determine the canopy diameter at breast height (DBH), $d_t$, characteristics was assessed. The resulting data suggested that the 49 plots would allow the average plot DBH to be determined to about 3% of the site average plot DBH, and the DBH of the population of trees to within about 1.3%.

Although it was possible to lay out up to ten plots per day, establishment of all plots took several weeks because of the infrequency of site visits. All plots were located along planting row transects, (figure 2.5). The end point of each transect was marked at the entrance and exit of each row to allow relocation of the transect at a later time. For the south western side of the site, the plough lines ran either parallel or perpendicular to the layout of the grid so that the distance between plots was measured in 100 m increments along the plough line using a measuring tape. For the north eastern side of the site, the plough lines ran at 45 degree angles to the grid so that the distance between plots was measured in 140 m increments along plough lines.
Figure 2-5 Schematic of sample plot establishment scheme. Squares represent plot location, grey lines the forest track and ride dissecting the site, and the black line the plough lines followed when installing the plots.

At the approximate location, a plot corner was randomly determined by randomly dropping a compass three times. The first reading determined the direction of the plot corner point, the second reading determined the distance (from 0 to 3.6 m) of the plot corner point and the third reading determined the plot corner (e.g. a reading between north and east would indicate placement of the north east corner of the plot). The plot was marked on the ground with plastic twine, which was pegged to the ground at the corners of the plots.

All trees in the newly established plots were tagged with marking ribbon and numbered sequentially; the first tree tagged was also labelled with the plot number. All trees where tagged if they consisted of an independent bole above breast height. Lodged trees were noted but were not tagged or measured. Tree DBH (see section 2.6.1) was measured in all plots and GPS readings were made in some plots at the time of plot establishment, (the GPS readings were later determined to be too inaccurate for identifying plot location).

During the course of the experiment, additional measurements were made in selected plots. A table of these ancillary experiments is contained in appendix C. Selection of plots for additional measurements was based on the distribution of plot DBH sizes in
order to represent adequately the experiment site. In certain circumstances the requirement for power, reasonably rapid access to the plot or accessibility issues determined sample plot usage. However, for ancillary measurements, plots were usually selected so that they fell near the mid point and/or the upper and lower standard deviations of the distribution of DBH. When possible, additional measurements were situated in 'unused' plots. This approach improved marking of the access to the plots and also access to the plot through trimming of the understory branches along the path leading to the plot.

2.4.2 Microclimate measurement sites

Climatological and canopy microclimate variables are determining factors in the exchanges of mass and energy from the site. Any model developed from or applied to these data will require microclimatological information to effectively represent the exchanges of this ecosystem. Important variables associated with the transfer of carbon dioxide and energy include solar and thermal radiation, temperature, humidity, carbon dioxide and wind speed both in and above the canopy. Additionally, precipitation, atmospheric pressure, wind direction, and soil moisture measurements allow parameterisation of the exchange processes.

It was decided to select two primary microclimate measurement locations and instrument plots in the vicinity of those locations. Measurements requiring power and routine maintenance were situated in these locations. The primary microclimate site was situated approximately 70 m north of the intersection of the road and ride at the center of the site, figure 2.4. A tower was installed at this location for making canopy profile and above canopy measurements. A range of average tree sizes existed in the plots surrounding this tower allowing deployment of ground based instrumentation under a variety of canopy cover. A second set of microclimate measurements was located in association with the flux tower placement described in the next section.

2.4.3 Flux measurement site

The flux measurement methods used in this thesis (see chapter 5) allowed for direct determination of the fluxes of mass and energy. However, the expense of these
methods prohibited duplication of measurements, and therefore, only one flux measurement site was selected.

A key factor in the selection of a flux measurement site is the site's homogeneity. Two concepts used in determining representative surface scales of interest are ‘blending height’ (Claussen 1995) and ‘flux footprint’ (Hsieh et al. 2000; Schmid & Oke 1990). Blending height implies that at some height above the surface, atmospheric mixing makes variations in the surface fluxes undetectable. The flux footprint concept is more informative because it quantifies the spatial scales represented in a flux measurement. Footprints may be used to ascertain the
appropriate fetch for a measurement. Here fetch describes the upwind distance of homogeneous surface required to make a representative flux measurement. Recent progress on this topic has seen it grow to cover a wider range of atmospheric and surface conditions (Schmid 2002).

This topic will be employed further in chapters 5 and 8; at this point it will suffice to say that the measured flux will be more sensitive to variations in the surface flux at distances close to the tower (10 to 50 m) and less so at greater distances. From the aerial photo of Griffin, figure 2.6, we see that the variation in surface cover in the vicinity of the flux tower is quite uniform. However under very stable conditions and northerly flow the flux from the rather large 'bare' spot at the intersection of road and ride may represent upwards of 15% of the flux. Under unstable and near neutral conditions it will only contribute on the order of 1 to 2% of the flux, (Based on integration of footprints calculated following Fan et. al. (1992).

Site topography must also be considered. While small or gradual changes in topography may not have significant impact on the measured flux, in some situations, topography may play an important role in determining the flux at a specific location in the landscape (Finnigan & Brunet 1995). As our interest was the determination of ecosystem flux and not the effects of topography, any potential landscape effects, were avoided as far as possible. The topographic features of the site can be represented as a contour map of site slope, as shown in figure 2.7. The slope in the vicinity of the flux tower appears to be relatively uniform with a slope of approximately 5 degrees. However, in the southern and south western directions, there are rapid increases in slope beyond 500 m radius from the tower. However, the predominant wind flow along the valley axis makes the slopes to the southwest less influential. If it were not for the spatial inhomogeneity to the north and northeast of the tower it may be recommended, for any future studies, that the tower be moved further in those directions.
2.5 Site management

2.5.1 Land use

Since 1980 the experiment site has been a commercial forestry plantation, planted and managed on behalf of HSBC by Tillhill Economic Forestry Ltd. The site is also used for hunting and fishing expeditions, which provides an aspect of wildlife management that is necessary for plantation development, and provides additional income to Tillhill. In addition, the site’s forest roads provide a link for hikers and bicyclists. No problems of vandalism or theft were encountered during the course of the experiment.

2.5.2 Site preparation and planting

For previously unforested peaty gley soils, such as at Griffin, it is common to plough the site and plant seedlings on the overturned soil. This practice provides an aerated
soil that is initially weed free that improves root development of the seedlings. The ploughing pattern at Griffin appears to have been planned to maximize the slope along the plough lines to improve runoff. This resulting slope will have exceeded the optimum slope of 2 degrees beyond which erosion is encouraged (Forestry Commission 1991). Sample plot transects, (figure 2.8), show typical topographic cross sections perpendicular to the plough line. A microtopographic variation in surface height of 30 to 60 cm is observed at this site.

The plantation was established during the years 1980 and 1981. Trees were planted at a spacing of 2 m, equivalent to a planting density of 2500 while the surveyed planting density was slightly lower at 2215 trees ha\(^{-1}\). The site was planted in blocks containing Sitka spruce (*Picea sitchensis*), Douglas fir (*Pseudotsuga menziesii*), and Japanese larch (*Larix kaempferi*). A survey of the experiment site revealed the overwhelming predominance of *Picea sitchensis* (table 2.3). No thinning had been performed on the forest before or during this experiment. However, it is planned to thin the experiment site in 2004 as a continuation of this experiment.

![Plot transects showing example of microtopographic variations of plots 51 and 33. Heights were measured relative to the initial measurement point of the transect.](image)

**Figure 2-8** Plot transects showing example of microtopographic variations of plots 51 and 33. Heights were measured relative to the initial measurement point of the transect.

**Table 2.3** Measured species composition for Griffin.

<table>
<thead>
<tr>
<th>Species</th>
<th>Measured Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Picea sitchensis</em></td>
<td>97.3</td>
</tr>
<tr>
<td><em>Pseudotsuga menziesii</em></td>
<td>2.1</td>
</tr>
<tr>
<td><em>Betula</em> spp.</td>
<td>0.3</td>
</tr>
<tr>
<td><em>Salix</em> spp.</td>
<td>0.3</td>
</tr>
</tbody>
</table>
2.5.3 Fertilization

Fertilization of Sitka spruce plantations is often necessary to improve the competition of the spruce against heather (*Calluna vulgaris*) and other dominant surface vegetation during early years. For Griffin, applications of nitrogen, phosphorous, and potassium by helicopter since 1989 are shown in figure 2.9. In the region of the flux tower the total application of fertilizer was approximately 350 kg ha\(^{-1}\), which was applied in the year 1996. Some regions in the experiment plot have had no applications since 1989 while other regions with suspected heather check received repeated treatments with total applications of up to 1400 kg ha\(^{-1}\).

![Fertilization map of the Griffin experiment site. Applications from the years 1990 through 1999 are included. (Data from Tillhill Economic forestry Ltd.)](image)

2.6 Canopy description

Descriptions of a forest canopy for the purposes of a scientific experiment may be quite detailed, providing information on canopy structure and biomass. Resources were not available to carry out such a detailed study of the Griffin canopy, but the
more easily acquired canopy characteristics were measured and used with previous
determined allometric relationships to estimate required values (Wang et al. 1991).
Primary measurements included: tree diameter at breast height, canopy height, leaf
litter, and leaf area index (as part of another research project (Lisa Wingate 2003)).
From this information, canopy biomass content and change was determined from
allometric relationships. These measurements and relationships are described in the
following sections.

2.6.1 **DBH, basal area, and biomass**

The growth of a forest is usually estimated using allometric relationships based on
easily measurable tree characteristics. A tree's diameter at breast height, and canopy
height are two measures commonly used to estimate the amount of standing stock in a
forest.

Measurements of DBH were taken using rounded down diameter tapes. The diameter
of each tree in the plots described in section 2.4.1 was measured. Although
measurements of DBH were conducted annually, two main measurement campaigns
were conducted (in 1997 and 2001).

During the 1997 campaign, the trees were measured at the same time the plot was
established. Once the plot boundaries were marked, each tree was sequentially
numbered with tagging tape and its DBH measured. During this campaign three
people made the measurements, taking the tasks of tagging, measuring and recording.
When tagging trees, it was found most useful to work systematically along rows
within the plot. This practise made re-identification easier in subsequent years in
cases where a tree tag was missing or its number obscured. Because some of the tree
tags were prone to more rapid deterioration, some of the more commonly used plots
were retagged with aluminum markers in 1998.

During the measurement campaign in 2001 approximately 26 of the initial 47 plots
were identified and measured. Many plots remained unmeasured because of the
difficulty in relocating the plots.
When measuring DBH, effort was made to measure consistently at similar heights. Difficulty arose because of the roughness of the surface, which had height variations of up to 1 m or more (see section 2.5.2). If branches occurred at the level at which the DBH measurement was to be made the closest position on the bole which was unobstructed by branches was used to determine the tree's DBH. Forked trees were measured independently if the fork occurred below the level of measurement. All trees were measured and no lower limit of DBH was imposed on the recorded values.

In addition to the direct measurements of DBH, 86 cores were taken from 54 trees and historical estimates of DBH were estimated from these cores. Cores were extracted and wrapped in plastic to minimize shrinkage. In the laboratory the cores' annual ring widths were measured and summed to determine annual DBH values.

As an intermediate step in determining forest biomass the inventory DBH, \(d_n\) values were converted to basal area, \(B_a\) (\(m^2\) ha\(^{-1}\)) by summing the basal areas of trees within a plot, \(m\), adjusting the units and then averaging all plots measured during the year, \(n\), to obtain the stand average basal area:

\[
B_a = \frac{\sum_n \left( \frac{\pi}{4} d_i^2 \right)}{n}.
\]  

Using the values of \(B_a\), the amount of above ground biomass \(W_a\) (Mg ha\(^{-1}\)) was estimated using the empirical relationship (Wang et al. 1991):

\[
W_a = 5.90025 + 2.30985 \cdot B_a + 0.0218 \cdot B_a^2.
\]  

This relationship was developed for Sitka spruce stands aged between 13 and 22 years, yield class 10 - 28 and stocking densities of 2520 to 4160 trees ha\(^{-1}\). Griffin fits within the first two ranges specified though the forest’s stocking density was at the low end of the densities used in deriving the relationship in equation (2.3).
Figure 2.10 shows the distributions of DBH values for matching trees from the 1997 and 2001 inventory surveys. They exhibit a shift in the mean DBH from 9.9 to 12.5 cm, (the mean of all measured DBH values in 1997 was 9.3 cm). This increase corresponds to an average annual DBH increase of 0.65 cm y$^{-1}$.

Figure 2-10 Frequency distributions of DBH for the 1997 and 2001 sampling campaigns. For 1997 both the compete data set and the samples set which correspond to the 2001 sampling are presented.

Figure 2-11 Four year increases in plot aboveground biomass compared to initial (1997) biomass.
Comparing the increase in biomass, equation (2.3), from 1997 to 2001 against the 1997 biomass, we see the largest increases in biomass for plots with 1997 biomass values larger than 50 Mg ha\(^{-1}\), figure 2.11. This pattern is also observable in the increases of individual tree DBH sizes. Closer examination revealed that both forked and non-forked trees exhibited this pattern. However, there was insufficient data to draw conclusions about species other than *Picea sitchensis*. The non-linearity of this pattern likely represents the effect of tree LAI reduction with canopy closure and the resulting reduction in production capacity (Ryan *et al.* 1997) and is unlikely to be related to changes in hydrologic or nutrient limitations that may affect more mature stands (Gower *et al.* 1996; Hunt *et al.* 1999; Ryan *et al.* 1997). Back calculation using equations 2.2 and 2.3 suggests that the peak of the curve in figure 2.11 corresponds to a DBH value of 13.5 cm.

![Figure 2-12 Mean annual biomass estimates from tree core and circumference tape measurements. Tree core biomass estimates for hillside, experiment site and combined data are shown. Error bars represent ± one standard error (n= 4 to 47).](image)

Estimates of stand biomass calculated from tree cores, as part of a student project assessing topographic effects, (figure 2.12), are lower than those estimated from direct measurements of DBH. The smaller mean values may result from the inclusion of cores taken from trees on steeper slopes outside the immediate experimental plot. It is apparent from figure 2.12 that the biomass estimates taken from tree cores within the experiment plot are similar to those obtained using circumference tape determined
estimates of biomass. While the purpose of figure 2.12 is to show the adherence of the chronological change in biomass at Griffin to expected growth curves (Forestry Commission 1991; Rennolls 1995), the observed differences in tree core data indicate that site measurements may not be valid over larger areas.

The spatial distribution of biomass shown in figure 2.13, was interpolated from plot biomass data. The figure suggests a larger forest biomass in the south central region of the experiment site, though a finer resolution sampling would be required to identify if any of the existing plots represented anomalously high or low values. The larger values to the north east of the track may be associated with a stream in that vicinity and the associated possibility of thicker soils and higher soil moisture.

![Figure 2-13 Spatial distribution of above ground biomass.](image)

The annual average above-ground plot biomass values are given in table 2.4. Also given are the equivalent carbon values assuming a 48% carbon content (Lamlom & Savidge 2003) (see also section 2.6.4.3). These biomass estimates are compared with the tree core estimates in figure 2.12 and show a similar pattern of increase in biomass over the experiment period. As mentioned previously the smaller tree core biomass estimates may have been the result of including cores taken from trees on slopes outside the experimental plot.
Table 2.4 Measured values of DBH and derived values of basal area, biomass and carbon content evaluated for the Griffin forest experiment site.

<table>
<thead>
<tr>
<th>Year</th>
<th>DBH cm</th>
<th>Basal Area m² ha⁻¹</th>
<th>Biomass Mg ha⁻¹</th>
<th>Carbon g m⁻²</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>9.9 ± 4.4</td>
<td>20.3 ± 11.4</td>
<td>61.7 ±39/-34</td>
<td>2964</td>
</tr>
<tr>
<td>1998</td>
<td>10.4 ± 4.8</td>
<td>22.0 ± 12.1</td>
<td>67.2 ±43/-36</td>
<td>3224</td>
</tr>
<tr>
<td>1999</td>
<td>11.3 ± 4.0</td>
<td>23.8 ± 11.0</td>
<td>73.3 ±39/-34</td>
<td>3519</td>
</tr>
<tr>
<td>2000</td>
<td>10.7 ± 3.6</td>
<td>26.4 ± 11.2</td>
<td>82.2 ±42/-36</td>
<td>3944</td>
</tr>
<tr>
<td>2001</td>
<td>12.4 ± 5.1</td>
<td>31.2 ± 13.9</td>
<td>99.0 ±55/-47</td>
<td>4754</td>
</tr>
</tbody>
</table>

From this data, the annual above ground carbon accumulation of the site was estimated to be 448 g C m⁻². This value does not include biomass accumulation for below ground components or accumulation by surface vegetation.

2.6.2 **Height**

Canopy height, \( h_c \) was measured to provide information for growth modelling. It also provided input on estimating zero plane displacement and roughness length. Canopy height was measured in 1997 and 2001; however, the spatial representation of canopy height measurements was limited because of the difficulty in making these measurements in a closed canopy.

![Figure 2-14 Schematic of tree height measurement methods.](image-url)
For the first set of tree height measurements taken in 1997, a height pole was used to obtain the height of the canopy, figure 2.14. The errors associated with the use of the height pole would have been restricted to parallax associated with observing the top of the tree and pole and with errors in the height measurement of the pole itself.

Measurements of canopy heights in 2001 employed a different strategy because most trees were too tall to measure using a height pole; instead, trigonometric methods were employed. The height of the tree was determined from the inclination angle of the top of the tree as measured from a known height on the tower, $\alpha_c$, and the horizontal distance from the tower to the tree, $x$. The value of $h_c$ was calculated as:

$$h_c = \frac{\sum (z - x \cdot \tan(\alpha_c))}{n}$$ (2.4)

The trigonometric method assumed that the ground was horizontal between the tower and the tree. This assumption will have associated systematic errors of approximately $\pm 1$ m, depending upon the slope in the vicinity of the tower. The site microtopography will not have had as significant an effect as its magnitude would imply since trees were consistently planted on the tops of ridges. The slope-associated error would have been azimuthally dependent. Because measurements were taken of trees from many azimuths, it was assumed that this error would have cancelled out. Other errors in this method will have resulted from random errors in the estimation of the distances of $x$ and $z$ and of the angle $\alpha_c$.

During the first set of tree height measurements the mean canopy height was 6.7 m, with a maximum observed height of 10.2 m and a coefficient of variation of 26%, as shown in figure 2.15. In 2001, the mean and maximum had increased to 10.5 m and 15.8 m, respectively and the CV had decreased to 20%. Despite a known non-linear relationship of $h_c$ to DBH over the life of a forest (Forestry Commission 1991), over the period of this experiment a linear relationship could be used to relate $h_c$ to $d_t$.

$$h_c = 2.44 + 0.55 \cdot d_t$$ (2.5)
2.6.3 Canopy dimensions

Canopy base height and canopy radius were measured in 2001 to provide information for growth modelling. The values of canopy radius and base height also provided a quantitative measure of the occurrence of canopy closure.

Canopy radius was measured as the maximum horizontal extent of any single branch on a tree. On smaller trees and in open canopy this task was straightforward. In areas of closed canopy the intermingling of the branches and height of the lowest live branches made identification of the longest live branch more difficult. It is subjectively estimated that the error in radius measurements is on the order of ± 0.5 m for closed canopies, but probably half that for open canopies. Canopy base height was measured as the lowest level of live leaves on a branch. Again similar difficulties arose when evaluating this parameter for closed canopies with elevated bases. When determining radius and base height, maximum and minimum values were used and no account was taken of tree asymmetry resulting from preferential tree growth due to openings in the canopy adjacent to the tree being measured. This will have the effect of decreasing the base height and increasing the radius of larger trees in canopies that are not completely closed.
For open canopies it is expected that the canopy base height will remain near the surface. Only when the canopy closes will the canopy base begin to rise as the lower needles, deprived of sunlight, weaken and die, as was observed for Griffin, figure 2.16. The data suggest that closure occurs and the maximum canopy base height is reached at a canopy height of 10 to 11 m. Closure of the canopy for tree heights greater than 10.5 m suggests that canopy radius will have reached its maximum for trees of that height. The average maximum canopy radius for trees with \( h_c > 10.5 \) m was 1.9 m, suggesting a 1 m overlap of branches in a closed canopy, figure 2.16.

### 2.6.4 Litter fall

Litter fall was measured in four plots. In each plot, four litter trays (labelled A through D) were randomly placed using pre-selected random distances ranging between 0 and 10 m north and west of the southeast corner of the plot. Two types of litter trays were used. The first type was employed in two plots, (plots 25 and 51), and consisted of a 50 cm by 50 cm square wooden frame over which fabric was stretched and stapled. This frame was laid upon the ground at the specified random location. In the other plots the traps consisted of a cloth sample bag attached to a bicycle wheel rim. These traps were elevated approximately 40 cm above the ground.
Samples were collected monthly, unless weather conditions rendered the samples impossible to collect accurately. Such situations arose under conditions of heavy snow or persistent heavy rain. Under these circumstances the samples were left for a further month before being collected.

Collected litter samples were air dried in the laboratory for one to three weeks until they were easily removable from their collection containers or until time was available for sample drying. The samples were then transferred to pre-weighed aluminium trays and dried in an 80°C drying oven. Samples typically remained in the oven for three days but may have remained longer if there was insufficient time to reweigh the samples after 3 days.

During the first year of sample collection, the litter fall samples were further separated into sub samples of leaves, twigs, cones, and other materials. The separated samples were weighed independently to obtain their component weights. In addition a single set of litter samples from July-August 2000 were dried, separated and weighed to obtain the mass of the litter components. These components were then ground in a ball mill, sub-sampled and analysed to determine their carbon, nitrogen, and hydrogen content.

2.6.4.1 Trap type and plot comparisons

Because two litter-trap configurations were used, the possibility of differences in trap collection efficiency existed. In one plot (51), trap type was changed for two of four traps, after seven months. To check for differences, a Mann-Whitney rank sum test was used to compare the ratios of litter collected in the changed traps to litter collected in all other traps and as a ratio with the other two traps in the plot.

The test results indicate a significant difference (p<0.001) when the comparison was based on the ratio using all available traps but were insignificant when using only the traps within the plot. The significant difference observed when using all plots probably resulted from changes in litterfall for two of the plots coincident with the timing of the change in plot traps on 13/7/1998, as can be seen in figure 2.17. Therefore, the statistical results of the within plot comparison was considered more valid and that no difference between traps types could be identified.
Unlike deciduous species and some coniferous species, the litter fall of *Picea sitchensis* does not follow regular annual patterns, often correlating with unusual conditions of wind, drought or pestilence (Owen 1954; Pedersen & Bille-Hansen 1999). The time series of monthly litter collections (figure 2.18) exhibits an irregular pattern of litter fall with strong peaks in the autumn of 1998 and 2000. No obvious correlations to climatic conditions were discovered. However, a common cause of large litter falls in Sitka is the occurrence of aphids, (Straw *et al*. 1998; Straw *et al*. 2000). While this may be the cause of the observed peaks no record was kept of insect activity with which the litter fall records could be correlated.

A strong correlation that was apparent in the litter fall data was its relation to canopy development. The effect can be clearly seen in the pattern of cumulative litter fall, (figure 2.17). It is apparent that, after 1998, the litter collected in plots 16 and 25 was accumulating at nearly twice the rate of that collected in plots 15 and 51. The most apparent difference between these plots is their different average DBH values. Table 2.5 give the DBH values for these four plots for the years 1997 and 2001. From this information we see that the tree size in plots 15 and 51 in the year 2001 were not yet as large as the tree sizes of plots 16 and 25 in the year 1997.
If we refer back to figures 2.16 and 2.15 we can infer that the canopy radius reached its maximum size, and probably canopy closure, at plot DBH values of about 13. The data in table 2.5 indicate that the canopy in plots 15 and 51 had not yet closed in 2001 while the canopies in plots 25 and 16 would have closed in, or slightly before, 1997. Based on these findings, the rate of litter fall for unclosed canopy may be described by plots 15 and 51 and that for a closed canopy by plots 16 and 25.

### Litter fall components

During the first year of collection, litter was separated into its component parts. The results, (figures 2.19 and 2.20), indicate that needles are the primary component of litter fall throughout the year. However, there is an increase in the contribution of sticks and cones from a few percent of the litter in the summer to more than 20% in the winter. A possible cause for this annual variation in the components of litterfall may be the annual variation in wind speeds, (see chapter 4). At least for 1998, the
increased proportion of sticks and cones in the litter corresponds with higher wind speeds.

![Figure 2-19 Litter components for the year 1998](image)

Unfortunately, separation into components was only done for samples collected between February 1998 and January 1999, an unusual period in terms of litter fall because of the large litterfall rates observed in plots 16 and 25 as a result of canopy closure. The results separated by plot and shown in figure 2.20, are ambiguous in that plot 51 is similar to plots 16 and 25 in its proportion of leaf litter while plot 15 contains a smaller proportion of leaf litter. Analysis of other years are needed to reveal if the large peaks observed in 1998 and 2000 are the result of increased leaf
loss or an overall increase in all litter components and if there is a change in the components of litterfall as a result of canopy closure.

2.6.4.3 Litter nutrient contents

The nutrient content analysis of the single sample from July-August 2000, (figure 2.21), suggest that the woody component of litter contains the highest carbon content, 49.2%, while the unidentified litter contains the lowest carbon content 47.8%. In contrast, the unidentified litter contains the highest nitrogen content 1.6%, while cones contain the lowest nitrogen content, 0.3%.

A possible reason for the higher nitrogen content of the ‘Other’ material may be caused by high nitrogen content of seeds contained within cones (Zackrisson et al. 1999) being lost to the ‘Other’ component during the drying process. It is therefore recommend to air dry and separate components before further drying in future determinations. The reasons for the low carbon content of plot 15 and the low nitrogen content of plot 51 are not known. Again, more extensive analysis is needed to identify patterns of nutrient content of litterfall.

![Figure 2-21 Litter carbon and nitrogen contents expressed as percent of mass. Graph on left shows the contents by component and figure on the right shows contents by sample plot. Error bars represent one standard deviation.](image-url)
2.6.4.4 Litter contribution

An annual litter carbon contribution to the soil of 85 g (C) m\(^{-2}\) a\(^{-1}\) for open canopy plots and 290 g (C) m\(^{-2}\) a\(^{-1}\) for closed canopy plots was calculated from the cumulative litter fall values shown in figure 2.17 and the approximate 48% carbon content of litter. Employing an exponential model similar to Titus and Malcolm (1999), gave a decay constant of approximately 0.55, similar to the coefficient range of 0.05 to 0.80 observed by Titus and Malcolm (1999). This similarity suggests that their estimate of the long term (~30 years) retention of 16% of litter carbon is also appropriate for this site.

2.6.5 Leaf area

Leaf area measurements were obtained from results of an ancillary experiment carried out by PhD student Lisa Wingate in 1998. Both hemispherical photos, and direct sampling methods were employed. Results of this analysis suggest a 1998 LAI of about 6, which could be projected forward to a leaf area of approximately 8 by 2001.

2.7 Soil description

The management (ploughing) of the soil at Griffin during afforestation, mentioned in section 2.5.2, created both complications and opportunities when attempting to describe the characteristics of the soil. The ploughing of the soil resulted in a horizontal micro-topographic variability on the scale of the planting density, \textit{i.e.} 2m. The resulting soil surface was characterised by plough ridges, furrows and unploughed areas. Although this required greater sampling to characterize the soil, the properties of the overturned soil on the plough ridges created an opportunity to estimate carbon and nitrogen accumulation in the soil surface since the time of planting. A representation of the soil properties as a result of ploughing is presented in figure 2.22. This figure shows a schematic representation of the soil and associated density and nutrient content of the soil prior to planting, immediately after afforestation, and 20 years after afforestation. It is the difference of the nutrient content profiles of the ridges immediately after and 20 years after afforestation that has allowed estimation of soil carbon and nitrogen accumulation at this site over the corresponding period.
2.7.1 Structure

Soil structure was determined from inspection of both soil pits and soil cores. Soil pits were dug, in the vicinity of the eddy covariance and scaffold towers, to a depth lower than the level of the surrounding plough furrows. Each pit intersected both a plough ridge and trough, but not the unploughed area of the forest floor. Although soil cores were taken for the determination of other variables, they did provide soil structure information from a wider spatial distribution of locations, albeit from not as great a depth.

From the first soil identification done near the flux tower in 1996, the soil was classified as a brown earth. A second classification in 2000 identified the soil as a stagnohumic gley of the Strichen soil association. Table 2.6 contains an excerpt of the 2000 classification, done by Lisa Wingate; it describes the layer depths and characteristics. Further work needs to be done to determine if the observed soil classification differences are the result of spatial differences or are the result of misclassification.
Table 2.6 Description of Stagnohumic gley soil characteristics.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth</th>
<th>Characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>LF</td>
<td>0 – 2 cm</td>
<td>dark/light brown; no mineral grains; very fibrous; moist; layers of fibrous roots; lots of fine roots and needles; no stones; diffuse boundary.</td>
</tr>
<tr>
<td>O&lt;sub&gt;1&lt;/sub&gt;</td>
<td>2 – 21 cm</td>
<td>Dark brown; no identifiable mineral grains; amorphous; moist; many fine and course roots; crumbly; no stones; no sharp boundary.</td>
</tr>
<tr>
<td>O&lt;sub&gt;2&lt;/sub&gt;</td>
<td>21 – 30 cm</td>
<td>brown almost greyish/dark brown; amorphous; moist; no mineral grains; blocky structure; many fibrous and coarse roots; no stones; wavy gradual boundary.</td>
</tr>
<tr>
<td>Bg&lt;sub&gt;1&lt;/sub&gt;</td>
<td>30 – 40 cm</td>
<td>light brownish/yellowish/reddish (mottled); silty loam; fine distinct mottles; moist; not friable; few fine fibrous roots; no stones; indistinct boundary.</td>
</tr>
<tr>
<td>Bg&lt;sub&gt;2&lt;/sub&gt;</td>
<td>40 – 55 cm</td>
<td>greyish/orange-reddish mottles; iron pyrite flakes; friable; abundant (10-20 cm) mottles; moist; blocky; occasional fine roots; stones large; coarse sandy loam; horizon continues.</td>
</tr>
</tbody>
</table>

2.7.2 Bulk density

Soil bulk density is required for calculations of total nutrient content and soil heat storage. Its value was determined from soil core samples, either specifically for that purpose or in association with other measurements.

In 1998, soil core samples were taken in plots 18, 33 and 51. These cores were taken in an attempt to determine a relationship between surface topography (unploughed, ridge, or furrow) and soil bulk density. Cores were taken with a soil corer every 0.3 m along a transect crossing the plot perpendicular to the plough lines. The samples were extracted and the depth of the extraction hole was measured. The cores were generally less than 14 cm and no attempt was made to split the cores into different sections. The samples were placed in plastic bags and their wet weight was obtained. The samples were then dried and their dry weight measured. Bulk density was calculated from the dry soil weight, corer diameter and core depth. Soil gravimetric water content was also determined from the sample dry and wet weights. The results for individual plots from the 1998 sample set are presented in table 2.7.
### Table 2.7 Comparison of soil bulk density samples.

<table>
<thead>
<tr>
<th>Samples</th>
<th>Surface</th>
<th>Year</th>
<th>Average Soil bulk density g cm⁻³</th>
<th>Average Core depth cm</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plot 18</td>
<td>All</td>
<td>1998</td>
<td>0.52 ± 0.16</td>
<td>11.8 ± 4.8</td>
<td>33</td>
</tr>
<tr>
<td>Plot 33</td>
<td>All</td>
<td>1998</td>
<td>0.44 ± 0.17</td>
<td>14.7 ± 3.6</td>
<td>31</td>
</tr>
<tr>
<td>Plot 51</td>
<td>All</td>
<td>1998</td>
<td>0.55 ± 0.36</td>
<td>15.6 ± 5.2</td>
<td>28</td>
</tr>
<tr>
<td>Closed Canopy</td>
<td>Unploughed</td>
<td>2001</td>
<td>0.53 ± 0.23</td>
<td>0 - 15.0</td>
<td>22</td>
</tr>
<tr>
<td>Open Canopy</td>
<td>Unploughed</td>
<td>2001</td>
<td>0.50 ± 0.52</td>
<td>0 - 15.0</td>
<td>10</td>
</tr>
<tr>
<td>Closed Canopy</td>
<td>Ridge</td>
<td>2001</td>
<td>0.93 ± 0.52</td>
<td>0 - 15.0</td>
<td>22</td>
</tr>
<tr>
<td>Open Canopy</td>
<td>Ridge</td>
<td>2001</td>
<td>0.89 ± 0.46</td>
<td>0 - 15.0</td>
<td>21</td>
</tr>
<tr>
<td>Closed Canopy</td>
<td>Unploughed</td>
<td>2001</td>
<td>0.98 ± 0.43</td>
<td>15.0 - 28.0</td>
<td>18</td>
</tr>
<tr>
<td>Open Canopy</td>
<td>Unploughed</td>
<td>2001</td>
<td>1.30 ± 0.48</td>
<td>15.0 - 33.0</td>
<td>26</td>
</tr>
<tr>
<td>Closed Canopy</td>
<td>Ridge</td>
<td>2001</td>
<td>0.88 ± 0.30</td>
<td>15.0 - 47.0</td>
<td>24</td>
</tr>
<tr>
<td>Open Canopy</td>
<td>Ridge</td>
<td>2001</td>
<td>0.78 ± 0.39</td>
<td>15.0 - 47.0</td>
<td>24</td>
</tr>
</tbody>
</table>

In a second set of measurements done by Franz Conen in 2001, soil bulk density was assessed for 16 soil cores. Eight cores were taken in two plots with both open and closed canopies to see if the presence of ground vegetation or differing microclimate would have affected the soil characteristics. Of the eight cores taken in each plot, four were taken from soil undisturbed by ploughing, and four were taken from the tops of ridges. The extracted cores were split into 5 cm layers. The core sections were oven dried and estimates of bulk density calculated from the segment length.

When analysed in their entirety, the 2001 cores do not exhibit mean bulk densities significantly different than the 1998 cores (p=0.19). However, when the 2001 cores were separated by core segment depths less or greater than 15 cm, the 1998 cores and the 0-15 cm undisturbed cores were significantly different than the inverted cores and the lower end of the undisturbed cores. This result simply indicates a high mineral content to the ridges. Plotting the 2001 cores against depth, figure 2.23, shows the character of the inverted and undisturbed cores. We see in these inverted cores that the increase of bulk density with height above the previous surface level is similar to the increase in bulk density with depth below the undisturbed surface level for both the undisturbed and inverted cores. Note also the thin layer of low bulk density at the
top of the inverted core profiles, indicating the accumulation of organic matter on the surface of the inverted cores. More comprehensive sampling would be required to discern spatial variation of soil characteristics at the site.

![Graph showing bulk density profiles for open and closed canopy locations](image)

**Figure 2-23** Bulk density profiles for open canopy (circles) and closed canopy (triangles) locations. Two samples each from undisturbed (black) and ploughing ridges (hollow) are presented for each location.

### 2.7.3 Nutrient content

The primary purpose of measuring the carbon and nitrogen contents of Griffin soil cores was to determine if the amount of accumulated soil carbon could be determined from the ridges of inverted soil. Soil content of carbon and nitrogen was assessed for the 16 soil cores used in the bulk density determinations. To determine the carbon and nitrogen content of the soil, the dried samples were sieved to remove roots and mineral matter greater than 2 mm. The remaining material was ground using a ball
grinder. These samples were then sub sampled and processed with a C/N analyser to determine their carbon and nitrogen percentages. The carbon and nitrogen contents were determined by multiplying the percentage contents by the corresponding sample bulk density.

Figure 2-24 Total soil carbon content expressed as percent of mass for open canopy (circles) and closed canopy (triangles) locations. Two samples each from undisturbed (black) and ploughing ridges (hollow) are presented for each location.

In both unploughed and ridge soil cores we see increases in the content of both carbon (figure 2.24) and nitrogen (figure 2.25) in the top 10 cm of the soil cores, reflecting nutrient inputs by vegetative detritus. In the undisturbed cores, the high near-surface values of C and N gradually decline with depth. Because no historical information about the undisturbed cores exists, it is impossible to determine the additions of C and N to the soil from the forest. However, below the top 10 cm in the inverted cores, the content of C and N increase to a maximum at a depth corresponding to the level of the undisturbed surface and then decrease again. This structure reflects the inversion of the soil resulting from ploughing during site preparation.
This pattern allows the determination of additions of carbon and nitrogen to the soil surface since the time of ploughing. It is assumed that immediately after ploughing the C and N content of the ridge cores would be lowest at the new surface and highest at the previous, buried, surface. We may then use the inflection in the C and N content profiles occurring at about 5 cm below the surface of the ridge core profiles to identify the depth of the layer of discernable additions of C and N to the upturned soil. The amount of added C and N was determined by subtracting the expected amount of C and N in the soil based on the decline with depth in the undisturbed profiles, (confer figure 2.22).

The amount of carbon and nitrogen added to the soil since time of planting, as obtained by this method are given in table 2.8. Assuming constant rates of litter input
since the time of planting, the values in table 2.8 would imply an annual soil carbon contribution of approximately 111 g (C) m\(^{-2}\) a\(^{-1}\). It is realized that this approach is only an approximation because the variations in litter contributions over the lifetime of the forest and the effect disturbance of the soil will have on soil temperature and soil moisture. Such variations may have affected leaching, and respiration of nutrients contained within the inverted soil cores as compared to an undisturbed soil. However, this approach has provided a best guess estimate of soil carbon accumulation that would have been difficult to estimate through sampling of only undisturbed soil cores.

Table 2.8 Estimates of soil carbon and nitrogen increases for Griffin forest since 1980.

<table>
<thead>
<tr>
<th>Location</th>
<th>Carbon added g m(^{-2})</th>
<th>Nitrogen Added g m(^{-2})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Open canopy</td>
<td>2451 ± 263</td>
<td>66 ± 97</td>
</tr>
<tr>
<td>Closed canopy</td>
<td>2230 ± 741</td>
<td>45 ± 103</td>
</tr>
</tbody>
</table>

2.8 Summary

Description and review of experiment layout provided suggestions on possible modifications to tower placement, allometric measurement methods, power supply, and requirements for greater spatial extent of measurements of soil characteristics and aboveground biomass. Relationships of canopy allometry (except DBH) and litter fall were determined and found to have a strong dependence upon canopy closure. Soil carbon content accumulation was determined by exploiting the site preparation practise of ploughing.

The estimates of above-ground and soil carbon accumulations were used to estimate the total site carbon accumulation as shown in table 2.9. However, because no estimates of below-ground biomass carbon accumulation was made, it was estimated to be 25% of above-ground accumulation (ref). Using this assumption the average annual accumulation for Griffin over the period of this experiment was estimated to be approximately 654 g m\(^{-2}\).
Table 2.9 Estimates of average annual carbon accumulation for Griffin forest for the period 1997 through 2001.

<table>
<thead>
<tr>
<th>Carbon accumulation</th>
<th>g m⁻²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Above-ground carbon accumulation</td>
<td>448</td>
</tr>
<tr>
<td>Below-ground carbon accumulation (assumed 20% of above-ground)</td>
<td>90</td>
</tr>
<tr>
<td>Soil carbon accumulation</td>
<td>111</td>
</tr>
<tr>
<td>Estimated total carbon accumulation</td>
<td>654</td>
</tr>
</tbody>
</table>
3 Canopy microclimate measurement techniques

This chapter presents methods general to the measurement of climate variables at Griffin. The site is described in terms of climatological measurements, and the methods used in gathering and maintaining the climate data set. The specifics of climate variables measurements are presented in Appendix A while a statistical presentation and characterization of climate measurements will be presented in chapter 4.

3.1 Sampling of climatological variables

As discussed in chapter 2, statistical characterization of variability at Griffin may require extensive measurements. Again we employ equation 2.1, previously used to define the number of sample plots, to quantify the required number of microclimatological measurements. Using one month’s data, spatial variability was estimated as the median of the standard deviation of values from available instruments making similar measurements, table 3.1. For measurements that did not vary spatially (e.g. incoming solar radiation on a clear day), a single measurement using a properly calibrated instrument would suffice to characterize the variable. For measurements of spatially varying conditions (e.g. within the canopy), numerous samples were required to achieve the desired accuracy. From table 3.1 we observe a high number of samples required for soil heat flux, transmitted PPFD, and net radiation, and a very high number of samples required for wind speed, reflected PPFD, and soil moisture. The high sample values for net radiation are likely to be the result of calibration differences between sensors (see appendix A.1.6) while those for wind speed, reflected PPFD, and soil moisture probably stem from both sensor calibration inaccuracies and insufficient spatial sampling of pilot measurements used in determining the required sample size. The results in table 3.1 can be used as a guideline for sampling requirements in future experiments under similar conditions as those experienced at the Griffin site.
Table 3.1 Estimation of spatial samples required to provide the desired accuracy for various signals. Spatial variability was determined as the median standard deviation for a subset of data using the given number of measured samples. Desired accuracy is specified as approximately 10% of annual range of average diel values.

<table>
<thead>
<tr>
<th>Signal</th>
<th>&quot;Spatial&quot; variability</th>
<th>Desired accuracy</th>
<th>Required samples</th>
<th>Measured samples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil heat flux</td>
<td>0.733</td>
<td>1 W m(^{-2})</td>
<td>16</td>
<td>7</td>
</tr>
<tr>
<td>Soil temperature</td>
<td>0.284</td>
<td>1.2 °C</td>
<td>2</td>
<td>6</td>
</tr>
<tr>
<td>Transmitted PAR</td>
<td>1.818</td>
<td>1 (\mu)mol m(^{-2}) s(^{-1})</td>
<td>72</td>
<td>24</td>
</tr>
<tr>
<td>Air temperature</td>
<td>0.243</td>
<td>1.5 °C</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>Net Radiation</td>
<td>5.174</td>
<td>12 W m(^{-2})</td>
<td>14</td>
<td>3</td>
</tr>
<tr>
<td>Global short-wave</td>
<td>1.713</td>
<td>20 W m(^{-2})</td>
<td>4</td>
<td>2</td>
</tr>
<tr>
<td>Precipitation</td>
<td>0.0034</td>
<td>0.03 mm</td>
<td>7</td>
<td>2</td>
</tr>
<tr>
<td>Wind speed</td>
<td>0.348</td>
<td>0.5 m s(^{-1})</td>
<td>227</td>
<td>2</td>
</tr>
<tr>
<td>Reflected PAR</td>
<td>1.016</td>
<td>2 (\mu)mol m(^{-2}) s(^{-1})</td>
<td>121</td>
<td>2</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>0.262</td>
<td>0.05</td>
<td>12819</td>
<td>2</td>
</tr>
</tbody>
</table>

When duplicate measurements were made, practical and resource constraints imposed limits on the number of replicates collected. Therefore, duplicate measurements were typically taken, not for the purpose of assessment of site variability, but instead to maximize data capture and improve signal quality assessment. Duplicate measurements were usually recorded on separate recording devices (loggers) to maximize the likelihood of data capture in the case of logger failure. When making subjective quality assessment, duplicate measurements allowed a more accurate assessment of quality through comparison of one or more identical or similar sensors. A more detailed discussion of this topic is covered in section 3.5.

The temporal variations of microclimatological signals occur at all time scales. Short time scales between 0.1 second and 1 hour are considered important primarily for the determination of turbulent transport of mass and energy. Synoptic and sub-synoptic variations on the time scales of hours to days introduce randomness into most microclimatological time series, though their influence is dwarfed by diel and annual time scales, which, respectively, account for approximately 40% and 20% of signal variance, (Baldocchi et al. 2001). At still lower frequencies, decadal time scales contribute a small proportion to signal variability but are of key importance in monitoring climate change. Therefore, excepting eddy covariance measurements, to discern variations on time scales of one hour and longer, Shannon's Sampling
theory (Ifeachor & Jervis 1993) requires that we must obtain samples at least every 30 minutes. Thus, the low frequency sampling rates of between 1 second and 3 minutes used in determining half hour averages was more than sufficient for capturing the signal fluctuations observed in the microclimatological time series.

3.2 Data acquisition

Three approaches were taken in the acquisition of microclimatological and biophysical data: manual measurement at irregular intervals, lower frequency automated sampling averaged over a run length of 30 minutes, and higher frequency automated sampling also run averaged. Most microclimatological measurements were automated for the purposes of increasing data capture and maintaining measurement standardization, however manual samples were taken for initial measurements of soil water content and ancillary precipitation measurements.

Automated high frequency sampling was employed in making eddy covariance measurements. Values of wind velocity and sonic temperature were sampled at 166 Hz but were output at 20.83 Hz by the anemometer\(^1\). Other high frequency sampled signals were digitised at 10 Hz using the 12 bit A/D converters incorporated into the sonic anemometer. Digital signals output by the sonic anemometer were captured and recorded using the Edisol software program (Moncrieff et al. 1997) running on a laptop computer. Because individual sample points were retained with this method, extensive analyses of these data were possible. A more thorough discussion of techniques and sampling requirements associated with this method is described in Chapter 5.

Automated low frequency sampling constituted the majority of microclimatological measurements made at Griffin. The frequency at which measurements were taken was either 1 or 5 seconds, depending upon the number of tasks assigned to the logger. Loggers performing complicated instructions or monitoring numerous sensors could only perform at the lower sampling interval of five seconds because of the time required to carry out logging instructions (Campbell Scientific 2000).

\(^1\) Instrumentation models, manufacturers, and specifications are given in Appendix D and are not listed in the text.
The low frequency sampling of microclimatological signals was accomplished using solid-state A/D conversion devices. These loggers contained microprocessors to allow for programmed instructions of the data acquisition and signal control, as well as volatile memory for storing the averaged signals until they could be retrieved. In addition, several loggers were associated with non solid-state multiplexers, which expanded the signal connection capacity of the loggers. Each multiplexer allowed a logger to use a single channel to acquire either 64 single ended or 32 differential voltage measurements. Because of limited memory capacity, one logger was attached to a compact flash recording device to allow increased storage.

All loggers used in this experiment obtained power from 12-volt batteries. The voltage levels of all but one of these batteries were maintained by AC powered 12-volt DC battery chargers while the fifth battery was charged by a solar panel. Neither approach of charging logger batteries appeared more successful. However, if the current drain on the logger battery is small and towers are available for deployment, the use of solar panels to maintain battery charge was superior in that it did not require the infrastructure and maintenance of distributed mains electricity supply.

Both the sensors and the sampling devices have an associated accuracy and precision that were important in assessing the validity of the run mean values. The accuracy and precision of individual sensors is discussed in the sections describing those sensors. For the data recording devices, measurement accuracy and precision depend upon the method of measurement and the sensitivity of the analogue to digital (A/D) conversion process of the loggers. There was no signal error induced by multiplexing of sensor signals. The accuracy and precision associated with the data conversion processes used in this experiment are listed in appendix D.
3.3 Instrument deployment

Microclimatological sensors were deployed at three locations within the experiment site. A small set of sensors was associated with the power generating equipment, a larger set was placed on and around the eddy covariance tower, and an extensive set of sensors was placed on and around the profile tower. Information concerning instrument deployment dates is given in appendix E.

![Figure 3.1 Map of the centre of the Griffin site showing the position of the eddy covariance tower, profile tower, and the power generation hut. The circles adjacent to the hut are the locations of the tipping bucket rain gauges. The greyed areas represent regions of forest cover.](image)

Power hut

Two precipitation sensors were associated with the logger placed in the power hut. One sensor was placed in the center of the ride some 15 m west of the power hut, figure 3.1. The second sensor was mounted above the power hut, on the solar panel support structure, at a height of 5 m. These sensor placements were selected as two possible alternatives of minimizing interference resulting from precipitation interception by surrounding forest.
Eddy covariance tower

The position of the eddy covariance tower (figure 3.1) was selected for its more homogeneous southerly fetch. The suite of sensors located on and about the mast holding the eddy covariance instrumentation (figure 3.2) complemented those on the profile tower. The initial positions of sensors from July of 1996 to January of 1997 were lower on the tower (not shown in figure 3.2), coincident with or below the top of the canopy. The instrument positions were raised in January of 1997 to obtain measurements more representative of above-canopy atmospheric conditions. However, because of the presence of eddy covariance instrumentation at the top of the tower, other instruments could not be situated at or near the top of the tower. The resulting placement led to shading of some of the sensors by the tower at low sun angles in mid summer.

Table 3.2 Comparison of mean DBH values for the four 0.01 ha plots surrounding the base of each tower and a mean for the experiment site. DBH values were taken from the 1997 sampling.

<table>
<thead>
<tr>
<th>Exp. Site</th>
<th>Profile Tower</th>
<th>Eddy Cov. Tower</th>
</tr>
</thead>
<tbody>
<tr>
<td>DBH, cm</td>
<td>10.3</td>
<td>9.9</td>
</tr>
</tbody>
</table>

Profile tower

To minimize interference of the flow for the eddy covariance sensors on the eddy covariance tower, most microclimate sensors were placed on the profile tower. Additionally, the sizes of the trees surrounding this tower were more representative of the forest than those immediately surrounding the eddy covariance tower, table 3.2, thus providing a more representative environment for within canopy, surface level instrumentation and surface sensing radiation measurements.

Incoming, reflected, and diffuse radiation sensors were situated at the top of the tower in order to provide an unobstructed view of the sky. To minimize obstruction from adjacent instruments, sensors were placed so that their sensing elements were at the same level. Only the diffuse sensor shade band produced partial sky obstruction to these sensors. For this reason, the shade band was placed as far north from the other sensors as was practical on the tower’s top platform.
Figure 3.2: Schematic of instrument placement on the profile and flux towers at the Griffin site.
Installation levels for profile measurements were chosen on a double logarithmic spacing. It was assumed that wind velocity and scalars would increase roughly logarithmically with height above their zero plane displacement levels, which varied between 4 and 6 m over the course of the experiment (see section 4.12.2). It has also been observed that, within canopies, scalars show rapid change close to the ground (Gillespie 1971; Rannik et al. 2003; Raupach 1989). Utilizing these two spacing assumptions resulted in the vertical instrument spacing shown in figure 3.2.

Most surface measurements near the profile tower were grouped into two locations approximately south and south-southeast of the tower, figure 3.3. Selection of these sites was limited by distance to power and loggers. Within these limits, the two sites were selected based on their degree of canopy closure. At the time of installation, the southerly location had obtained full canopy closure with no surface vegetation while the canopy at the south-southeasterly location was still open with grasses and mosses covering the surface. No attempt was made to determine if the gap size...
associated with the open canopy location was representative of site forest gaps. The two plots will hereafter be referred to as the open canopy and closed canopy plots. A similar suite of sensors was placed in each of the two plots. When sufficient sensors were available, they were placed to characterize the microtopography resulting from the ploughing carried out as preparation for tree planting.

3.4 Data logistics

The memory capacity of logging devices used in this experiment varied from eight days to several months. It was the smallest of the capacities that determined the frequency of data collection, and thus site visits. If the hard disk of the high frequency sampling logger (i.e. the eddy covariance laptop personal computer, PC) neared its capacity, raw data collection was stopped while mean values continued to be recorded so that loss of averaged values did not result from reaching the limit of the systems storage capacity. However, if the low frequency sampling loggers filled their storage buffers, they would begin overwriting the earliest recorded data in the buffer, resulting in loss of data from the earliest part of the data collection period. The desire to maintain a continuous data set, perform sensor maintenance and ensure that power was available at the site dictated that weekly visits be made to the site. During a site visit, data from each of the loggers was retrieved and the logger updated as required.

A portable computer was used to interrogate the low frequency sampling loggers. While interrogating the loggers, sensors were assessed for signal quality in real time and data was downloaded via a serial port connection using logger specific software\(^2\). Data was downloaded to the PC in delimited text format while the compact flash storage card employed by one logger was swapped with a card which had had its data removed. Sensor problems determined by real time monitoring were resolved if possible; otherwise, information detailing the problem was recorded in the field log as a reminder for further action. The loggers’ times were

\(^2\) Details of software applications used in the experiment are given in appendix F.
synchronized if they were more than one minute different from coordinated universal time (UTC)\textsuperscript{3}.

Upon arrival, operational status of the high frequency data logging system was determined, because a PC was used to record the high frequency sampled data, it was possible to monitor signal quality via that PC. If necessary, the system and sensors were restarted. Questionable sensor operation was repaired or a note was made in the field log for further action. Logger time was checked and reset as for the other logging devices. In order to retrieve data from this logging device it was necessary to terminate operation of the logging program. Raw and run averaged data were transferred from the PC to removable storage "Zip" disks. To minimize the amount of eddy covariance data lost, the period used for data retrieval coincided with instrument checks, maintenance, and calibration.

Upon return to Edinburgh, data were transferred from the PC, compact flash card, and removable storage disk to a computer in the laboratory. Because the file labels did not indicate the time of file creation or termination, data from the low frequency sampled data were stored in a subdirectory specific to that field visit. Files containing averaged results from the high frequency sampled data were stored in a subdirectory specific to that year of data collection. Similarly, the high frequency raw data were stored in a year-specific subdirectory. Periodic backups were made of the raw and run-averaged data and stored off-site.

Run-averaged low and high frequency sampled data were loaded from their delimited text file formats into database tables (\textit{Paradox} version 5) with a separate table maintained for each data logging device. An upload utility was developed that allowed repeatable uploads with adjustment for incorrect date/time stamps in the run-averaged text files.

A schematic of the data management and processing workflow, figure 3.4, details the flow of data from the field to the final output. This schematic incorporates

\textsuperscript{3} Coordinated universal time is the currently accepted format of military and civilian timekeepers. For the purposes of this thesis, it is equivalent to Greenwich Mean Time (GMT). Unless otherwise specified all times given in this thesis are given as UTC.
concepts from section 3.3 as well as processes used in obtaining results given in the remainder of chapter 3 and in chapter 4.

Figure 3.4 Schematic of data management and processing workflow. Loggers are represented by rounded rectangles, data storage by circles and data processing by rectangles.

3.5 Data quality assessment

3.5.1 Quality assessment concepts

Data quality assessment as applied in this thesis does not treat the known static and dynamic characteristics that define a sensors response; it is assumed that these characteristics were known and behaved consistently between calibrations. As employed, quality assessment refers to the treatment of transient sensor
characteristics (i.e., noise) and invalidation of underlying measurement theory. Noise is the random departure of the signal from the expected signal caused by the sensor or interaction of the sensor with its environment (Brock & Richardson 2001). Invalidation of measurement assumptions results primarily from inappropriate deployment and measurement techniques, which may be either static or transient.

The quality assessment of microclimate variables involved evaluation of instrument calibrations, signals, and run-averaged results to determine if values and responses to environmental conditions were reasonable and proportionate. For identification purposes, noise was categorized as either signal drift, spikes and dropouts, or high frequency noise.

Signal drift was treated as a calibration quality problem because it corresponds to undesired signal oscillations at frequencies much lower than signal oscillations of interest (i.e., the run length). Drift was treated as a quality problem of the run-averaged data only when data could not be corrected using a reliable reference sensor. Spikes and dropouts refer to undesired signal oscillations in the frequency range of primary interest. They can often be identified as an individual data point, or sets of contiguous points, whose values deviate from the expected distribution of values. High frequency noise, as the name implies, refers to undesired signal oscillations at a frequency range higher than that of primary interest. The difference between spikes and high frequency noise is indistinct as it depends upon the users definition of the frequency range of interest.

The primary target of data quality assessment was the run-averaged data, although quality assessment was also applied to, or made use of, associated metadata, such as field logs, calibration data and the high frequency raw data. Run means and standard deviations (when available) were primarily assessed by comparison with spatially or temporally corresponding data. Run-averaged data provided only partial information about data quality because of lack of information on possible high frequency noise and spikes, which, because of their transient nature, were difficult to detect in run-averaged data unless they were very severe. When available, the best resource for assessing data spike and noise problems was the underlying high frequency sampled raw data.
Frequent calibration data provided both a method to assess signal drift and a way to remove it, but did not provide information with respect to spikes or instrument noise. Additionally, calibration data were also subject to their own quality assessment.

The field logbook detailed specific activities, calibrations, problems, and other comments associated with the operation of the experiment site. As these notes were only available for the times when an operator was on site, they provided a limited view of the equipment status. The notes did not address signal quality problems directly but instead gave insight into potential causes of data quality problems. It was often necessary to interpolate log notes based on run-averaged signal characteristics, or on previous experience with the particular item. However, these notes did provide an invaluable source of information for enhancing and justifying later stages of data quality assessment.

### 3.5.2 Quality control methods

Two similar approaches were used to the same end when applying quality assessment to the run-averaged data. The first approach was to assign a quality control flag or flags to individual run-averaged data and remove the data from the processing stream at a later stage. The second approach was to remove the data from the processing stream as part of the quality testing process. Neither approach removed the initial data, which were retained in their original format. The use of quality control flags was deemed superior as it allowed, when necessary, quantitative analysis incorporating the data that had been quality flagged.

Based on previous experience it was known that although field log entries can be of great use, they can be difficult to re-interpret, and present difficulties in relocating notes on a particular problem. Therefore, all field log notes were entered into a database with each note categorised by date, time, sensor/equipment, and operation. An additional field of comments was also available for field calibrations, and more lengthy notes, figure 3.5. The resulting capability to search and order the recorded
log information, proved invaluable to data quality assessment. Sensor-specific log entry reports were created to assist quality assessments during reviews of signal quality.

<table>
<thead>
<tr>
<th>Date</th>
<th>Year</th>
<th>Begin</th>
<th>End</th>
<th>Sensor ID</th>
<th>Category</th>
<th>Record</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/25/1997</td>
<td>1997</td>
<td>19:30</td>
<td>255</td>
<td>17:00 Licor fre 2</td>
<td>Change</td>
<td>Adjusted purge flow rate to 0.2 SCFH</td>
<td></td>
</tr>
<tr>
<td>1/25/1997</td>
<td>1997</td>
<td>19:30</td>
<td>255</td>
<td>17:00 Licor fre 2</td>
<td>Calibration</td>
<td>Step 2 Set CO2 span using 350 ppm tank - dial 4.19±0.54, values</td>
<td></td>
</tr>
<tr>
<td>1/25/1997</td>
<td>1997</td>
<td>19:30</td>
<td>255</td>
<td>17:00 Licor fre 2</td>
<td>Calibration</td>
<td>Step 1 Reset zeros (dial CO2 0.977.50, values CO2 0.90±0.50)</td>
<td></td>
</tr>
<tr>
<td>1/25/1997</td>
<td>1997</td>
<td>17:30</td>
<td>255</td>
<td>17:30 Plots</td>
<td>Observation</td>
<td>soil moisture sampling @ plot SW of lower 2</td>
<td></td>
</tr>
<tr>
<td>1/26/1997</td>
<td>1997</td>
<td>17:40</td>
<td>255</td>
<td>17:40 Logger 5</td>
<td>Data</td>
<td>dump data</td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.5 Sample of field log database entry.

The calibration data from sensors with frequent calibrations were manually analysed and checked for unrealistic results. Such results were removed from the data stream before determining calibration factors. If erroneous sensor calibrations could not be resolved, that information was used as justification for the setting of quality flags.

Infrequent calibrations were available for many sensors. In some circumstances, if a reliable sensor were available for comparison, it would be used as an *ad hoc* reference for adjusting signal drift and determining the presence of noise and spikes. Note, however, that signals adjusted for drift in this manner could no longer be considered reliable within the range of error observed for the correcting signal and the expected spatial variability observed between the two sensors.

While run-averaged values derived from high frequency sampled data benefited from quality assessment of the raw data, the sheer volume of raw data required the use of automated techniques to assess data quality. A modified version of Hojstrup's (1993) spike detection method was implemented. Quality assessment of noise, amplitude resolution, and stationarity was based on work by Vickers and Mahrt (1997), while quality assessment based on turbulent intensity and integral turbulence tests employed the techniques of Foken and Wichura (1996). Typical and higher order signal statistics were also incorporated into quality assessment.
Results of these techniques for variables derived from signals (for which high frequency sampled data were available) were used to verify subjective, manually determined quality control flags.

The relatively large amount of low frequency averaged data (approximately 100,000 values of 200 variables) suggested that quality assessment would benefit from automated techniques, yet many of the errors encountered were of a nature which made automated techniques impractical. Although statistical values other than means and standard deviations could have been retrieved - as could have the raw signals, such a broadening of the data stream would have required greater resources, making these changes uneconomic. Automated analysis was restricted to tests for realistic ranges of signal means and standard deviations (appendix G). These ranges were slightly less restrictive than those specified as typical by FLUXNET (2002) based on the experience that bad signals were usually well beyond expected maxima. Data that did fall beyond the predetermined bounds were excluded without subjective analysis.

Visual inspection was the primary method of quality assessment of run-averaged values that fell within a reasonable range. This process depended almost exclusively on the use of signal mean values, as they proved easiest to interpret when viewed as a time series. When possible, visual quality assessment was applied in parallel to time series of two or more measurements. These data sets consisted of values either from duplicate sensors or from sensors of similar expected values. For example, a pyranometer may have been compared with a PPFD sensor because of the similar responses of these sensors to solar radiation. The method employed was to assign a numerical quality control flag of 0, 1, or 2 to each recorded value. A flag of 0 indicated the value contained no known problems, while a flag of 1 indicated a problem with the value but that the magnitude of the error, the character of the error or the potential impact on other signals was not significant enough to cause immediate exclusion. A flag of 2 indicated that the result was unusable for further processing and analysis, and should be excluded from the data stream at the appropriate location.
3.5.3 Quality control effects

The effect of application of quality control on the amount of available data is shown in table 3.3. Data for the experiment ranges from about 70 to 90% (average of 76%) for climate variables and is 70% for fluxes. The effect of quality control removes on average a further 9% of data, though this amount varies greatly,
between about 1 and 40%. It is probably accurate to say that the amount of data removed by quality control efforts reflects the difficulty in operation and maintenance of the measured variable. No efforts were taken to determine the quantitative effects of not removing quality-controlled values upon the variables’ results.

Table 3.3 Data coverage prior to quality control and further reductions in data coverage as a result of quality control expressed as percent of experiment period for variables and fluxes.

<table>
<thead>
<tr>
<th>QC reduction %</th>
<th>Data coverage, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil moisture</td>
<td>0 26</td>
</tr>
<tr>
<td>Air pressure</td>
<td>1 77</td>
</tr>
<tr>
<td>Net radiation</td>
<td>3 81</td>
</tr>
<tr>
<td>Air temperature</td>
<td>4 81</td>
</tr>
<tr>
<td>Wind speed</td>
<td>4 95</td>
</tr>
<tr>
<td>Transmitted radiation</td>
<td>5 75</td>
</tr>
<tr>
<td>Air temperature profile</td>
<td>5 68</td>
</tr>
<tr>
<td>Wind direction</td>
<td>6 95</td>
</tr>
<tr>
<td>Bole temperature</td>
<td>6 76</td>
</tr>
<tr>
<td>Diffuse radiation</td>
<td>7 75</td>
</tr>
<tr>
<td>Precipitation</td>
<td>8 84</td>
</tr>
<tr>
<td>Global radiation</td>
<td>9 80</td>
</tr>
<tr>
<td>Soil heat flux</td>
<td>10 75</td>
</tr>
<tr>
<td>Humidity</td>
<td>11 81</td>
</tr>
<tr>
<td>Soil temperature</td>
<td>11 76</td>
</tr>
<tr>
<td>Wind speed profile</td>
<td>15 73</td>
</tr>
<tr>
<td>Reflected radiation</td>
<td>24 78</td>
</tr>
<tr>
<td>CO₂/H₂O profiles</td>
<td>38 73</td>
</tr>
<tr>
<td>Momentum flux</td>
<td>5 70</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>9 70</td>
</tr>
<tr>
<td>CO₂ flux</td>
<td>9 70</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>11 70</td>
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3.6 Summary

The experiment, as designed, was appropriate for capturing temporal variability on time scales of minutes to years; however, the number of sensors employed (for within canopy work) was insufficient to represent spatial variability accurately. Although no extraordinary measures were required to maintain proper logging of data, data loss, primarily a function of power restrictions and site distance, did limit temporal coverage. Computerized field/lab notes and thorough backups and records
of data flow helped maintain data quality. Nevertheless, a lack of sufficient information about field behaviour required manual review of data to maintain the highest degree of data quality. Quality control measures removed a substantial (9%) amount of poor quality data.
4 Griffin climatology

4.3 Introduction

This chapter presents the climatology of the Griffin field site over the period 1997 to 2001. Results and analysis for the measurements of radiation, temperature, humidity, carbon dioxide, wind speed, wind direction, precipitation, pressure, soil moisture, and soil heat flux are presented. Emphasis is placed on comparison across years, interesting observations and departures from expected behaviour. The details of measurement methods, including sensor operation, calibration, installation and maintenance, as well as instrument intercomparisons, are contained in appendix A.

4.4 Global short-wave irradiance

Global short wave irradiance, $R_g$, is defined as the total energy received as radiation per unit surface area tangential to the earth's surface; its measurement includes both direct and diffuse components. Also used throughout this thesis is the value of potential short wave radiation, $R_p$. This value represents the value of $R_g$ expected at the surface for standard clear sky conditions; its calculation is described in appendix B.4.

![Figure 4-1 Date/Time contour plot of $R_g$ for the Griffin site. Half hour mean values were not further averaged over the time axis but were averaged over two week periods for date axis data.](image)

The temporal variation of $R_g$ at annual and diel time scales, figure 4.1, exhibits peaks corresponding to mid-day and mid-summer, associated with the maxima in
solar elevation angles at those times. Nocturnal values are appropriately near zero while maximum variability appears to occur during midday.

![Graph showing monthly means of solar radiation](image)

**Figure 4-2** Five-year (filled circles) and annual (open circles) monthly means of $R_0$. Modelled $R_p$ (no symbol) scale with the right vertical axes.

Five-year and annual monthly mean values of measured $R_0$ and $R_p$ are presented in figure 4.2. Also shown is a comparison of the five-year average $R_0$ and long term average $R_0$ obtained from the University of East Anglia’s gridded solar radiation data set, New et al. (1999) and from data of Page and Lebens (1986). The high winter values of the UEA data set suggest that these data may be erroneous. The Page and Lebens data compare favourably with the Griffin. The pattern of $R_0$ follows that of $R_p$ with the exception of the month of May, which stands out as the defining characteristic of the annual pattern, this anomaly also appears to be present in the Page and Lebens (1986) data but is weaker and is also present in April and June. This difference suggests that the phenomena may be related to the inland location of Griffin as compared to the coastal location of the data from Page and Lebens (1986). Based on the annual pattern of the ratio of measured to $R_p$ shown in figure 4.3, and assuming the ratio for the month of May is an outstanding value, it was determined that the May peak of 200 Wm$^{-2}$ is approximately 15% larger than expected. This difference corresponds to an additional 31 W m$^{-2}$ during this period. The timing of this peak corresponds to the period during which mean land surface temperature of the British Isles is closest to that of the, warmer, surrounding sea surface temperature (Mayes et al. 1997). During the remainder of the year, the sea surface temperature is higher than that of the land, resulting in conditions conducive
to cloudiness. A link between high radiation and high atmospheric pressure suggests an alternate cause of sunny conditions in May, see figure 4.73. The summer peak in monthly average value of $R_g$ varies annually by approximately $\pm 25$ W m$^{-2}$. The years 1997, 1999, 2000, and 2001 exhibited above average radiation in autumn, summer, late spring and spring, respectively. During 1998, $R_g$ was below average for almost the entire year. Below average conditions were observed in spring and autumn in 1997 and 1999, while 2000 had below average radiation in autumn and 2001 had below average radiation from late spring through autumn.

![Figure 4-3 Ratio of measured to potential short wave radiation. The open circle represents an arbitrary 'expected' value based on a smooth annual curve.](image)

The variability of radiation at all temporal scales can be assessed in more detail by inspecting the spectra of the time series of $R_g$ (figure 4.4). In this figure, the spectrum of $R_p$ is presented to allow identification of variability in $R_g$ of astronomical origin. The spectrum of $R_p$ has strong peaks at frequencies corresponding to diel (0.042 Hz) and annual (0.0001 Hz) cycles. Both of these peaks also have smaller peaks at frequencies higher than their associated annual and diel peaks. These subordinate peaks of the diel cycle represent the redistribution of energy from the primary peak caused by the truncated sinusoidal shape of the diel potential radiation cycle associated with the sharp transition from day to night. Measured $R_g$ has annual and diel peaks identical to those of $R_p$. Variability at other time scales is largest about the diel peak and appears to fall off linearly at both higher and lower frequencies. Very weak peaks are seen at time scales of four days (0.01 Hz) and 21 days (0.002 Hz). The four-day cycle is related to the passage of synoptic scale weather systems but the cause of the 21-day peak is not known. These results are nearly identical to those of Baldocchi et al. (2001) who have
made a similar examination of spectra for biological and meteorological variables on similarly long time scales.

Figure 4-4 Spectral power curves of measured $R_g$ (circles) and $R_p$ (grey line) for frequencies from 0.00002 Hz (5 Years) to 1 Hz (1 hour). The spectral powers at each frequency, $S(n)$, have been normalized using the ratio of frequency to signal variance (Stull 1988).

Figure 4-5 Five-year weekly values of the standard deviation of run mean $R_g$ (triangles), coefficient of variation of $R_g$ (circles), and coefficient of variation of $R_p$ (solid line).

Irrespective of frequency, the temporal predominance of variability can be assessed using the coefficient of variation, CV, of the fluctuations in $R_g$, figure 4.5. The standard deviation of $R_g$ is a factor of five larger in summer than in winter, but
when expressed as CV the data suggest that the variability of \( R_g \) is relatively larger in winter than during summer. However, the close correspondence between the patterns of CV of \( R_g \) and \( R \) indicates that the larger relative variability can be attributed to the diel cycle and that the variability resulting from cloudiness is similar throughout the year.

In addition to knowing the amount of radiation, the quality of radiation is important because of its role in photosynthesis. The presence of clouds will affect the proportion of diffuse radiation (Lam & Li 1996) and thus the penetration of radiation into the canopy (Landsberg et al. 1973; Norman & Jarvis 1975). The distribution of radiation within the canopy will in turn affect photosynthesis and transpiration (Gu et al. 2002; Wang & Jarvis 1990). While diffuse radiation was measured (see sec 4.4), the number of measurements was limited (see appendix A.1.2), and an alternative method was required to estimate clear sky conditions. Two common definitions of bright sunshine are \( R_g \) of at least 150 W m\(^{-2}\), or a minimum threshold of 40% of the extra-atmospheric short-wave radiation (Campbell Scientific 1985). For the latter definition of bright sunshine, a threshold of 45% of the potential surface radiation has been employed as an equivalent definition. Additionally, as a test of this approach, a subjective estimation of the number of clear and half-clear days was tallied based on visual inspection of the global radiation data. It should be emphasized that the subjective analysis curve is presented here as a lower boundary limit and as a check of pattern representation; it does not provide a practical method of determining sunshine.

Results for both calculated and subjective approaches shown in figure 4.6, suggest that monthly averaged sunshine hours per day varied between 3 in the winter to 9 in the summer, similar to the range observed by Page and Lebens (2002). This annual pattern is an artefact of the change in day length over the year. However, as with annual curves of \( R_g \), we see a peak during the month of May with an extra hour of sunshine per day more than other summer months. This peak is predominant in the curves representing percent of potential radiation. These curves also have a smaller annual range than does the curve based on a minimum threshold value of \( R_g \).
The discrepancies between the minimum threshold value and percent of potential estimates raise the question of which approach provides a better estimate of sunshine conditions. Both the minimum threshold $R_g$ case and 45% of potential case show similar patterns in figure 4.6; however, when presented as percent sunshine hours on a diurnal basis (figure 4.7), it is observed that the curves representing the minimum threshold $R_g$ and 45% of $R_p$ have very different patterns. The minimum threshold $R_g$ value curve varies with the diurnal $R_g$ cycle while the percentage curve is relatively constant over the day. Two extreme months are presented in this figure but similar patterns exist for other months.

Assuming there is no diurnal preference for clear skies at this site, the patterns in figure 4.7 suggest that the minimum threshold $R_g$ estimation under-represents clear sky conditions for low solar elevation angles (morning, evening, winter) and over-represents them at high solar elevations (summer mid-day). The estimates employing percentage $R_p$ appear to provide a more consistent estimate. Cases (not shown) employing the more restrictive 75% of percent of potential radiation had similar patterns.
Figure 4-7 Comparison of estimation of sunshine for the months of January and June as a function of hour of the day.

Figure 4-8 Comparison of monthly averaged diel patterns of $R_g$ for high radiation conditions ($R_g > 70\%$ of $R_p$, dotted line) and low radiation conditions ($R_g < 30\%$ of $R_p$, solid line). Curves have been normalized by their respective monthly mean to facilitate curve comparison.

The curves employing the percentage of potential radiation (e.g. figure 4.7) show a slight increase in percent sunshine over the day during most months (not shown), as well as an increase during sunrise and sunset. The rise in percent sunshine during morning and evening may be the effect of an inadequate model of $R_p$ or sensor cosine response but warrants further investigation. However, the diurnal increase
appears in the plots of monthly averaged diel curves normalized for average diel radiation, (figure 4.8). These plots show a shift in the diel curve of $R_g$ to later times under lower average diel radiation (< 30% of potential) as compared to higher radiation conditions, suggesting a decrease in cloudiness over the diurnal period on cloudy days (low $R_p$).

4.5 Global PAR

Although photosynthetic photon flux density, $Q_{pg}$, or PPFD (units of $\mu$mol m$^{-2}$ s$^{-1}$), is the quanta equivalent measure of PAR (units of W m$^{-2}$), their units are not directly convertible because of the difference in energy transferred by photons of different wavelengths. An accurate conversion would require knowledge of the respective flux densities for all wavelengths within the 0.4 to 0.7 $\mu$m band. Studies which provide factors for the conversion between units (Jacovides et al. 1999; McCree 1984) suggest that the conversion is similar under many conditions. However, to avoid the potential for conversion errors, comparisons of $Q_{pg}$ with short-wave radiation will be in their respective units of $\mu$mol m$^{-2}$ s$^{-1}$ and W m$^{-2}$.

![Figure 4-9 Relationship of global PPFD to global short wave radiation.](image)

Plots of Date/time contours and annual monthly deviations from the five-year mean for $Q_{pg}$ are not presented as they exhibit patterns nearly identical to those for $R_g$, as a result of the close relationship of $Q_{pg}$ to $R_g$, as exemplified in figure 4.9. The discussion of annual and diurnal patterns presented in the preceding section is therefore similarly valid for $Q_{pg}$.
Because high frequency sampled $Q_{pg}$ values were available, it was possible to extend the spectrum of $Q_{pg}$, figure 4.10, to higher frequencies than those presented for $R_g$ in figure 4.4. As expected, the spectral region from 5 years to 1 hour for $Q_{pg}$ is nearly identical to that for $R_g$. Additional information in the spectral region between 1 hour and 0.1 second show the spectral power of $Q_{pg}$ to gradually roll off until approximately 200 cycles per hour (0.055 hz), at which point noise becomes a predominant feature. The sharp drop off at frequencies between $10^3$ and $10^4$ cycles per hour is the result of a low pass filter, with a 1 s time constant, applied to the data. This additional high frequency information is notable because variations in $Q_{pg}$ appear to fall off more rapidly at a frequency of about 1 hr$^{-1}$. More detailed analysis would be required to determine if this falloff is related to cloud size and velocity.

![Figure 4-10](image)

**Figure 4-10** Spectrum of $Q_{pg}$; this spectrum is a combination of two spectra. The region from $10^0$ to 1 was obtained by transforming half hour values of $Q_{pg}$ while the region from 1 to $10^4$ was obtained by transforming high frequency sampled data from the year 1999. The spectral magnitudes were matched at the point of intersection so the magnitude axis may not be accurate. The spectral powers at each frequency, $S(n)$, have been normalized using the ratio of frequency to signal variance.

While the gross patterns of $Q_{pg}$ are similar to those of $R_g$, there are subtle differences between the two measures because of the more limited spectral region covered by $Q_{pg}$. The ratio of $R_g/Q_{pg}$ is used to investigate these differences. In a cloud-free, clean atmosphere, Rayleigh scattering of visible radiation is larger than
that of near infrared radiation (Wallace & Hobbs 1977), resulting in larger values of $R_g/Q_{pg}$ in direct as compared to diffuse radiation. For overcast skies, MIE and optical scattering of visible and NIR radiation by water droplets should result in negligible effects of cloudiness upon the $R_g/Q_{pg}$ ratio. However, Evans and Puckrin (2003) have found evidence of strong absorption of near infrared radiation by cloud droplets, supporting the smaller values of $R_g/Q_{pg}$ found under cloudy conditions in this experiment. These effects are also apparent in the values of $R_g/Q_{pg}$ from other experiments (Ross & Sulev 2000; Wesely 1982) given in table 4.1. The values of $R_g/Q_{pg}$ in table 4.1 are presented in order of expected ascendance. The values in table 4.1 show an increase under conditions in which near infra-red (NIR) radiation energy is enhanced relative to PPFD. Only the clear or overcast global data from Ross and Sulev (2000) show a deviation from this monotonic trend.

Table 4.1 Values of $R_g/Q_{pg}$ for different sky conditions and for canopy reflected and transmitted radiation. Values presented are in units of $J\text{ mol}^{-1}$.

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<thead>
<tr>
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</thead>
<tbody>
<tr>
<td>Clear sky diffuse</td>
<td>0.470</td>
<td>0.433</td>
<td></td>
</tr>
<tr>
<td>Overcast global</td>
<td>0.526</td>
<td>0.553</td>
<td></td>
</tr>
<tr>
<td>Clear sky global</td>
<td>0.592</td>
<td>0.551</td>
<td>0.435</td>
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<tr>
<td>Clear sky direct</td>
<td>0.569</td>
<td>0.569</td>
<td>0.459</td>
</tr>
<tr>
<td>Overcast reflected</td>
<td>1.47</td>
<td>2.150</td>
<td></td>
</tr>
<tr>
<td>Clear sky reflected</td>
<td>1.64</td>
<td>2.164</td>
<td></td>
</tr>
<tr>
<td>Clear sky transmitted</td>
<td>0.88 to 5.03</td>
<td>0.87 to 4.90</td>
<td></td>
</tr>
</tbody>
</table>

The pattern of $R_g/Q_{pg}$, in relation to solar elevation angle shown in figure 4.11, verifies the expected decrease in $R_g/Q_{pg}$ associated with shorter atmospheric travel paths at high solar elevation angles. The approximately 7% increase in $Q_{pg}$ relative to $R_g$ with increasing solar elevation angle is consistent with other research (Dickinson 1983; Gonzalez & Calbo 2002). To verify this relationship the atmospheric transfer model SBDART (Ricchiazzi et al. 1998) was run using typical atmospheric inputs and outputs of total shortwave and PAR insolation spectral energies. The ratio of these values were calculated for a range of solar elevation angles and adjusted for average optical depth at each solar elevation angle. The resulting patterns are very similar although the energy to quanta conversion factor
required to match the two curves, i.e. 4.05 μmol J⁻¹, was approximately 10% smaller than the accepted value of 4.57 μmol J⁻¹ (Dye 2004). This difference can be accounted for by the difference between quantum and energy measurements of quantum PAR, as discussed by Ross and Sulev (2000).

Although Raleigh scattering can explain the effect of solar elevation angle on $R_g/Q_{pg}$ under clear skies, the scattering of light in the presence of clouds is more complicated. To determine if cloudiness was affecting proportions of PPFD and NIR radiation components, $R_g/Q_{pg}$ data grouped into three 15-degree solar elevation angle classes are presented in relation to percent potential radiation in figure 4.12. As expected, the effect of increasing cloudiness (low %R_p) and lower solar elevation angles was to reduce the proportion of NIR radiation relative to that of PPFD, thus decreasing $R_g/Q_{pg}$. Reductions of up to 16% over the range of %R_p and 8% in relation to solar elevation under clear conditions. Although $R_g/Q_{pg}$ had the same pattern with respect to %R_p for all solar elevations, the effects of solar elevation angle disappeared under cloudier conditions because the effects of Raleigh scattering become insignificant in comparison to the attenuation of NIR radiation by clouds.
Figure 4-12 The distribution of values of the ratio of $R_g$ to $Q_{pg}$ at levels of %$R_p$. The %$R_p$ greater than 90% have small sample sizes. Error bars represent one standard error (n>25). The model curve (open squares) were obtained from the SBDAT atmospheric transfer model, (Ricchiazzi et al. 1998). Model assumes a conversion of $Q_{pg}$ using the empirical coefficient 4.05 μmol J⁻¹.

The significance of these effects can be determined from the frequency distribution of the percent potential radiation, figure 4.13. A bimodal distribution of percent potential radiation exists with a peak at ~20% and a smaller peak at ~90% of potential short-wave radiation. It is apparent that overcast and mostly cloudy conditions predominate, while mostly clear sky conditions are slightly more predominant than partly cloudy conditions. These conditions contrast sharply with similar observations over a more continental site (Gu et al. 1999) for which the distribution peak is between 70 to 80% of potential. Conditions in which excess radiation is observed (> 100% of $R_p$) are primarily limited to low solar elevation conditions and may reflect either small errors in either measurement of short-wave radiation or the model of $R_p$. The distribution suggests that for a majority of observations, i.e. cloudy conditions, the ratio $R_g/Q_{pg}$ will change primarily as a function of %$R_p$. For a smaller number of observations the value of $R_g/Q_{pg}$ will change as a function of both %$R_p$ and solar elevation angle.
Figure 4-13 The frequency distribution of values of $R_g$ as a function of \%$R_p$. Additionally, data are grouped by four 15-degree solar elevation angles.

Figure 4-14 Contour plots of $R_g/Q_{pg}$ against solar elevation angle and Date for the years 1997 to 2001. The four plots correspond to levels of \%$R_p$. Values of $R_g/Q_{pg}$ (units of J $\mu$mol$^{-1}$) were averaged after grouping by 5 degree bins for solar elevation angle and one month bins for date. Values were excluded for periods when $R_g < 40$ W m$^{-2}$ or $Q_{pg} < 80$ $\mu$mol m$^{-2}$ s$^{-1}$, and averaged values were excluded if they contained less than five values.
To assess any residual seasonal effects, contour plots of $R_g/Q_{pb}$ against solar elevation angle and date for four levels of $\%R_p$ are shown in figure 4.14. The annual pattern of $R_g/Q_{pb}$ exhibits little variability under cloudy conditions ($< 50\% R_p$). Under less cloudy conditions a summer minimum and autumn maximum are more exaggerated under less cloudy conditions and at lower sun angles. The summer minimum may be the result of increased NIR absorption because of higher atmospheric water content during those periods (Gonzalez & Calbo 2004; Rapti 2000). The cause of the higher $R_g/Q_{pb}$ values in autumn may similarly correspond to low atmospheric water content associated with lower boundary layer air temperatures.

4.6 Diffuse radiation

Diffuse radiation comprises the radiation reaching the earth's surface that has been reflected from air, clouds or particulate matter in the atmosphere. It therefore depends greatly upon the composition of the atmosphere and cloud cover.

Because the available data was limited and sparsely distributed over a two and a half year period, it was necessary to find an appropriate method to extrapolate the diffuse radiation data. As clouds have a significant effect in absorbing direct radiation, it can be assumed that they also affect the proportion of diffuse radiation. With this in mind, and assuming that the $\%R_p$ is a reasonable estimator of cloud cover and has been used as a factor in assessing the behaviour of diffuse radiation, figure 4.15.

The available data were grouped by days having similar categories of average diurnal $\%R_p$. This effectively classified the data into five levels of sky conditions ranging from very cloudy ($\%R_p$ of 0 to 25) to mostly clear ($\%R_p$ of 75 to 100). The class with $\%R_p > 100$ consisted of a smaller set of data ($n = 10$ to 20) than the other four classes ($n = 50$ to 200). In addition, because data were obtained from different times of the year, time was plotted as percent of the diurnal and nocturnal periods instead of time of day.
Figure 4-15 Comparison of diffuse (open circles) and global (filled circles) short-wave radiation for days with different average levels of percent potential radiation. Time of day is presented as percent of diel period on the x axis.

Figure 4-16 Relationship of percent diffuse radiation to the percent underestimate of $R_p$. Black spots are for solar elevations greater than 30 degrees while grey spots are for solar elevations less than 30 degrees. The model is described in appendix B.6.

It is observed in figure 4.15 that the proportion of diffuse radiation increases with global radiation for the 0 to 25% and 25 to 50% classes of $\%R_p$. Values of $R_d$ then appear to reach a maximum and decline to smaller values under sunnier conditions. This relationship indicates that $\%R_p$ can be related to the amount of diffuse radiation. This relationship, shown in figure 4.16, indicates a strong relationship between percent $R_d$ and $\%R_p$ at solar elevation angles greater than 30 degrees.
similar to the findings of Lam and Li (1996). At smaller solar elevation angles, the relationship remains valid under cloudy conditions but breaks down under clearer conditions. Accuracy in determining diffuse radiation at low solar elevation angles under clearer conditions is likely the cause of greater scatter under such conditions. It is also likely that inaccuracy in the estimates of $R_d$ at low solar elevation angles contributes to the observed scatter.

![Graph](image)

**Figure 4.17** Relationship of percent diffuse to $\%R_p$ (circles with std err bars) and fitted with a 5th order polynomial (dotted line). The solid line is the slope of the polynomial model of percent diffuse radiation with respect to percent potential radiation. Error bars represent one standard error ($n>50$).

A second concept obtained from figure 4.15 is that a $\%R_p$ exists that corresponds to an optimum in the proportion of diffuse radiation. In figure 4.15 it appears that this optimum is achieved at a $\%R_p$ of between 50 and 75 percent. In figure 4.17, all available $R_d$ data were plotted against $\%R_p$ and fit with a polynomial. The slope of this polynomial suggests that the optimum fraction of diffuse radiation occurs at a $\%R_p$ of approximately 70%. It is also interesting to note that this corresponds with the minimum in the frequency distribution of $\%R_p$ observed in figure 4.13. Further analysis is required to convolve this optimum with those appropriate for ecosystem processes.

### 4.7 Reflected short-wave radiation and PPFD

Reflected radiation is any radiation emitted from the Earth’s surface that originated as extraterrestrial radiation. It is measured as the total up welling radiant energy
received per unit surface area tangential to the earth's surface. It is by nature composed entirely of diffuse radiation.

The features of the date/time contour plot (figure 4.18) and inter-annual variability plot (figure 4.19) of $R_r$ are very similar to the corresponding plots for $R_g$ (figures 4.1 and 4.2). One slight difference apparent in figure 4.19 is that the peak of $R_r$ in
May is less dominant (10% less than June value) than it was for $R_g$ (7% larger than June value). This suggests that either the albedo of the canopy is reduced under direct radiation or the values of reflectance are higher in June and July than they are in May, (this will be discussed further in section 4.6).

As in section 4.3, values of $Q_{pr}$ will be discussed in relation to those of $R_r$. As expected from the results of Ross and Sulev (2000) the values of $R_r$ are nearly twice those of $Q_{pr}$ (table 4.1) resulting in values of $R_r/Q_{pr}$ between 0.6 and 1.8. The values of $R_r/Q_{pr}$ are larger than those of $R_g/Q_{pg}$ because of the proportionally greater absorption of PPFD by the canopy (Russell et al. 1997b).

![Figure 4-20 Relationship of $R_r/Q_{pr}$ to solar elevation angle for Griffin. Error bars represent one standard error (n=600).](image)

As with the values of $R_g/Q_{pg}$, we again anticipate a relationship between $R_r/Q_{pr}$ and solar elevation angle, figure 4.20. An increase in $R_r/Q_{pr}$ is observed, a pattern opposite in nature to the decrease in $R_g/Q_{pg}$ with solar elevation angle in figure 4.11. It is known that the reflected proportions of both $R_g$ and $Q_{pg}$ increase at low solar elevation angles because of the bi-directional reflectance characteristics of the canopy (Deering et al. 1999), see section 4.6. However, it appears that the reflected proportion of NIR radiation increases more rapidly than does the proportion of PAR radiation. The results of Deering et al. (Deering et al. 1999) provide conflicting evidence in the two NIR wavebands presented. An alternative, or contributing explanation is that the differences in the cosine response of the sensors (Gonzalez et
causes \( R_r \) to have a smaller response when the low solar elevation causes the largest proportion of canopy reflectance to be seen at small sensor declination angles.

![Graph showing the ratio of \( R_r \) to \( Q_{pr} \) at different levels of \( \% R_p \) for four solar elevation angle classes.](image)

**Figure 4-21** Magnitude of the ratio of \( R_r \) to \( Q_{pr} \) at levels of \( \% R_p \) for four solar elevation angle classes. Sample sizes at \( \% R_p \) greater than 90% are relatively small. Error bars represent one standard error (n>25).

A plot of \( R_r/Q_{pr} \) against \( \% R_p \), figure 4.21, is used to assess the effect of cloudiness on the differences between \( R_r \) and \( Q_{pr} \). For larger solar elevation angles, the value of \( R_r/Q_{pr} \) appears to remain relatively constant at a level of about 1.7 J μmol\(^{-1}\). The gradual rise observed in the increase of \( R_g/Q_{pg} \) with \( \% R_p \) did not occur in the response of \( R_r/Q_{pr} \). At \( \% R_p \) between 20 and 80% there is little change in the proportion of \( R_r \) to \( Q_{pr} \), suggesting an increased absorption of PPFD relative to NIR by the canopy for increasing \( \% R_p \). A response of \( R_r/Q_{pr} \) to \( \% R_p \) appears only under overcast conditions (\( \% R_p < 20% \)) with the effect being enhanced at low solar elevation angles. It is speculated that the increase in reflected PPFD relative to reflected NIR radiation at low levels of \( \% R_p \) stems from the small amounts of NIR and the greater absorption of diffuse PPFD by the canopy (Russell et al. 1997a).

Again, seasonal effects are addressed using a contour plot of \( R_r/Q_{pr} \) against solar elevation angle and date for four different levels of \( \% R_p \), (figure 4.22). The obvious effect related to solar elevation angle is apparent at the bottom of all four panels. At solar elevation angles above 15 degrees, there is a mid summer depression in the values of \( R_r/Q_{pr} \) of between 0.2 to 0.3 J μmol\(^{-1}\). Values of \( R_r/Q_{pr} \) appear higher for
most other months, though a weak maximum may have occurred during spring. The mid summer reduction corresponds with the development of new shoots during June and July. New shoots reflect a greater proportion of PPFD radiation compared to NIR (Knyazikhin et al. 1997), likely because of their lower chlorophyll content (O’Neill et al. 2002). O’Neill et al. (2002) also report a greater absorption of radiation by first year Sitka leave at the wave bands just beyond the upper limit of PAR, though at longer wavelengths there was no significant difference for needle age.

![Contour plots of RR/Qpr against solar elevation angle and date for the years 1997 to 2001. Values of RR/Qpr were averaged after grouping by 5-degree bins for solar elevation angle and one-month bins for date. Values were exclude for periods of snow cover, when Rg < 40 or Qpg < 80, and averaged values were excluded if they contained less than five values.](image)

**Figure 4-22** Contour plots of $R_r/Q_{pr}$ against solar elevation angle and date for the years 1997 to 2001. Values of $R_r/Q_{pr}$ were averaged after grouping by 5-degree bins for solar elevation angle and one-month bins for date. Values were exclude for periods of snow cover, when $R_g < 40$ or $Q_{pg} < 80$, and averaged values were excluded if they contained less than five values.

### 4.8 Albedo and $Q_{pg}/Q_{pr}$ Ratio

Albedo is the ratio of reflected to global short-wave insolation and represents the absorptivity of the underlying surface; the corresponding value for PPFD is the ratio
The measurements used to determine these values are discussed in appendix A.1.1.2 and A.1.4.

Figure 4-23 Annual monthly mean (open circles) and five-year monthly mean (closed circles) Albedo (left panels) and $Q_{pq}/Q_{pr}$ (right panels).

The five-year, and annual, monthly average values of both albedo and $Q_{pq}/Q_{pr}$ are presented, figure 4.23. The five-year monthly mean values of albedo exhibit a spring minima of about 0.082, followed by a July maximum of 0.102. The annual pattern of $Q_{pq}/Q_{pr}$ is nearly identical, with corresponding minimum and maximum values of 0.027 and 0.035. The reduction in magnitude of $Q_{pq}/Q_{pr}$ by about 60%, with respect to albedo, results from the larger absorption of PPFD by the canopy (O’Neill et al. 2002; Russell et al. 1997b).

The annual patterns of albedo and $Q_{pq}/Q_{pr}$ are markedly different from the annual patterns of the corresponding radiation components. The annual pattern of albedo and $Q_{pq}/Q_{pr}$ exhibit small magnitudes through May after which both measures increase in magnitude by about 20%. This increase corresponds to the development of new shoots at the end of May and beginning of June. Annual deviations from the five-year mean roughly follow the patterns exhibited by the measured radiation
components, suggesting that the environmental conditions affecting radiation may also affect albedo.

![Graph showing relationship of albedo to solar elevation angle for different months.](image)

**Figure 4-24** Relationship of albedo (top panel) and $Q_{pg}/Q_{pr}$ (bottom panel) to solar elevation angle for different months. Error bars represent one standard error ($n>10$).

A commonly observed relationship of albedo is a diurnal pattern exhibiting a midday minimum and maxima during morning and evening. This pattern results from an increase in specular reflectance at low solar elevation angles (Dickinson 1983). Typically, this relationship is weaker for conifer canopies because of their greater canopy roughness (Jarvis *et al.* 1976). A similar relationship is observed at Griffin with both albedo and $Q_{pg}/Q_{pr}$ exhibiting a decrease with increasing solar elevation angles. To present this relationship (figure 4.24) it was necessary to reduce seasonal effects of shoot development by plotting the relationship of albedo and $Q_{pg}/Q_{pr}$ for each month. Winter months are not presented in figure 4.24 because of the smaller amount of data available to define the relationship after application of quality control restrictions. It is observed that both albedo and $Q_{pg}/Q_{pr}$ exhibit the expected decreases with increasing solar elevation angle.

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Figure 4.25 Relationship of normalized albedo and $Q_{pg}/Q_{tr}$ to solar elevation angle. Values are normalized by their corresponding monthly average value. Error bars represent one standard error (n>600).

To facilitate comparison of the changes of albedo and $Q_{pg}/Q_{tr}$ with solar elevation angle, values were normalized by their corresponding monthly means before determining the response curve, figure 4.25. As before, both albedo and $Q_{pg}/Q_{tr}$ decrease with increasing solar elevation angle. It is observed that the response of $Q_{pg}/Q_{tr}$ is only slightly more sensitive than albedo to solar elevation angle at high solar elevation angles. At low solar elevation angles, we might expect that the stronger cosine response of the short-wave sensor would cause a greater sensitivity of albedo to solar elevation angle. Instead, we observe a greater sensitivity of $Q_{pg}/Q_{tr}$, suggesting that the albedo of long-wave radiation may be reduced at low solar elevation angles. A slight peak is also observed in both albedo and $Q_{pg}/Q_{tr}$ at solar elevation angles greater than 50 degrees. It is known that the reflectance of conifer canopies is enhanced for measurement angles close to the solar elevation angle (Bicheron et al. 1997), suggesting a possible cause for the peaks observed at high solar elevation angles. To further investigate the differences in behaviour of albedo and $Q_{pg}/Q_{tr}$, the data are separated into four classes of $\% R_p$ and again plotted against solar elevation angle, figure 4.26.

A comparison of the responses of albedo and $Q_{pg}/Q_{tr}$ to solar elevation angle for different $\% R_p$ suggests that these responses are also a function of the quality of radiation. Ignoring values at solar elevation angles less than 20 degrees or greater than 50 degrees because of their possible confounding effects, we observe that at high levels of $\% R_p$ both albedo and $Q_{pg}/Q_{tr}$ decrease with increasing solar elevation
angle. As $\%R_p$ decreases, this relationship gradually diminishes so that at low $\%R_p$ both albedo and $Q_{pg}/Q_{pr}$ show little, or possibly even a slight positive response, to solar elevation angle. This is expected as the bi-directional canopy reflectance characteristics responsible for the decreases in albedo and $Q_{pg}/Q_{pr}$ with solar elevation angle become irrelevant under conditions of predominantly diffuse radiation.

![Graph](image)

**Figure 4-26** Relationship of normalized albedo and $Q_{pg}/Q_{pr}$ to solar elevation angle for four $\%R_p$ classes. Values are normalized with monthly means to remove seasonal effects. Error bars represent one standard error ($n>28$).

A second observation is that $Q_{pg}/Q_{pr}$ appears to be more sensitive to solar elevation angle than does albedo. A paired t test of the change in normalized albedo and $Q_{pg}/Q_{pr}$ with solar elevation angle supports this observation ($P = 0.001$, Power=0.050, alpha = 0.985), though only for conditions of $\%R_p$ between 75 and 100%. The smaller sensitivity of albedo to solar elevation angle is consistent with the greater canopy absorption of an increasingly larger NIR component of $R_g$ under clearer skies, as can be inferred from figures 4.21 and 4.12, respectively. With this in mind, we may review figure 4.26 and suggest that some of the discrepancy
between albedo and \( Q_{pg}/Q_{pr} \) at low solar elevation angles may be caused by the ratio of reflected to incoming NIR being less sensitive to solar elevation angle.

### 4.9 Absorbed and intercepted PAR

The amount of photosynthetically active radiation absorbed by the canopy (APAR) is proportional to the amount of radiation energy available to the canopy for photosynthetic processes. It may be expressed in the units of PPFD, \( Q_{pa} \), or may be expressed as the percentage of \( Q_{pg} \) absorbed by the canopy, \( Q_{pf} \). The values of \( Q_{pa} \) and \( Q_{pf} \) are not measured directly, but instead are calculated from the difference between \( Q_{pg} \) and the PPFD transmitted through the canopy \( Q_{pt} \). In this section, values of absorbed PAR will be presented as the fraction of \( Q_{pg} \) absorbed by the canopy, \( Q_{pf} \). Because of the great spatial variability of light beneath the canopy, it was necessary to provide sufficient spatial coverage to obtain an accurate representation of \( Q_{pf} \). The data provided in table 4.1 suggest that the number of samples measured (24) was less than those necessary (74) to obtain a resultant value of \( Q_{pf} \) within 10% of the site mean.

![Figure 4-27 Date/Time contour plot of the fraction of absorbed PAR, \( Q_{pf} \), for the Griffin site. Data incorporates all surface \( Q_{pa} \) sensor measurements](image)

Both the diel and annual patterns of \( Q_{pf} \) (figure 4.27) are noticeably different from those of the above canopy (figures 4.1) and below canopy (not shown) radiation components. Instead, \( Q_{pf} \) remains nearly constant over the diurnal period as well as over an annual period that spans from approximately June, to June of the following...
year. As with albedo, the annual pattern of $Q_{pf}$ is probably a result of the development of new shoots at the end of May through June, (McWilliam 1972) see section 8.3.3.3.

While the effect of shoot development upon albedo and $Q_{pf}/Q_{pg}$ is cyclic, it results in a steady increase in $Q_{pf}$ for canopies that are not already closed. This effect is shown in figure 4.28, in which separate curves of the monthly average values of $Q_{pf}$ are plotted against date for each of the three plot installations and the transect, (see figure 3.1). In figure 4.28, no effect of shoot development is observed in the closed canopy plot while obvious effects are seen in the open canopy plot adjacent to the eddy covariance tower.

![Figure 4-28 Monthly average fraction of intercepted PAR, $Q_{pf}$, in four plots at Griffin for the years 1997 through 2001. Error bars represent one standard error (n>800).](image)

Table 4.2 Annual (June – June) values of $Q_{pf}$ for four sensor installations. For each plot, values marked with an asterisk are not significantly different.

<table>
<thead>
<tr>
<th>Year</th>
<th>Eddy cov tower Open canopy</th>
<th>Profile tower Open canopy</th>
<th>Profile tower Closed canopy</th>
<th>Profile tower Transect</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>85.48</td>
<td>91.35</td>
<td>96.02 *</td>
<td>91.14</td>
</tr>
<tr>
<td>1998</td>
<td>91.25</td>
<td>93.63</td>
<td>96.03</td>
<td>93.34</td>
</tr>
<tr>
<td>1999</td>
<td>93.47</td>
<td>94.72</td>
<td>95.73</td>
<td>94.45</td>
</tr>
<tr>
<td>2000</td>
<td>94.58 *</td>
<td>95.93 *</td>
<td>96.33 *</td>
<td>95.66 *</td>
</tr>
<tr>
<td>2001</td>
<td>94.71 *</td>
<td>95.86 *</td>
<td>96.02 *</td>
<td>95.68 *</td>
</tr>
</tbody>
</table>
Statistical comparison of the June to June annual values (P < 0.001), table 4.2, revealed significant annual increases in the values of $Q_{pf}$ for canopies that had not closed. The data also suggest that closure is obtained at values of $Q_{pf}$ of approximately 96%, and that no increases in $Q_{pf}$ are observed after obtaining closure.

![Graph showing relationship between $Q_{pf}$ and %$R_p$ for five solar elevation angle classes. Error bars represent one standard error (n>50). (Statistical tests suggest that the increase at high solar elevations angles is not significant P < 0.05.)](image)

Figure 4-29 Relationship of $Q_{pf}$ to %$R_p$ for five solar elevation angle classes. Error bars represent one standard error (n>50). (Statistical tests suggest that the increase at high solar elevations angles is not significant P < 0.05.)

From the observed annual and diurnal behaviour of $Q_{pf}$ it might be expected that there is little or no variability in its magnitude other than that caused by canopy development. However, it is known that light penetration into the canopy is greater under more diffuse radiation conditions (Russell et al. 1988) suggesting that $Q_{pf}$ may also vary as a function of radiation quality. In figure 4.29, the response of $Q_{pf}$ to %$R_p$ is plotted for different solar elevation angle classes. At larger solar elevation angles, there is a slight increase in $Q_{pf}$ with increasing %$R_p$, but little effect at intermediate solar elevation angles. This suggests that direct beam radiation is more effectively absorbed by the canopy through reduced transmittance, reduced reflectance, or both. At small solar elevation angles, $Q_{pf}$ exhibits a maximum at intermediate levels of %$R_p$. This behaviour may be the result of reductions in $Q_{pf}$ at low %$R_p$ because of increased penetration (Russell et al. 1988)
and at high $\%R_p$ because of increased reflectance (Deering et al. 1999) (figure 4.26).

### 4.10 Net radiation

Net radiation, $R_n$, is the difference between down welling and up welling radiation of all wavelengths; encompassing both solar short wave and terrestrial long wave radiation. Similar to global short-wave radiation, it is defined as the net radiation energy received per unit surface area tangential to the earth’s surface.

![Figure 4-30 Five-year monthly means (filled circles-solid line) and annual monthly means (open circles-dotted line) of $R_n$, for each of the five years of the experiment.](image)

The patterns of monthly mean and interannual variability observed in $R_n$, (figure 4.30), are nearly identical to those observed for incoming solar radiation components. The peak value, 130 W m$^{-2}$ of the five-year monthly mean again occurs during the month of May. The minimum in $R_n$ is observed during December (-8 W m$^{-2}$) though net negative $R_n$ occurs during both the months of December and January. As with $R_g$, $R_n$ is above average during autumn, summer, late spring and

![Graph](image)

**Figure 4-31** Ratio of $R_n$ to $R_p$ averaged by month using all five years of data.

The pattern of $R_n$ as a proportion of total potential short wave radiation indicates that $R_n$ is smaller in the last third of the year than during the corresponding months at the beginning of the year, figure 4.31. It is seen in the following text that this difference results from increases in long wave radiation loss in the later part of the year.

![Graph](image)

**Figure 4-32** Diel patterns of monthly averaged $R_n$. Error bars represent one standard error (n>150).
The diel patterns of $R_n$, separated by month are shown in figure 4.32. The patterns are similar for most months, with a peak magnitude reached slightly after midday. To more closely examining the variability in monthly diel patterns of $R_n$, the curves from figure 4.32 were normalized to their diel ranges, split into net short wave ($R_g - R_l$) and net long-wave ($R_n - (R_g - R_l)$) components and plotted as contour plots against month and percent of diurnal period (figure 4.33). From this graph, it is clear that variations in the monthly patterns are driven primarily by variations in short wave radiation even though long wave radiation exhibited greater month-to-month variability. From this figure, we observe that the morning rise in $R_n$ is quite consistent throughout the year but that a later afternoon decline occurs during the months of February and August. The later diurnal decline in August also appears in the pattern of long wave radiation, though it persists into September and is probably related to greater thermal heat loss following the peak in the annual temperature cycle.

![Contour plot](image)

**Figure 4-33** Contour plot of normalized diurnal patterns of long wave (LW), short wave (SW) and $R_n$, by month and percent of diurnal period.

To examine the magnitudes of net short wave and net long wave radiation components we plot the diel patterns of monthly averaged values (figure 4.34). From the diurnal magnitudes, it is apparent that net long-wave radiation is approximately a factor of five less than that of short-wave radiation. The proportion of long-wave radiation is not constant, being smaller during the spring.
and summer months and greater during autumn and winter months. Such a pattern is consistent with the monthly average difference between air and soil temperatures, (see section 4.9.2).

Figure 4-34 Diel patterns by month of net loss of long wave radiation and net short wave radiation.

Although net long-wave radiation is function of sky/surface temperature differences (Stull 2000), Linacre (1969) found it to be well described by an empirical relationship with percent potential radiation and temperature. A similar empirical relationship for Griffin was not improved by the inclusion of temperature but the magnitude of short wave radiation did improve the relationship (figure 4.34).

\[ R_L = \ln(R_g - 66.9) \cdot \left( 1.512 + 0.164 \frac{R_g}{R_p} \right) \]  \hspace{1cm} (4.1)

It is suggested that, at high radiation levels, variations in radiation heating of the canopy are more closely related to variations in long wave radiation loss than are variations in temperature. At low radiation levels, this relationship is observed to decline and temperature will play a more important role in determining long wave radiation loss; however, a clear relationship to temperature could not be discerned from the analysis residuals.
4.11 Temperature

Temperature is a measure of the molecular kinetic energy of a system. For an ecosystem, this measure is closely associated with radiant energy because it forms the primary source and loss of energy to an ecosystem. The unequal distribution of temperature within an ecosystem results in the flow of energy from areas of high to low energy density, resulting in the characteristic patterns of temperature and heat exchange observed in nature. Temperature also has direct effects on evaporation, transpiration, photosynthesis, respiration, and the storage of heat in vegetation, air and soil. A description of the theory, materials, methods, and adjustments associated with the measurement of temperatures at Griffin are described in appendix A.2.

4.11.1 Above canopy air temperature

In contrast to those for radiation, the temporal patterns of air temperature, $T_a$, exhibit greater variation over the annual cycle than over the diel cycle. This difference in the temporal patterns of temperature, figure 4.36, are accompanied by a diel shift in the peak to about two hours after mid-day, and a similar annual shift to about one and a half months after mid-year. These peak shifts are caused by the
heat capacitance of the ecosystem, which behaves approximately like a first order
differential system such that the temperature response to radiation heating has an
inherent phase shift (Brock & Richardson 2001). Linacre (1992) explored
this relationship using climagrams and found a typical 30-degree phase shift in the
annual cycle of temperature compared to that of radiation, though this phase shift is
generally shorter at higher altitudes and further inland.

![Date/Time contour plot of above canopy air temperature, $T_a$, for the Griffin site.](image)

The annual climagram for Griffin, figure 4.37, is compared with a theoretical
climagram for a latitude of 56 deg, and a theoretical climagram with mean
temperature adjusted for site annual mean temperature (Linacre 1992). From this
figure, we observe that the effect of the oceanic climate of Britain has a strong
effect on temperature by creating a positive 6 °C offset, relative to the theoretical
model, and by reducing the observed annual temperature range. The temperature
range had a lower limit of approximately 3 °C, with temperatures during the months
of January through April nearly constant and higher than would be expected for a
more continental climate. During the months of June through August, temperature
was similarly lower than expected, probably as a result of the heat capacitance of
the surrounding ocean and because of greater cloudiness causing reduced radiation
in association with an oceanic environment.
Figure 4-37 Griffin climagram showing theoretical and actual lagged relationships between solar radiation and air temperature. Dashed curve is based on theoretical (Linacre 1992) temperature and radiation range at Griffin latitude. Solid curve is theoretical curve adjusted for higher mean temperature. The dotted line represents five-year monthly mean data and error bars represent one standard error. Each point is labelled by its month.

Figure 4-38 Griffin diel climagram representation showing actual and synchronized relationships between solar radiation and air temperature. Solid curve assumes no phase shift in temperature data. Error bars represent one standard error (n=1826).

A similar representation of the diel cycle, shown in figure 4.38 does not exhibit a similar lower limit to temperature. However, by assuming air temperature has reached a stable equilibrium at sunrise, it is found that over the diel period the phase shift in temperature gradually increases to about 25 degrees at midday and obtains a maximum of 35 degrees prior to sunset. During the nocturnal period, the phase
shift gradually moves back towards synchrony with radiation by tending towards a constant value. This diel pattern of phase shift is caused by the formation of an increasingly large atmospheric boundary layer capable of interacting with surface heat transfer and acting as a buffer volume. The nocturnal re-synchronization results from the thermal isolation of the surface layer following the breakdown of the atmospheric boundary layer.

Spectral representation is again used for a more in-depth look at the temporal variation. In figure 4.39, we can see the stronger annual and other long-term variations in temperature as compared to radiation ($Q_{pg}$). Most notably, there are strong weekly to monthly variations in temperature that are not observed in radiation. This suggests that synoptic scale features impact more strongly on temperature behaviour than on that of radiation.

![Figure 4-39 Five-year spectral representation of above canopy air temperature, top panel. Kaimal et al. (1972) model temperature spectra, (dotted line) are presented for comparison. The bottom panel is a repetition of the corresponding $Q_{pg}$ spectra for comparison.](image)

The key feature at shorter than diel scales, is a reduction in temperature variation relative to that observed for radiation, indicating that hourly variations in radiation
are much more significant that hourly variations in temperature. While close comparisons with the neutral Kaimal spectra can not be expected because of the wide range of conditions averaged, it is probable that the lack of comparison at high frequencies results from the inclusion of periods with low signal to noise ratios, for which white noise would have created larger than expected values of spectral power at high frequencies.

![Monthly Averaged Air Temperature](image)

Figure 4-40 Five-year monthly means (solid line) and annual monthly means (dotted lines) of above canopy air temperature, $T_a$.

Returning to the monthly means, figure 4.40 shows both the annual and five-year monthly means of air temperature for Griffin. Also shown in this figure is a comparison of the five-year monthly average air temperature with a similar long-term monthly average for 24 Scottish weather stations. Griffin follows a very similar annual pattern but is about 2 °C cooler than the Scottish average, which is expected based on the mainly coastal locations of the weather stations employed. Warmer than average periods occurred during the summer of 1997, the winter of 1998 and the fall of 2001 while colder than average periods occurred during the summer and fall of 1998 and during the winter and summer of 2001. It is notable that, although both radiation and temperature both have minima in December, the peak in radiation occurs in May while that of temperature occurs 2 and a half months later, between July and August. The shift in the peak is associated with the heat capacitance of the earth and ocean surrounding Griffin while the apparent lack of shift in the minima is caused by the strong effect of the heat capacitance of the
surrounding ocean and energy supplied by the Gulf Stream to the regional climatology as shown in figure 4.37.

The monthly averaged soil surface temperatures exhibit a pattern identical to that of air temperature. However, the thermal properties of soil result in a slight decrease in the amplitude (~15%), and an additional phase shift relative to $R_g$ (~10 degrees). These differences are larger (about 80% and 30 degrees) for diel patterns. Despite the lagged relation observed between radiation and temperature, no similar relationship existed between radiation and temperature for the monthly deviations from the five-year means. Linear regressions between the monthly deviations of $R_g$ and $T_a$ lagged from between 0 to 4 months resulted in correlations coefficients always less than 5%. This implies that factors controlling the variability of radiation differ from those causing the variation of air temperature, as can be inferred from the different patterns observed in the spectral plots.

4.11.2 Temperature profiles

Profiles of within and above canopy air temperature were averaged by year, (figure 4.41), and season, (figure 4.42). In each figure, the data are separated into eight plots, each plot showing averaged data for a fraction of the day/night period. The data in each profile curve are presented as values relative to the above canopy air temperature. Because of the static profile positions, temperature data for all five years are plotted against relative canopy height ($z/h_c$). The vertical axis in both plots use values of $z/h_c$ for the year 1999.

Some general observations on the behaviour of the canopy temperature profile can be drawn from these graphs. The air temperature in the bottom half of the canopy is always cooler than $T_a$ above mid-canopy, reflecting the greater radiation absorption higher in the canopy. The diurnal maximum in $T_a$ is observed near the top of the canopy while a nocturnal maximum occurs at greater than twice canopy height. A minimum temperature occurs at the surface during the day while nocturnally this minimum occurs at approximately 25% of canopy height. These temperature profiles result in convectively unstable conditions above the canopy and stable
conditions within the canopy during the day and unstable conditions in the lower half of the canopy and stable conditions in the upper half of the canopy and in the air above the canopy at night.

PROFILE AIR TEMPERATURES SEPARATED BY YEAR SHOWN IN FIGURE 4.41, INDICATE THAT THE BELOW CANOPY AIR TEMPERATURE BECAME COOLER DURING THE DAY (~1 °C) AND WARMER AT NIGHT (~0.3 °C) AS THE CANOPY ADVANCED TOWARDS CLOSURE. THIS IS CONSISTENT WITH THE REDUCED EXCHANGE OF SURFACE RADIATION CORRESPONDING TO CANOPY CLOSURE, SEE FIGURE 4.28. SIMILAR VARIATIONS IN ABOVE CANOPY TEMPERATURES ARE ALSO OBSERVED WITH A SLIGHT DIURNAL REDUCTION AND NOCTURNAL INCREASE DURING THE LATTER PART OF THE EXPERIMENT. IT IS POSSIBLE THAT THESE CHANGES ARE DIRECTLY RELATED TO THE CHANGES IN LOWER CANOPY TEMPERATURES.
Figure 4-42 Profiles of air temperature by season and percent of day/night. Temperatures are plotted as difference from above canopy temperature and heights are plotted as relative canopy height \( z/h_c \) for the year 1999.

The diurnal patterns of the temperature profile by season presented in figure 4.42, show that, during autumn and winter, diurnal heating of the canopy only occurs at the very top of the canopy. During spring and summer, greater solar elevation angles coincide with diurnal heating of the canopy down to about mid-canopy level. This will correspond with more stable conditions at the bottom of the canopy during the day in spring and summer than in autumn and winter. Nocturnally, temperature profiles behave similarly during all seasons with the greatest differences observed in the soil and not in the air column.

The characteristic shape of the soil temperature, \( T_s \), profiles with depth, shown in figure 4.43, result primarily from the thermal properties of the soil, in contrast to the air temperature profiles which were strongly affected by direct radiative heating and cooling of the canopy within the profile. The strong effect of soil heat on air
temperature is seen in the similar behaviour of the near surface air temperature and the upper soil temperature profiles.

Determination of the temperature lag with depth in the soil, (figure 4.43), indicate that the lag varies from approximately 3 hours at a depth of 0.025 m to a lag of about 19 hour at a depth of 0.3 m. From the lag of soil temperature behind air temperature, \( t_{\text{lag}} \), for a specified depth in the soil, \( z \), it is possible to determine (Rosenberg et al. 1983) the thermal diffusivity of the soil, \( \kappa \), as:

\[
\kappa = \left[ \frac{z}{2 \cdot t_{\text{lag}}} \right]^2 \cdot \frac{\pi}{P_r}
\]  

The estimated values of \( \kappa \), range from 0.03\( \times 10^{-6} \) m\(^2\) s\(^{-1}\) near the surface, increase slightly and then decrease to a value of about 0.12 \( \times 10^{-6} \) m\(^2\) s\(^{-1}\) at lower depths. The values at lower depths correspond well with the range of values of \( \kappa \) for dry peat soil of 0.10 \( \times 10^{-6} \) m\(^2\) s\(^{-1}\) and for moist peat of 0.13 \( \times 10^{-6} \) m\(^2\) s\(^{-1}\) (Monteith 1973; Van Wijk & De Vries 1963). Some of the discrepancy in the numbers obtained may result from lag time resolution of 30 minutes, and variability caused by variations in soil moisture and mineral content, both of which will increase diffusivity.

![Figure 4-43 Cross correlation (lag) of soil temperatures with above canopy air temperature using all available soil temperature measurements, left panel. Right panel, soil thermal diffusivities estimated from temperature lag.](image-url)
4.11.3 Boles temperatures

The primary purpose of measurement of bole temperature was to determine the storage of heat in the canopy biomass. Thus, temperature profiles were determined from the two depth measurements of $T_b$, figure 4.44. An attempt was made to determine thermal diffusivity from the lag of $T_b$ behind air temperature at different depths into the bole, using equation 4.2. These calculations gave a thermal diffusivity of $\kappa = 0.12 \pm 0.18 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$, which is lower than the values of Moore and Fisch (1986) for tropical forest but closer to the values for pine obtained by Meesters and Vugts (1996). However, the associated error was quite large because of poor resolution of the lag time (30 minutes) and inaccuracies in the depth of the temperature probe. Further analysis of the data also indicated weak spatial variation (both horizontal and vertical), suggesting that the air temperature, which was used to determine the lag, may have been a further source of error.

![Figure 4-44](image)

Figure 4-44 Average and standard error of bole temperature gradients expressed as degrees per decimetre as a function of time. All available data were averaged by month to obtain the curves shown. Error bars represent one standard error ($n>140$).

4.12 Humidity

Humidity measurements are necessary for water budget, and energy exchange and storage determination. In addition, humidity is a key component in the calculation of secondary parameters and corrections (eg. density corrections (Webb et al. 1980), virtual temperature determination, and calculation of the sensible heat flux.
coefficient) and plays an important roll in the exchange of carbon dioxide though its effect on stomatal functioning (Aphalo & Jarvis 1991; Leuning 1995). Despite its importance, sensor behaviour often presents difficulties in the measurement of humidity (Betts et al. 1997).

4.12.1 Above canopy humidity

The annual pattern of monthly averaged relative humidity, figure 4.45, shows a consistent minimum during the month of May, followed by a near linear increase to a maximum value during the month of December. Less humid conditions were observed during the autumn and winter of 1997 and much of the year 1998. More humid than average conditions occurred in the summer of 1997, spring and summer of 1999, summer and autumn of 2000 and the winter of 2001.

![Graph of Monthly Averaged Relative Humidity](image)

**Figure 4-45 Five-year monthly mean above canopy relative humidity (closed circles) and monthly mean relative humidity for individual years (open circles).**

Plots of five-year monthly averaged vapour pressure, $e$, vapour pressure deficit, $D$, and relative humidity, $h_r$ (figure 4.46) show that the $h_r$ peak in May is the result of a lag in the increase in the magnitude of vapour pressure relative to that of $D$. Also
noticeable is the strong increase in $D$ during the month of May, which can be associated with the corresponding peaks observed in radiation and temperature.

On a diel scale, the pattern of RH is controlled almost exclusively by the diel pattern of temperature, and hence $D$, as the diel variations in $e$ are relatively small, (figure 4.47). A plot of long term average vapour pressure obtained from the University of East Anglia’s gridded data set, New et. al. (1999). The patterns of the two vapour pressure curves agree well but the higher values of the UEA data reflect the primarily coastal, and lower elevation locations of Scottish weather stations used to derive this data set.

To examine the diel patterns of vapour pressure more closely, the five-year time average of vapour pressure for each month has been normalized by the corresponding five-year monthly averaged vapour pressure. These normalized values show the diel pattern of variation of vapour pressure as a percentage of the monthly mean vapour pressure. From these data, figure 4.48, it is observed that during the autumn and winter there is little variation in $e$ with only a slight mid-day rise in its magnitude.

**Figure 4.46** Five-year monthly mean values of vapour pressure deficit (open circles), vapour pressure (closed circles), and relative humidity (no symbol). Error bars represent one standard error ($n>6500$).
During spring, a strong early morning peak in $e$ developed, followed by a rapid decrease leading to a second, late afternoon, peak. It is suspected that the morning peak is caused by evaporation of dew into the surface layer precedent to the breakdown of the nocturnal inversion. A midday decrease in vapour pressure then results from atmospheric mixing associated with boundary layer development. During the evening, the vapour pressure rises with the breakdown of the boundary layer mixing and then decreases again as cooling causes dew formation on the canopy, (Betts et al. 2001). It is suggested that the dampening, during the summer, of the diel variations in vapour pressure observed in May results from drying of the ecosystem under conditions in which evapotranspiration and runoff losses exceed precipitation gains, see section 8.5.6.
Figure 4-48 Contour plot of the monthly averaged diel course of vapour pressure normalized with the monthly averaged vapour pressure ($e/e_0$).

4.12.2 Humidity profiles

Figure 4-49 Profiles of water vapour concentration by year and percent of day/night. Concentrations are plotted as the difference from above canopy Concentrations and heights are plotted as relative canopy height $z/h_c$. 

115
The defining characteristic of the humidity profile measurements, shown in figures 4.49, and 4.50, are the rapid within canopy increase of humidity to a maximum value at or near the surface. The within canopy increase in humidity is strongest during the day with average forest floor humidity of approximately 0.9 mmol mol$^{-1}$ greater than above canopy values. A nocturnal increase exists but is notable only in the bottom half of the canopy. There seems to be very little effect of canopy closure on the within canopy water vapour concentrations, though a slight dampening of the below-canopy, day-night concentration range may have occurred.

Humidity profiles, separated by season, are shown in figure 4.50. The diurnal profiles exhibit a strong seasonal dependence with lower canopy spring and summer water vapour concentrations approximately 0.5 mmol mol$^{-1}$ greater than autumn and winter values while nocturnal profiles show almost no seasonal variability. This difference may be related to the stronger diurnal heating of the canopy observed during spring and summer, see figure 4.42.
4.13 Carbon dioxide

Carbon dioxide measurements were taken for the determination of ecosystem exchange of carbon. Other carbon compounds (e.g. methane, carbon monoxide and volatile organic compounds) were not monitored, justified by their less significant role in atmospheric warming (IPCC Working Group 2001). The nature of the CO₂ monitoring instruments, their deployment and their lengthy period of installation made accurate measurements of CO₂ concentration, \( C_e \), difficult. Sensor precision, however, was such that accurate measurements of CO₂ variability over periods of seconds to days were obtained. For this reason, values of CO₂ in this section are presented as relative instead of absolute measures.

4.13.1 Above canopy CO₂

![Graph](image)

**Figure 4-51** Diel curves of five-year bi-monthly averaged deviation of CO₂ concentration from daily average concentration.

Bi-monthly average diel patterns of CO₂ concentration relative to the diel average CO₂ concentration, (figure 4.51), show the seasonal influence on diurnal draw down and nocturnal build up of CO₂ above the canopy. The period of May through August is characterised by large predawn build up and afternoon draw down of CO₂ concentration, both between 5 and 10 ppm. In contrast, the period from November through February sees almost no build up at night and only a slight midday draw down of CO₂ concentration. The large summer build up and draw down indicate the actions of both greater soil respiration, because of higher soil and biomass.
temperatures, and the increased photosynthesis, because of increased solar radiation.

4.13.2 CO₂ profiles

The vertical profiles of canopy CO₂ concentration, figures 4.52 and 4.53, are similar to those of water vapour in that they both show large increases in concentration close to the forest floor. Although both CO₂ and water vapour obtained peak profile values near the surface, the peaks in water vapour occurred during the day while those of CO₂ occurred at night. This difference in behaviour is caused by canopy gas exchange, which is a diurnal sink for CO₂ and a diurnal source of water vapour. As with water vapour, the effect of increasing canopy closure on CO₂ profiles over the three-year period of observation is small.

![CO₂ concentration profiles](image)

**Figure 4-52 Profiles of CO₂ concentration by year and percent of day/night. Concentrations are plotted as difference from above canopy concentration and heights are plotted as relative canopy height z/h_c.**

As with water vapour profiles, CO₂ concentration profiles show a strong seasonal dependence (fig 4.53). Unlike water vapour, which exhibits a stronger seasonal dependence during the day, the seasonal dependence of the CO₂ profile occurs...
primarily at night. The largest forest floor concentrations of CO$_2$ are observed during July through September and the smallest concentrations during January to March. This observation corresponds well with the known dependence of soil respiration upon soil temperature (Lloyd & Taylor 1994). It is also interesting to note that the nocturnal profile of CO$_2$ during the summer exhibits strong gradients up to the top of the canopy, whereas during the winter the CO$_2$ gradient appears proportional but of much smaller magnitude, resulting in near constant CO$_2$ concentrations in the upper half of the canopy.

Figure 4-53 Profiles of CO$_2$ concentration by season and percent of day/night. Concentrations are plotted as the difference from above canopy Concentrations and heights are plotted as relative canopy height $z/h_c$.

4.14 Wind speed

Wind provides the primary mechanism of the exchanges of mass and energy between a surface and the atmosphere as well as providing information about the character of the surface and the surrounding landscape. In this experiment, cup and
sonic anemometers were employed in the determination of mean horizontal wind speed, $U$. The use of these sensors is described in appendix A.5.

### 4.14.1 Above canopy wind speed

If examined closely, the data/time contour plot of average wind speeds shown in figure 4.54 show a year-to-year decline in mean above canopy wind speeds. Such a decline would be the result of forest growth relative to the static height of the wind velocity sensors. Calculations indicate that forest growth resulted in an annual average decrease of 0.2 m s$^{-1}$ for wind speeds measured at a height of 15.2 m. Diel patterns of wind speed were not affected by forest growth.

![Figure 4-54 Date/time contour plot of above canopy wind speed for Griffin.](image)

**Figure 4-54** Date/time contour plot of above canopy wind speed for Griffin.

![Figure 4-55 Five-year monthly mean above canopy wind speed (closed circles) and monthly mean wind speed for individual years (open circles).](image)

**Figure 4-55** Five-year monthly mean above canopy wind speed (closed circles) and monthly mean wind speed for individual years (open circles).
A second interesting feature, seen in both figures 4.54 and in the five-year and annual monthly averages shown in figure 4.55, is the anomalously low wind speed in January and February of 2001. This decline is believed to be associated with global circulation patterns and can be expected from the behaviour of the north Atlantic oscillation (Bojariu & Gimeno 2003), (see also section 8.3.2).

A comparison of the five year monthly average air temperature with a similar long term monthly average for seven Scottish weather stations is also shown in figure 4.55. The wind speeds at Griffin have a similar pattern but are between 3 and 4 m s$^{-2}$ lower than the weather station averages, again reflecting the mainly costal locations of the weather stations employed in the Scottish average. The annual variability in relation to the five-year mean wind speeds, figure 4.55, more clearly shows a peak in monthly average wind speeds during February and a minimum during the month of August. Additional smaller peaks appear to fall during the months of June and October, though a longer data set would be required to verify this observation.

Figure 4-56 bottom panel: Five-year monthly averaged, normalized wind speeds by percent of day. Top Panel: Monthly average diel range of wind speeds and mean wind speed (reversed scale).

To compare the five-year average monthly diel patterns of wind speed, figure 4.56, normalized values have been plotted against percent of diurnal/nocturnal period.
The resulting patterns indicate a consistent peak in wind speed at about 60% of the diurnal period. Morning increases follow shortly after sunrise and reach 90% of peak values by midday. The afternoon decline begins at about 80% of the diurnal period and reaches 40% of the diel range at sunset followed by a more gradual decline throughout the night.

The diel range of wind speed (also figure 4.56) is inversely related to mean monthly wind speed, suggesting that the midyear decrease in mean wind speeds results from a larger decrease in nocturnal $U$ than in diurnal $U$. It is believed that this occurs because of surface layer isolation resulting from the larger amount of energy required to accelerate the deeper summer boundary layer as it decays, (see also section 8.33).

![Figure 4-57 Five-year average wind speeds by wind direction for sonic anemometer and top sensor of cup anemometer profile. Error bars represent one standard error (n>250).](image)

The directional dependence of average wind speed (figure 4.57, bottom panel) shows the highest wind speeds to be from southerly and westerly directions and smaller speeds from north-easterly directions. This pattern is consistent with the wind climatology of the British Isles in general, (Chandler & Gregory 1976),
though the finer scale patterns about the southerly direction may be attributed to the presence of complex topography with the strong peak at 200 degrees and trough at 220 degrees corresponding to topographic low and high points (see figure 4.62). The different patterns observed between the 15.2 m wind speeds and samples from the profile tower suggest local effects in the measurement of profile wind speeds. Scaling the averaged wind speeds as \( \log(U/U_{\text{min}})/\log(U_{\text{max}}/U_{\text{min}}) \), (top panel of figure 4.57), revealed proportionally higher within canopy winds for easterly and westerly directions and proportionally lower winds for southerly directions. This is consistent with the planting row orientation (see figures 2.5 and 2.6) and the presence of an opening in that row near the profile tower. This result also suggests that profile measurements from this tower may not be representative of the 'average' canopy.

### 4.14.2 Wind speed profiles

Considerable information is available on the vertical profiles of wind speed in and above canopies (Allen 1968; Gardiner 1994; Green et al. 1995; Landsberg & Jarvis 1973; Thom 1971). Above the canopy, under neutral conditions, the wind profile will assume its common logarithmic shape. As the profile approaches the canopy, the logarithmic shape is distorted by the presence of the semi-permeable canopy; this has the effect of distorting the vertical scale at which the velocity approaches zero. Above the canopy the velocity profile approaches zero more quickly than expected for a log profile above a smooth surface while within the canopy the velocity approaches zero less rapidly.

A sample of the diel pattern of wind profiles on a moderately windy day is presented in figure 4.58, along with corresponding profiles of temperature and CO2. The profiles are presented as scalar values (three left panels) and as normalized difference values (three right panels). The scalar values show the diurnal increase of above canopy wind speeds and corresponding increases of temperature and CO2 both above and below the canopy. If the fluxes measured above the canopy persisted to the surface, the normalized profiles should collapse to a single profile. The normalized wind speed and temperature profiles are quite similar in shape indicating that transport processes are very similar for all periods presented. Their
non-logarithmic shape and dispersion result from the effect of the canopy on transport processes and the dispersion from the variability of sources and sinks within the canopy. The CO₂ profiles behave similarly above the canopy, but diverge within the canopy, with diurnal and nocturnal profiles obtaining different signs. This pattern reflects the disparity between above canopy and surface exchange of CO₂ with the surface acting as a constant source and the exchange above canopy indicating a source at night and a sink during the day.

Figure 4-58 Profiles of wind speed, temperature, and CO₂ concentration from 12-13 September 1997. The three left panels are profiles of scalar values while in the right three panels the values are presented as the difference from the top-most value and have been normalized using corresponding scaling values, \( u^* \), \( T^* \), and \( C^* \).

From the vertical profiles of wind speed, it is possible to obtain the canopy characteristic length scales of zero plane displacement, \( d \), and roughness length, \( z_0 \). Several approaches are available for estimating these values from wind speed profiles. The approaches that incorporate extrapolation of the log profile layer are sensitive to measured values of \( U \) and typically require several measurements in the
inertial sub layer above the canopy (Monteith 1973; Rosenberg et al. 1983). Estimation of parameters must be done within the inertial sub layer, the lower limit of which may be approximated as $d + 20 z_0$ (De Bruin & Moore 1985). Such measurements are not usually available over forests and other approaches have been developed that incorporate friction velocity or an estimation thereof (De Bruin et al. 1985; Lo 1995; Molion & Moore 1983).

An initial attempt at using log plot estimation of $d$ and $z_0$, employing only the top three levels of wind speed under neutral conditions produced unrealistic values of $z_0$ (0.06 m) and values of $d$ (8.1 m) that were higher than 75% of the canopy. The method of De Bruin and Moore (De Bruin et al. 1985) was then employed in the estimation of values of $d$ and $z_0$. This approach was applied to the profile data from 1997 with the restrictions that mean wind speed was greater than 4 m $s^{-1}$, and $|z/L| < 0.1$. The resulting values of $z_0$ and $d$, shown in figure 4.59, exhibited a distinct dependency upon wind direction.

![Figure 4-59 Estimated values of zero plane displacement for year 1997](image)

The two obvious regions of deviation occurred from 60 to 140 degrees and 200 to 280 degrees. Winds from these regions correspond with the planting rows and it is believed that the low values of $d$ result from the greater penetration of wind into the canopy. These regions were excluded from further analysis. The other variation observed is associated with high values of $d$ for southerly directions and lower
values for northerly directions. It is believed that this pattern may be associated with local topographic variability near the profile tower, with a north-facing slope that is slightly steeper than the site topography. Values of $d$ and $z_0$ were then calculated for both the growing periods starting in June of 1997 and June of 1998. The values for following years were not used because of data loss and because of poorer quality of upper level wind speeds because of canopy growth. Because of the expected wind direction dependency, and the variability in the number of samples from different wind directions, values of $d$ and $z_0$ were first averaged by 5-degree wind direction bins, which were than averaged to obtain a site average value, table 4.3.

Table 4.3 Values of zero plane displacement, $d/h$, and roughness length for the periods 6/97 to 6/98 and 6/98 to 6/99. Error represents one standard error.

<table>
<thead>
<tr>
<th></th>
<th>Zero plane displacement</th>
<th>$d/h$</th>
<th>Roughness length</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997-1998</td>
<td>4.68 ± 0.03</td>
<td>0.70</td>
<td>1.03 ± 0.01</td>
</tr>
<tr>
<td>1998-1999</td>
<td>5.50 ± 0.03</td>
<td>0.72</td>
<td>1.05 ± 0.01</td>
</tr>
<tr>
<td>Landsberg and Jarvis (1973)</td>
<td>0.88</td>
<td>0.34</td>
<td></td>
</tr>
<tr>
<td>Gardiner (1994)</td>
<td>0.79</td>
<td>1.4</td>
<td></td>
</tr>
<tr>
<td>Green et. al. (1995)</td>
<td>0.75</td>
<td>0.70</td>
<td></td>
</tr>
<tr>
<td>4x4 m</td>
<td></td>
<td>0.75</td>
<td></td>
</tr>
<tr>
<td>6x6 m</td>
<td></td>
<td>0.70</td>
<td></td>
</tr>
<tr>
<td>8x8 m</td>
<td></td>
<td>0.61</td>
<td></td>
</tr>
</tbody>
</table>

The resulting values of $z_0$ are intermediate of other observations of Sitka spruce, while values of $d/h$ compare better with the observations by Green et al. (1995) on widely spaced planting of Sitka. This later observation may be the result of the placement of the profile tower at a location in which the canopy was not closed. No attempt was made to determine stability effects on values of $d/h$ because of the observed occurrences of katabatic flow.

Katabatic, or drainage flows are known to occur on ground with slopes as gentle as 2 to 3 degrees (Mahrt & Larsen 1990). Models suggest that the development of katabatic flow depends on local stability, and is thus a function of heat loss (Horst & Doran 1986; Ye et al. 1990) while the intensity of the flow is a function of slope.

Figure 4-60 Profiles of wind speed, temperature, and CO₂ concentration from 24-25 September 1997. The three left panels are profiles of scalar values while in the right three panels the values are presented as the difference from the top-most value and have been normalized using corresponding scaling values, $u_*, T_*$, and $C_*$. Considering that the topographic slope at the Griffin site ranges up 10 degrees in the experimental area and up to 20 degrees within the catchment area, figure 2.7, it can be expected that such flows also exist at this site. Examination of wind speed differences between the profile’s top two anemometers revealed numerous periods of potential nocturnal drainage flow. An example time series of the profiles of $U$ under katabatic flow conditions, figure 4.60, indicate that steady state katabatic flow was achieved when the canopy cooled to more than 3 °C below the above canopy temperature, in agreement with the findings of (Horst et al. 1986). The characteristic katabatic flow velocity profile, with a maximum just above the canopy top, appears to weaken immediately after sunrise (Horst et al. 1986) but the
characteristic near logarithmic above canopy flow does not redevelop until 4 to 6 hours after sunrise. The normalized wind speed profiles do not obtain the common curve observed in figure 4.58. The change in sign of the profiles and greater dispersion indicating changes in the direction of transport of momentum.

Similarly, the profiles of $T_a$ and CO$_2$ in figure 4.60 are different from corresponding profiles under windy conditions, (figure 4.58). Profiles of CO$_2$ and $T_a$ exhibit a greater difference between canopy bottom and top (30 ppm and 3 deg) then observed under windy conditions (10 ppm and 1 deg). Both profiles exhibit a layer of homogeneous conditions, which differ from above canopy conditions, for up to half canopy height under the example katabatic flow conditions. Such profiles will produce sharper gradients of CO$_2$ and temperature, near the top of the canopy, than are observed under windier conditions.

4.15 Wind direction

Wind direction, $\theta$, along with wind speed, characterise the flow of air at a location. This section discusses the behaviour of wind direction measurements and how they are influenced by environmental conditions.

The variation in wind direction has been observed to be a function of wind velocity, figure 4.61 (Davies & Thomson 1999). An equation describing the relationship between the standard deviation of wind direction, $\sigma_\theta$ and $U$, using data over non-complex topography, has been developed by Hanna and Jolle (Hanna 1981):

$$\sigma_\theta = \sqrt{\sigma_0^2 + \left(\frac{b_0}{U}\right)^2}$$

(4.3)

In this model, the coefficient $a_0$ represents the wind direction variability resulting from surface layer turbulence while the term $b_0 / U$ represents the contribution of larger scale variability in wind flow.

Equation 4.3 did not appropriately describe the data from Griffin with the model values of $\sigma_0$ being too small at low wind speeds. This poorer fit of the Hanna and
Jolle model resulted in average $R^2$ values of about 0.5 for the model fit to data from 30-degree wind direction bins. Employing an exponential decay model:

$$\sigma_\theta = a_0 + b_0 \cdot e^{-c_0 \cdot \theta}$$  \hspace{1cm} (4.4)

in which the coefficient $a_0$ may again be used to represent the effect of surface layer turbulence while the larger scale variability is now represented by the term containing the coefficients $b_0$ and $c_0$. This model provided a better correlation between model and data, see table 4.4.

![Figure 4-61 Average wind direction standard deviation for wind velocities between 0 and 6 m s$^{-2}$. Averages were separated into 30-degree wind direction bins. Error bars represent one standard error (n>5).](image)

The asymptote values ($a_0$) obtained with both models are more than twice as large as those observed by Davies and Thomson (1999), a result consistent with the rougher canopy from this experiment. In addition to providing a better fit, equation 4.3 was also capable of capturing the change in the relationship for winds coming from wind directions between 180 and 300 degrees, which exhibited higher values of $\sigma_\theta$ for wind speeds between approximately 0.3 and 2 m s$^{-1}$. Considering the greater topographic complexity of the Griffin site, it may be inferred that the greater
variability at moderate wind speeds suggests that an additional term accounting for topographic complexity is required.

Table 4.4 Regression coefficients for the fit of equation 4.4 to $c_0$ for 30 degree wind direction bins, using five years of data from Griffin.

<table>
<thead>
<tr>
<th>Wind Direction</th>
<th>Coef $a_0$</th>
<th>Coef $b_0$</th>
<th>Coef $c_0$</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>15</td>
<td>21.97</td>
<td>85.46</td>
<td>1.99</td>
<td>0.79</td>
</tr>
<tr>
<td>45</td>
<td>24.42</td>
<td>89.98</td>
<td>2.34</td>
<td>0.80</td>
</tr>
<tr>
<td>75</td>
<td>21.56</td>
<td>90.27</td>
<td>2.09</td>
<td>0.84</td>
</tr>
<tr>
<td>105</td>
<td>20.57</td>
<td>89.71</td>
<td>1.89</td>
<td>0.85</td>
</tr>
<tr>
<td>135</td>
<td>16.05</td>
<td>90.85</td>
<td>1.53</td>
<td>0.80</td>
</tr>
<tr>
<td>165</td>
<td>16.71</td>
<td>94.34</td>
<td>1.62</td>
<td>0.77</td>
</tr>
<tr>
<td>195</td>
<td>18.14</td>
<td>80.08</td>
<td>1.05</td>
<td>0.82</td>
</tr>
<tr>
<td>225</td>
<td>23.10</td>
<td>74.47</td>
<td>1.04</td>
<td>0.78</td>
</tr>
<tr>
<td>255</td>
<td>22.05</td>
<td>80.16</td>
<td>1.20</td>
<td>0.79</td>
</tr>
<tr>
<td>285</td>
<td>17.37</td>
<td>79.98</td>
<td>1.10</td>
<td>0.83</td>
</tr>
<tr>
<td>315</td>
<td>17.70</td>
<td>80.75</td>
<td>1.33</td>
<td>0.83</td>
</tr>
<tr>
<td>345</td>
<td>18.38</td>
<td>85.90</td>
<td>1.66</td>
<td>0.82</td>
</tr>
</tbody>
</table>

Figure 4-62 Panoramic map of elevation and distance for different directions as viewed from the flux tower.

For winds from the region of steepest topography (figure 4.62), we observe a decrease in both coefficients $b_0$ and $c_0$. A simultaneous reduction in both of these coefficients has the effect of increasing the variability at moderate wind speeds while maintaining a similar variability at low and high wind speeds. We can infer
that complex topography does not affect the variability of wind direction at low wind speeds or high wind speeds but does increase it at moderate wind speeds. More sites will need to be evaluated to determine the magnitude of the variability increase and how far it extends into higher velocities with increases in topographic complexity.

Figure 4-63 Frequency distributions of wind direction by wind speed (separate panels) and by percent of the diurnal/nocturnal period (see legend).

The topographic complexity of the Griffin field site also contributed to the patterns observed in the run mean wind directions. The five-year wind direction distributions (not shown) have distinct peaks associated with the orientation of the valley in which the experiment site is situated. While such an average distribution
was expected, a more interesting insight was obtained by examining the wind direction distributions for different levels of wind velocity and fractions of the diel period, figure 4.63.

At wind speeds below 1 m s\(^{-1}\), we observe flow to come predominantly down the valley at night and up the value during the day, though down slope flow appears not to diminish until the first 25% of the diurnal period had passed. At wind velocities between 1 and 2 m s\(^{-1}\) the enhanced down slope flow appears to remain at night but the diurnal flow appears to have begun to obtain a distribution more similar to that at higher wind speeds. At higher wind speeds, we observe strong similarity between the diurnal and nocturnal wind direction distributions. Interestingly, however, as wind speeds increase beyond 3 m s\(^{-1}\) we observe both the diurnal and nocturnal down valley flows to develop an additional distribution peak at 200 degrees associated with the southerly topographic gap seen in figure 4.62, which appears to be funnelling wind flow only at higher wind velocities.

Figure 4-64 Wind direction frequency distribution (hundredths of a percent) by month. Distributions incorporate all available data.

The annual patterns of wind direction shown in figure 4.64 demonstrate the regularity of the wind directions associated with along valley flow. More frequent
occurrence of southerly flows are apparent for the months of April through August. More northerly flows develop from September, reaching a maximum in February, and have associated with them flow through the topographic gap at 200 degrees.

4.16 Precipitation

While drought was not a factor during the period of this experiment, the monitoring of precipitation played important roles in verifying the hydrologic budget of the experimental site and explaining the sites energy and carbon interaction. The primary measurement was that of gross precipitation, $P_g$, which described the input to the site of liquid water in the units of millimetres of precipitation during each half hour run. For shorter periods during the experiment, canopy precipitation stem flow and through fall were available to complement $P_g$.

4.16.1 Precipitation

![Figure 4-65 Date/time contour plot of precipitation for the Griffin site.](image)

As can be observed in the date/time contour plot of precipitation, figure 4.65, the occurrence of precipitation appears quite random when compared with similar figures for variables such as radiation and temperature. However, a pattern of greater autumn and winter precipitation is observed when considering the five-year monthly mean values of precipitation, figure 4.66, though year to year variability remains large in relation to this pattern. The annual pattern and magnitude of five year monthly average precipitation at Griffin also compares well with the average
of 19 Scottish weather stations, though the spatial variability of precipitation associated with these stations is quite large.


**Figure 4-66** Five-year monthly precipitation (closed circles) and monthly average precipitation for individual years (open circles) for data obtained from the automated precipitation gauges.

**Figure 4-67** Five-year bi-monthly averages of precipitation by time of day. The solid line without symbols indicates the magnitude of the average standard error for the set of curves.
A diel pattern of precipitation is observed that changes seasonally, (figure 4.67). These diel patterns of average run-total precipitation have diurnal values that are nearly twice as large as the corresponding nocturnal values during the months of January through August. During the months of September/October, precipitation is heavier and falls more evenly throughout the day, and in the months of November and December precipitation remains heavy but exhibits an inverted pattern with nocturnal precipitation being greater than diurnal precipitation. The cause of this annual change in pattern is not known but it may indicate a shift away from convection related precipitation during autumn months.

Two additional sets of statistics of interest in the estimation of canopy interception and precipitation runoff are the temporal extent of precipitation events and periods between precipitation events. These values, figure 4.68, may be used in models which require information on the temporal variability of precipitation (Zeng, et al. 2000). The data presented in figure 4.68 show that the typical length of rain events remains quite constant over the year while the median length of the interval between events is generally longer by about an hour during autumn and winter and 2 hours during spring and summer. This longer time interval between rain events in the summer accounts for the corresponding reduction in the number of precipitation events.
events. Perhaps what is most notable about the data presented in figure 4.68 is the generally short length of precipitation events. This characteristic is likely associated with resolution problems of the automated precipitation gauges in detecting the continuous light precipitation often observed at this site. A more accurate precipitation occurrence detector would be required to obtain a higher quality statistic.

4.16.2 Throughfall, stem flow, and stream flow

It is known that the storage capacity of the canopy responds to both environmental conditions and canopy structure (Llorens & Gallart 2000; Rutter et al. 1975). However, because of the coarse time scale of collection of throughfall and stem flow amounts, it was not possible to investigate the dynamic nature of canopy interception. The results presented here therefore refer to the static relationships for weekly to biweekly average values.

A linear relationship of throughfall to precipitation has been observed for several forest types, (Huber & Iroume 2001; Llorens et al. 2000; Teklehaimanot & Jarvis 1991). A similar relationship is observed for the Sitka spruce canopy at Griffin, figure 4.69, which indicates that throughfall is approximately 75% of precipitation, similar to observations by (Hancock & Crowther 1979; Hutchings et al. 1988). Using figure 4.69, and following Massman (1983), we may infer the storage
capacity of the canopy from the intercept of the linear regression with the precipitation axis. This method of estimation gives an equivalent precipitation depth value of 2.9 mm for the canopy storage capacity. This value is slightly higher than the range of storage capacity values observed for sitka and other species (Llorens et al. 2000). It must be noted, however, that this approach was meant for use with fine temporal resolution measurements of single precipitation events and will not be strictly valid for period averaged precipitation-throughfall relations.

\[ y = a \left(1 + \frac{x}{x_0}\right)^b \]

\[ a = 0.867 \]
\[ b = -1.728 \]
\[ x_0 = 34.17 \]

Figure 4-70 Regression of total stem flow against total precipitation for throughfall/stem flow collection periods. Error bars represent one standard deviation (n=24).

Stem flow for a specific forest stand has also been observed to increase linearly with the amount of precipitation (Hanchi & Rapp 1997). The data from Griffin forest, figure 4.70, exhibit similar tendencies with approximately 1% of precipitation reaching the ground as stem flow. However, the lack of data from high precipitation periods because of receptacle overflow results in the observed non-linearity and prevents a more detailed estimate.

The relatively small size of the catchment surrounding the Griffin experiment site, and drained by Culullich burn, resulted in the close correspondence between precipitation and stream flow observed in figure 4.71. Viewed simplistically, the behaviour of stream flow at this site is similar in nature to that of precipitation.
interception in that the site has a certain capacitance for water storage, in the way the canopy has a storage component. For the moment, if we ignore the site characteristics that may affect stream flow associated with specific precipitation events, we find that monthly total stream flow is approximately linearly related to monthly total precipitation at precipitation levels greater than about 40 mm, figure 4.72.

![Graph](image1.png)

**Figure 4-71** Sample time series of run total precipitation (grey bars) and catchment stream flow loss.

![Graph](image2.png)

**Figure 4-72** Relationship of monthly total stream flow to monthly total precipitation.
4.17 Atmospheric pressure

Atmospheric pressure, $P$ (units of kPa), represents the downward force exerted by the mass of the overlying air column. Its value is required in many of the calculations associated with unit conversions and flux calculations. Unless otherwise stated, the values presented here have not been reduced to their equivalent sea levels.

Mean air pressure at the Griffin experiment site is approximately 97 kPa. This value did not change for the years 1997 through 2000 but was slightly higher in 2001. No changes to the data logging system or sensor could be related to this increase in 2001. Variability on seasonal time scales was also observed, (figure 4.73). A 1 kPa decrease in $P$ is observed in winter. This annual pattern exhibited better correspondence with solar elevation than it did with air temperature.

This seasonal pattern was consistent with regional pressure patterns of 1970–2000, 0.5 degree grid interpolated sea level pressure (Basnett & Parker 1997) and long term average monthly pressure from six Scottish weather stations. In this comparison, Griffin air pressure data were converted to sea level pressures using monthly mean air temperatures and the hypsometric relationship between height and pressure (Wallace et al. 1977).
Analyses of run mean pressure in relation to environmental variables generally showed weak correspondence. However, the deviation of Griffin pressure from the 1970-2000 average does appear similar to the ratio of $R_s$ to $R_p$, (see figure 4.3), suggesting a link between pressure and cloudiness. At shorter time scales, weak relationships were found between environmental conditions and pressure change for periods less than 3 hours. To determine what time scales of variation were important for pressure fluctuations, spectra were calculated for run mean atmospheric pressure for the period November 1998 through December of 1999.

![Figure 4-74](image)

**Figure 4-74** One year spectral representation of above site air pressure, bottom panel. The top panel is a repetition of the five-year air temperature spectra for comparison.

The peaks in pressure spectra, figure 4.74 bottom panel, are distinctly different from those of air temperature (top panel). The peak in pressure fluctuations, at between 6 to 15 days, likely represents the passage of synoptic scale systems. Above this peak, a rapid drop off of variability at high frequencies suggesting that variations in pressure do not have the diurnal and smaller scale relations to radiation that exist in the temperature oscillations.
4.18 Soil moisture

The shallow rooting depth of the Sitka spruce (Henderson et al. 1900; Nieuwenhuis et al. 2003) and their sensitivity to plant water status (Silim et al. 2001; Watts et al. 1976) suggests that soil moisture could play an important role in their functioning. Although precipitation is plentiful throughout most of the year, periods of low precipitation could easily lead to soil drought conditions for trees that have developed under a predominantly wet environment. Time domain reflectometry, TDR, supported by gravimetric samples were employed to monitor soil water content, $\theta$, in this experiment.

![Image of soil moisture and precipitation data]

**Figure 4-75** Volumetric soil moisture (solid and dashed lines) and precipitation (dots) for January through March 2001.

Soil moisture measurements exhibited the expected responses to large precipitation inputs, *i.e.* a step increase and exponential decay (Deeks et al. 2004; Hewlett 1982), (figure 4.75). For smaller precipitation inputs, the resolution of the soil moisture probes was insufficient to identify related patterns of soil moisture.
Figure 4-76 Soil volumetric water content time series measured by TDR.

The time series of the available $\theta_v$ data shown in figure 4.76 demonstrate the strong annual variability of soil moisture at Griffin. The soil moistures measured at two locations have annual ranges of 20 and 40%. This annual range is, however, equivalent to the variation of 30 to 40% observed between these two measurement locations. A more extensive set of $\theta_v$ values is needed to obtain an accurate measure of Griffin soil moisture.

Figure 4-77 Monthly average soil moisture (filled circles) and half hour precipitation (open circles) for 2001. Error bars represent one standard error ($n>1000$).
Although soil moisture showed a close relationship to precipitation in figure 4.75, the annual pattern of soil moisture does not follow the annual pattern of precipitation as closely. As seen in figure 4.77, the pattern of soil moisture is proportionally greater than that of precipitation from November through approximately April. That the soil moisture is higher than expected from precipitation inputs during these months is a result of lower evaporative demand. This will be covered further in chapter 8.

4.19 Soil heat flux

Soil heat flux, $G$, is the transfer of sensible heat into or out of the soil at the soil surface. As with radiation, soil heat flux represents the transfer of energy and not its quantitative state. It is presented in this section because its measurement methods are similar to those used for climate state variables. The flux of heat into the soil may be estimated by measuring soil temperature profiles and inferring the heat transport down that profile based on soil properties (Kimball & Jackson 1999; Rosenberg et al. 1983). A more common method is to measure the heat flux using soil heat flux plates. These devices are sensitive thermopiles that are capable of measuring the temperature difference between their top and bottom sides. Because the plates have a known conductivity, the heat transfer through the plate will be proportional to the temperature difference across the plate. Soil heat flux plates measurements assume that the thermal conductivity of the soil is identical to that of the plate. Errors in this method may exist if the soil is of different conductivity than the plate. Errors may also be caused by vapour transport from below the level of the plate, advective transfer of heat by air or water, and because of spatial variability in soil properties or radiative forcing (Kustas et al. 2000; Mayocchi & Bristow 1995).

The annual monthly means of soil heat flux, figure 4.78, indicated that transfer of heat into the soil begins in April and continued through September. A peak transport into the soil of 3.0 W m$^{-2}$ was observed in July and a maximum average loss of heat from the soil of 2.7 W m$^{-2}$ in November. Because temperature gradient determines $G$, equation 4.5, the pattern of soil heat flux corresponds closely to that
of air temperature for both annual and diel patterns, though the pattern of $G$ is inverted because of sign convention.

![Figure 4-78 Five-year monthly means (solid line) and annual monthly means (dotted lines) of soil heat flux, G.](image)

When available and in the absence of soil heat flux plates, soil temperature profiles were used to estimate values of soil heat flux using the relationship,

$$G = K \frac{\partial T_z}{\partial z}$$

in which $K$ is the thermal conductivity of the soil.

It is known that values of $K$ vary with soil type and soil moisture. Because the soil temperature profiles and soil heat flux plates were installed in close proximity, it was assumed that soil composition and soil moisture was similar for both the measurements of $G$ and $T_z$. With this assumption, it was possible to determine values of $K$ for different soil moisture conditions by rearranging equation 4.5.
Average values of $G$ and $\partial T_r/\partial z$ were used to calculate $K$, which was then grouped by 5% bins of volumetric soil moisture content. The median value for each bin was chosen as the representative value of $K$, figure 4.79.

![Graph](image)

**Figure 4-79** Soil thermal conductivity calculated from site average soil heat flux and temperature difference between the 5 and 6 cm depths.

The resulting soil thermal conductivities are close to the conductivities of water (0.586 W m$^{-1}$ K$^{-1}$) at high soil water contents. At lower soil water content, the conductivities fall to values between 0.28 and 0.35 W m$^{-1}$ K$^{-1}$. These values are slightly higher than the values specified for peat soil (De Vries 1975; Monteith 1973), consistent with water content (Kellner 2001) and increasing mineral content at this depth, see figure 2.22.

A comparison of open and closed canopy soil heat flux indicated that closed canopy soil heat flux was significantly more positive ($n=24452$, $p<0.001$), though the difference in medians was small, table 4.5. Contrary to expectations, the quartile range observed in the closed canopy was greater than that observed in the open canopy. The cause of the greater range of soil heat flux under a closed canopy is likely an effect of under sampling, see table 4.1, as the number of samples taken would have been insufficient for determining the small expected difference between open and closed canopy locations. As an example, paired comparisons of individual plates revealed significant differences between all but two comparisons.
Insufficient information, on the physical characteristics of locations, was available to determine the cause of these differences.

Table 4.5 Median and quartile values of average open and closed canopy soil heat flux values.

<table>
<thead>
<tr>
<th>Canopy</th>
<th>Median</th>
<th>25% quartile</th>
<th>75% quartile</th>
</tr>
</thead>
<tbody>
<tr>
<td>Open canopy all</td>
<td>0.165</td>
<td>-1.847</td>
<td>2.170</td>
</tr>
<tr>
<td>Closed canopy all</td>
<td>0.218</td>
<td>-2.350</td>
<td>2.800</td>
</tr>
</tbody>
</table>

4.20 Summary

In this chapter, the temporal and spatial patterns of climate variables and their variability are analysed and explained in relation to controlling factors. Where possible results are examined in light of and comparisons made with other research findings.

The annual patterns of incoming and reflected radiation ($R_g$, $Q_{pg}$, $R_r$, $Q_{pr}$, $R_n$) have an approximately 15% larger than expected value in May, which may be related to seasonal variations in atmospheric circulation patterns or reduced land-ocean temperature difference. Average diel patterns are quite consistent over the year, although an afternoon decrease in cloudiness is observed on cloudy days. The variability of global radiation is similar throughout the year, with strong peaks at diel and annual periods and weak peaks at periods of 4 and 21 days.

The temporal patterns of other radiation-derived ratios (Albedo, $Q_{pr}/Q_{pg}$, $R_g/Q_{pg}$, $R_r/Q_{pr}$) behave differently than incoming and reflected radiation. These ratios are not determined by the annual or diel solar cycles. The values of $R_g/Q_{pg}$ range from 0.53 to 0.59 μmol J$^{-1}$, decreasing with elevation angle and cloudiness while $R_r/Q_{pr}$ ranges from 1.5 to 1.6, being constant under cloudy conditions and increasing with solar elevation under clearer conditions. Values of albedo (0.08 to 0.13) and $Q_{pr}/Q_{pg}$ (0.024 to 0.044) both decrease with solar elevation angle, but while albedo shows little effect of cloudiness, $Q_{pr}/Q_{pg}$ has more variability under cloudy conditions. Unlike the other radiation ratios, albedo and $Q_{pr}/Q_{pg}$ show a response to the growth and development of the canopy. Similarly, values of $Q_{pf}$ are closely tied
to the canopy, reaching a maximum value of about 96% at the time of canopy closure, and shows slight decreases with increasing percent of $R_p$.

Further analysis of radiation showed a prevalence of mostly cloudy and mostly sunny condition while partly cloudy, very cloudy and very sunny conditions were less frequent. The concept of sunshine (i.e. clear sky conditions) best represented as a percent of $R_p$ than by threshold value of $R_g$. Diffuse radiation was also modelled using a function of percent of $R_p$.

The annual patterns of temperature and humidity are similar to that of radiation, although they lag behind the pattern of radiation. The annual pattern of temperature is delayed by 1.5 months and is limited in range by the influence of the oceanic climate of Britain. The diel temperature pattern lags behind that of radiation, however it has a variable phase shift because of boundary layer development. Although temperature variations are inherently tied to radiation, the variability of temperature is larger than that of radiation at annual, monthly and weekly time scales; there is also larger variability at short time scales (minutes to seconds). Within-canopy profiles of temperature showed that air was cooler within the canopy than it was at canopy top, however at night and during autumn soil surface temperatures were often similar to or warmer than canopy top temperatures.

Annual patterns of wind speed are unrelated to available energy and appear to be more closely tied to global circulation patterns. The average diel patterns of wind speed, when normalized for mean wind speed magnitude, are regular throughout the year. Detailed analysis of wind speeds showed a directional variation associated with large-scale topography and canopy architecture (row planting). Wind direction variability also showed a directional dependence associated with large-scale topography. Wind and scalar profile measurements provided evidence of katabatic flow occurring above the canopy; this observation was supported by diel variations in wind direction. Wind speed profile measurements also provided estimations of zero plane displacement and roughness length, which were in agreement with other research findings.
Annual patterns of precipitation, although more erratic than those of other variables, indicate higher levels of precipitation in autumn and winter. The seasonal change in diel patterns show low diel variability in autumn and higher afternoon precipitation during the remainder of the year. The increase in autumn precipitation is primarily a result of increased nocturnal precipitation. The length of precipitation events is similar throughout the year while the time between events is longer during the summer. Estimates of canopy storage capacity were estimated from measurements of precipitation throughfall and stem flow and were found to be in agreement similar research. Soil moisture shows a close relation to precipitation on short time scales by less closely related on annual time scales.

Annual soil heat flux follows a pattern similar to that of air temperature. Significant variability in soil heat flux between locations indicates that greater sampling may be required. Estimates of soil thermal conductivity as a function of volumetric soil water content were estimated from soil heat flux and soil temperature profiles.
5 Ecosystem exchange measurement methods

5.3 Introduction

To understand the behaviour of an ecosystem it is important to know its resource requirements. Though numerous resources are mobilized in the growth of vegetative systems, three environmental variables (solar radiation, water and carbon dioxide) determine to a large degree the persistence of vegetation because of their role in the photosynthetic process. The level of current understanding of photosynthesis is such that models reasonably represent the formation of organic matter by vegetation (Farquhar et al. 1980; von Caemmerer & Farquhar 1981). Other models have attempted to represent the growth of entire plants or of entire ecosystems based on physical and empirical photosynthetic modelling concepts e.g. (2000; Amthor et al. 2001; Sellers et al. 1996; Wang & Jarvis 1990). Such models of ecosystem growth are important in the extrapolation of results from studied ecosystems in both time and space. Although useful in understanding the development of plants and ecosystems, these models do require high quality information about the soil and atmospheric environment of the system under study, as they represent the resource base of the vegetation. While such ecosystem models may be keys to our understanding, they also represent immense simplifications of systems containing an almost infinite variability of states and processes. To capture in its entirety the variability existing within an ecosystem would certainly prove unfeasible. This problem reflects itself in the difficulty in scaling up (Baldocchi 1993; Farquhar & De Pury, 1997; Jarvis 1995) small-scale exchange estimations, such as chamber (Davidson et al. 2002; Lavigne et al. 1997; Leuning & Foster 1990) and sap flow (Cienciala et al. 1999; Grime et al. 1995; Lundblad et al. 2001) techniques. Instead, it has been more fruitful to study such ecosystems en masse by measuring their interaction with their environment (Baldocchi et al. 1996; Running et al. 1999). By understanding such interactions, we are able to establish boundary conditions within which models of ecosystem and plant behaviour must reside. In addition, and irrespective of their benefit to modelling, such measurements have the capacity for explicit determination of average ecosystem behaviour with regard to resource utilization. Thus we are led to the point at which we
must determine appropriate methods to measure the interaction of an ecosystem with its environment.

First we must define a practical boundary within which lies the ecosystem of interest and across which mass and energy must move in order to interact with the ecosystem. These boundaries are typically a function of the method used to measure the transfer of mass and energy and therefore it is important to select a method appropriate for the scale of the ecosystem under study. In selecting an appropriate method for the ecosystem studied in this thesis, it is necessary to conceptualise the nature of the ecosystem’s environment.

Our ecosystem can be roughly divided into soil, vegetation and atmosphere. The soil and underlying geology forms a lower boundary, which changes slowly but can be difficult to monitor over the short time scales of most experiments. The vegetation lies ‘between’ the soil and atmosphere, extracting and transferring materials between earth and sky. It is more active than the soil over experiment time scales, making it of key interest, and simpler to monitor. The atmosphere forms a highly mobile and relatively homogeneous upper boundary via which many of the ecosystems resources are transferred. The atmosphere’s character makes monitoring of its lower boundary simple though it is necessary to select a position and methods relevant to that position for measuring the atmospheric transport of resources to and from the underlying ecosystem. Some understanding of the lower atmosphere is required to allow a wise choice of measurement techniques. The region of the atmosphere with which we are concerned in this thesis is called the atmospheric boundary layer, ABL. It is defined as ‘the part of the troposphere that is directly influenced by the presence of the earth’s surface, and responds to surface forcings on a time scale of one hour or less’, (Stull 1988).

In the most general sense, the boundary layer may be divided into four sublayers; the interfacial layer, the roughness sublayer, the surface layer, and the outer layer, as shown schematically in figure 5.1. The tropospheric layer overlying the ABL is referred to as the free atmosphere. It is well mixed and characterized by geostrophic winds and temperature and humidity decreasing with height (Stull 1988).
The upper portion of the ABL is referred to as the outer, or mixed layer, and is also well mixed. The outer layer does not have the structure of the free atmosphere above and is instead characterized by convective activity and a near constant velocity profile. The depth of the outer layer varies on a diel cycle, driven by energy input at the earth's surface. Its depth will vary from a pre-dawn depth of approximately 100 m to a depth of 1000 m or more as a result of surface forcing by absorbed solar radiation (Holton 1979). The region at the top of the outer layer in which the drier, warmer air from the free atmosphere mixes with air from the outer layer is called the entrainment zone.

The surface layer occupies the lowest 10% of the ABL and is characterized by a logarithmic velocity profile and turbulent mixing. Importantly, fluxes in this layer are assumed to be nearly constant with height, allowing flux measurements made within the surface layer to be representative of the fluxes originating from the underlying surface. At the base of the surface layer is the roughness sublayer. Though similarly turbulent as the surface layer, spatial variability related to the influence of individual
surface roughness elements can result in unpredictable flux distributions. A final laminar layer, the interfacial layer, with a depth of a few millimetres to centimetres is characterized by molecular diffusive transport.

It is this knowledge of the structure of the boundary layer that helps to explain the reasoning behind current methods of measuring surface exchange. Employing the diurnal growth and well-mixed character of the ABL, attempts have been made to estimate surface exchange by treating the entire boundary layer as a chamber. These boundary layer budget (Levy et al. 1999; Pattey et al. 2002) and inversion (Denning et al. 2003; Gimson & Uliasz 2003; Gurney et al. 2002) methods are capable of estimating exchange for large regions but suffer from poor temporal and spatial resolution of exchange measurements. This method is not limited to the ABL and has been applied to constant volumes near the ground for measuring fluxes with limited spatial extent (Czepiel et al. 1996; Denmead et al. 1998).

It is the relative ease of access to, and the turbulent characteristics and constant flux nature of the surface layer, combined with good temporal and intermediate spatial resolution that has been exploited for the estimation of fluxes. Several methods have been developed for surface layer estimation of fluxes, these include profile, eddy covariance, and budget methods.

The profile methods (aerodynamic and Bowen ratio) were one of the first methods developed. Both methods assume that the turbulence of the surface layer exchanges mass in a diffusive manner, similar to the behaviour of molecular diffusion. Though these methods provide reliable results when vertical gradients of variables are large, measurements above rough surfaces or during transitional periods when gradients are small may produce erroneous results (Angus & Watts 1984; Baldocchi et al. 1988; Kanemasu et al. 1979; Sinclair et al. 1975).

Eddy covariance techniques, which rely on simplifications of the conservation equation for the variable being studied, have become a popular method for estimation of surface exchange in the past twenty years (Baldocchi 2001; Swinbank 1951). These methods are also subject to weaknesses, associated with underlying assumptions (Finnigan et al. 2003; Massman & Lee 2002). Relaxed eddy
accumulation is a variation of the eddy covariance technique in which upward and downward travelling gases are captured in separate reservoirs, instead of taking instantaneous measurements (Baker 2000). This technique was developed to allow estimation of fluxes of variables for which no instrumentation exists with sufficient measurement accuracy and resolution, at the required sampling frequency, to apply the eddy covariance technique.

For this experiment the desire to obtain spatially averaged fluxes over a forested area of an extent that could be monitored by alternate methods suggested that eddy covariance approaches would be most appropriate.

The remainder of this chapter describes the theory, processes and results of ecosystem exchange measurements applied to the Griffin forest experimental site. In the next section, eddy covariance theory is developed with descriptions of individual terms and associated simplifying assumptions. The implications of this theory for its practical application to flux measurement are discussed in following sections. The theoretical derivation is followed by a description of the data processing techniques, described by comparison of differing methods and selection of appropriate techniques or acknowledgement of potential errors. Two following chapters provide more detailed examination of techniques associated with eddy covariance. These are followed by chapter containing analysis of the exchange results in terms of temporal, spatial and environmental relations of exchange measurements. The analysis of exchange measurements examines statistical and simple environmental relations of fluxes. (A description of associated instrumental measurement techniques is presented in appendix A).

5.4 Theory

For a surface that is acting as a uniform source (or sink) of some variable, \( c \), we may estimate the amount of \( c \) leaving the surface over a given period of time (i.e. the flux of \( c \) from the surface) by measuring the flow through the faces of an imaginary box overlying the surface of interest, figure 5.2. The required characteristics of this imaginary box include diffusive transport \( (D_x, D_y, D_z) \) and the flow of \( c \) into and out
of the box \((uc_{in}, uc_{out}, vc_{in}, vc_{out}, source, \text{ and } wc_{out})\) in the three ordinal directions \((x, y, \text{ and } z)\). We must also know the change of \(c\) within the box over the period of interest (i.e. change in storage) that occurs as a result of divergences in the respective flows in and out of the box.

![Figure 5.2 Schematic of atmospheric flows associated with the conservation equation of a scalar.](image)

Although it is useful to know the flow of \(c\) both into and out of the box on all sides it is really the change in flow of \(c\) across the box in the three ordinal directions in which we are interested. For a scalar variable that does not experience a change of state the condition of this imaginary box may be expressed as (Baldocchi et al. 1988; Kaimal & Finnigan 1994; Lumley & Panofsky 1964b; Stull 1988):

\[
\frac{\partial c}{\partial t} + u \frac{\partial c}{\partial x} + v \frac{\partial c}{\partial y} + w \frac{\partial c}{\partial z} = D_x \frac{\partial^2 c}{\partial x^2} + D_y \frac{\partial^2 c}{\partial y^2} + D_z \frac{\partial^2 c}{\partial z^2} + F \tag{5.1}
\]

The first term in equation 5.1 represents the change in storage of \(c\), the second, third and fourth terms represent the change in transport of \(c\) in the \(x\), \(y\) and \(z\) unit directions. The first three terms on the right hand side of equation 5.1 represent exchange by molecular diffusion and the term \(F\) represents the value of the surface flux.
It must be noted that this equation is a simplification when applied to non-scalar variables, or those which exhibit changes of state. Similar equations in which $c$ is a velocity vector will include terms describing the influence of gravity, Coriolis, and pressure forces. Similarly the equation for temperature conservation includes terms for radiative flux loss and latent energy loss/gain while the equation for water vapour includes a term for evaporation/condensation (Stull 1988).

5.4.1 Description of terms

5.4.1.1 Variable of interest
We use the quantity $c$ to represent any atmospheric component that is used as a resource by the underlying ecosystem or necessary for its understanding. This quantity may refer to a number of variables, though we restrict its use in our study to the variables of prime interest to forest growth: water vapour, carbon dioxide, and energy in the forms of latent heat, sensible heat, and momentum.

5.4.1.2 Assumptions
Most assumptions associated with the reduction of equation 5.1 are presented below, but there are other assumptions, subtler in nature, for which the reader is referred to alternate texts for their explanations. These assumptions include shallow motion (Mahrt 1986), Bousinesque approximations, lapse rate, and assumption of shallow convection (Stull 1988).

5.4.1.3 Source term
The source term, $F$, represents the exchange of quantity $c$ with the underlying surface. It is the variable of prime interest, which we intend to obtain from our measurements. Equation 5.1 assumes that $F$ does not vary spatially; it is however, recognized that $F$ does vary spatially. This potential problem is generally addressed by selecting a site that appears homogeneous. Adherence to this assumption will be discussed in section 5.2.2.3.

5.4.1.4 Diffusion terms
The terms describing diffusion of $c$ across the walls of the imaginary box:
are generally considered insignificant, (Lumley & Panofsky 1964a). This assumption is justified on the magnitude of both the kinematic viscosity \((D_c \approx 2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1})\), and the spatial rate of change of the magnitude of \(c\), (which should be zero over a homogeneous surface). It should, however, be remembered that very close to surfaces, in the interfacial layer, these terms become important and constitute the primary transport mechanisms.

### 5.4.1.5 Storage term

The storage term, \(\partial c/\partial t\), describes the variation in the magnitude of variable \(c\) as a function of time within the imaginary volume. If we assume horizontal homogeneity, this equates to divergence of the vertical transport term.

Changes in biomass and soil temperature are sometimes included in the sensible heat storage term. Because these terms represent fluxes occurring beyond the lower boundary of the budget box, i.e. the surface defining the source, their purpose is to account for all components when attempting to close the energy budget equation. Problems caused by incorporating these additional sensible heat storage terms would arise if these terms were added to sensible heat fluxes used for analysis other than that of energy budget closure. A similar can be applied to the fluxes/budgets of carbon dioxide or water vapour because the changes in content of \(\text{CO}_2\) or \(\text{H}_2\text{O}\) in the soil and biomass over the measurement periods employed. However, variation of these components over run periods is too small to be accurately measured and can only be determined over long time periods.

The calculation of the storage term associated with the imaginary box defined in figure 5.2 is typically written as:

\[
S_c = \sum_i K_{ei} \frac{\Delta c_i}{\Delta t} \Delta z_i
\]  

Where \(S_c\) is the storage term corresponding to the flux measurement period, \(\Delta t\). The change in \(c\) over the measurement period, \(\Delta c\), should represent the difference in the ensemble average of \(c\) as measured at the beginning and end of the measurement.
period. However, it is typically assumed that a finite set of sensors, arranged along a vertical transect of the box and measured at the beginning and end of the measurement period, is representative of these ensemble averages. In this approach each of the vertical measurement locations, indicated by subscript $i$, is considered representative of the ensemble average for the corresponding layer of thickness $\Delta z$. The units of $c$ are converted into units appropriate for expression of the change in storage as a flux using the variable $K_c$. The total change in storage is then calculated for each layer and summed to obtain the change in storage for the entire imaginary volume. For exactness, and assuming a logarithmic surface layer, the location of the measurement of $\Delta c_i$ within the layer $\Delta z_i$ should be located at the logarithmic center of the layer instead of the geometric center; the associated error will be small unless the layers are very deep or the profile of $c$ is very steep.

As mentioned above, for the determination of the energy budget, a more complete accounting of storage terms must be taken. Several researchers (McCaughey 1985; McCaughey et al. 1997; McCaughey & Saxton 1988; Moore & Fisch 1986) have described storage of energy, often including multiple components related to the various sensible heat, latent heat and photosynthetic energy components. A clearer definition is available if these various components are expressed in terms of the storage of kinetic energy $S_{kin}$ and potential energy $S_{pot}$.

\[ S_{Energy} = S_{Kin} + S_{Pot} \]  

In this definition, the kinetic component describes energy stored as heat or motion. Most definitions implicitly assume that the total energy stored as motion (i.e. as changes in wind velocity or in movement of vegetation or soils) is negligible. This assumption is maintained and only the storage of heat will be considered further. The potential energy component describes energy stored in chemical processes. The most predominant of these include changes in state of water, and the processes of photosynthesis and respiration, though it is possible that other processes may contribute.

While there are several approaches that can be taken in the breaking down of kinetic and potential storage terms, they are probably best considered in terms of the
substance which is storing the energy. Therefore both $S_{\text{Kin}}$ and $S_{\text{Pot}}$ storage terms will be considered as having air, water, and multiple solids storage locations; for convenience we will group solids into soil and biomass. Kinetic energy storage may then be stated as:

$$S_{\text{Kin}} = S_{\text{Kin-air}} + S_{\text{Kin-water}} + S_{\text{Kin-soil}} + S_{\text{Kin-biomass}}$$  \hspace{1cm} (5.5)

For the potential energy components we assume that there are no significant chemical reactions occurring solely in the air and that the only storage in the water component is associated with the latent heat associated with changes of state (i.e. evaporation, condensation, or sublimation). Of the many possible energy storage components associated with chemical reactions in solids, typically photosynthesis of plants and respiration of plants and soil are grouped together and based on the net exchange of CO$_2$. The same approach has been adopted.

$$S_{\text{Pot}} = S_{\text{Pot-water}} + S_{\text{Pot-soil/biomass}}$$  \hspace{1cm} (5.6)

More exact definitions of the various storage components are described in 5.3.2.12 and their analysis is presented in section 8.2.

**5.4.1.6 Transport terms**

The transport terms of equation 5.1 describe the mean advective and turbulent transport of the $c$.

$$u \frac{\partial c}{\partial x} + v \frac{\partial c}{\partial y} + w \frac{\partial c}{\partial z}$$  \hspace{1cm} (5.7)

Under the assumptions of incompressibility and horizontal homogeneity, these terms may be reduced to the single term (Stull 1988):

$$\frac{\partial wc}{\partial z}$$  \hspace{1cm} (5.8)

The assumption of horizontal homogeneity may be a weak assumption (Aubinet et al. 2003; Rao et al. 1974). Spatial inhomogeneity probably exists on most sites, and can be exaggerated by making measurements within the roughness sublayer, by complex topography, or by careless site selection. Similarly, inhomogeneity associated with temporal variations such as katabatic flows, and boundary layer or mesoscale convective flows may create or exaggerate spatial inhomogeneity. Research aims to investigate the occurrence of this potentially large term. While attaining energy
budget closure at a site suggests that transport terms are not affecting flux measurements, it does not negate the possibility of offsetting transport terms in the component energy budget fluxes, nor does it apply directly to fluxes of other scalars (eg. CO₂).

By applying Reynolds decomposition equation 5.8 may be decomposed into its mean and fluctuating components.

\[
\frac{\partial \overline{w'c'}}{\partial z} + \frac{\partial \overline{w'c'}}{\partial z}
\] (5.9)

By integration with respect to the vertical component, as was done for the storage term, we obtain transport in terms of the products of the mean and fluctuating components of vertical velocity and \( c \).

\[
\overline{w'c'} + w'c'
\] (5.10)

A further assumption that mean vertical velocity is zero allows removal of the first term of equ 5.10. This assumption results in some practical problems associated with the determination of vertical velocity. These problems are discussed in sections 5.3.2.5 and 5.3.2.9. Given that these problems are resolvable, the resulting equation for the determination of ecosystem exchange of \( c \) may now be expressed as:

\[
F = w'c' + S_c
\] (5.11)

### 5.4.2 Implications for measurements

Practical problems associated with the determination of the terms in equation 5.11 have implications for the application of the eddy covariance technique. The problems associated with each term are discussed below and are followed by a discussion of site-specific problems.

#### 5.4.2.1 Implications for flux measurement term

The term of equation 5.11 describing the fluctuations of vertical velocity and \( c \) requires knowledge of the 'infinite' mean and measurements of all instantaneous deviations from that mean for both variables. Though not stated explicitly in the theoretical development, it is the case that both variables must be measured at the same point in space and time. The necessity of making measurements with these
characteristics can present several practical problems that manifest themselves as an inability to determine appropriate signal means and an inability to resolve fluctuations at either very long or very short time scales. The ability of a sensor to respond to signal fluctuations is referred to as the sensor’s frequency response characteristics, and the correction for these limitations as the frequency response correction.

Practical problems arise in determining signal mean values because of instability of instruments over very long periods and because of averaging, velocity rotation and filtering methods. The determination of signal means is restricted not only by instrument stability but also by the length of measurement periods. This topic has been discussed by Finnigan et al. (2003) and will be covered in section 5.3.2.9. At the other end of the spectrum, finite temporal resolution of an instrument causes signal attenuation at high frequencies. Besides an instrument’s physical characteristics, frequency response limitations may result from sensor deployment geometry (Kristensen et al. 1997; Laubach & McNaughton 1998; Lee & Black 1994) and signal handling and processing.

The problem of signal frequency response was recognized quite early in the development of eddy covariance measurement systems (McIlroy 1961) but was only popularised in 1986 by Moore who created simplified algorithms for the corrections caused by inadequate sensor frequency response (Moore 1986). This will be discussed further in section 5.3.2.7 and chapter 7.

5.4.2.2 Implications for storage measurements

There exist two problems associated with determination of storage estimates. The first of these problems is caused by assuming that a point measurement is representative of a layer ensemble mean. To obtain a statistically accurate sample may require an unfeasibly large number of spatial samples. For example, using equation 2.1, it was determined that 74 samples would be required to accurately represent \( i.e. \) within 10\% a run-to-run change in temperature of 0.2 C if the standard deviation was on the order of 0.4 C. Because a similar number of samples may be required for most variables and because we represent the variable means with a single sample we are...
not able to address the potential for spatial variability of storage terms, (see section 8.2).

A second, related problem is caused by the difference of the location of storage measurement from that of the flux source. While both the flux and storage measurements have associated footprints, the locations of the two footprints can be quite different (Rannik et al. 2003; Schmid & Lloyd 1999). Because a storage measurement is made closer to the surface, its footprint is smaller and closer to its measurement point than is that of a flux measurement footprint. Though it is necessary to assume that both footprint locations are identical, this assumption will fail if storage varies spatially.

5.4.2.3 Implications for experiment design

The flux equation (equ. 5.11) likewise has implications for experimental design and implementation. Because horizontal homogeneity is a key assumption in determining equation 5.11, it is prudent to be aware of any existing site inhomogeneity. We have already gained some insight into the temporal variability from review of the observed spectral powers of climate variables in chapter 4; however, it is site spatial variability with which we are concerned. Because variability exists both in land cover and topography, more direct measures of spatial inhomogeneity for the Griffin site were obtained from site topographic and photographic information.

While topographic variability can be inferred visually from figure 2.7, a more quantitative measure can be obtained from a variogram of the same topographic information. A variogram describes the variability as a function of spatial separation. In general variability will increase with separation up to a distance (i.e. the range) at which the variability reaches a value (i.e. the sill) beyond which variability remains nearly constant. Variograms of local (1 km square) and regional (10 km square) topographic information, figure 5.3, reveal a near linear increase in variability out to a range beyond 5 kilometres, suggesting that the topography has scales of variability much larger than the 5 km separation used in this analysis. More notably, the local topography variogram suggests that, to a lag distance of nearly 300 meters, the rate of increase in topographic variability remains half that at greater lag distances, indicating
that, on average for any selected location, reasonable topographic homogeneity exists out to this distance. This analysis indicates that the current site selection, for which elevation changes are small to a distance of 500 m, is representative of the region.

![Figure 5.3](image1.png)  
**Figure 5.3** The right panel is a variogram of topographic heights fitted with a linear model. Topographic data used are from a 1 by 1 km square centred on the Griffin site (left panel) and a 10 by 10 km square region centred on the Griffin experiment site (right panel). Both variograms were calculated using Surfer v8 (Golden software).

![Figure 5.4](image2.png)  
**Figure 5.4** Variogram of aerial photo brightness values fitted with a rational quadratic model. Brightness values were obtained from a digitised aerial photo, which had been resampled from 300 dpi to 30 dpi. Raster brightness values were vectorized (IDRISI, Clark Labs). Data are for a 1 by 1 km square region centred on the Griffin experiment site. The lag
distances on the horizontal axis must be multiplied by a value of 5 to obtain a true lag distance.

Further information on site variability in the form of roads, forest rides, streams, and variable forest development was derived from aerial site photographs, figure 5.4. A variogram of the aerial photo brightness values was fitted with a Rational quadratic model, (Surfer, Golden software) in order to quantify the associated variability. In this variogram model, the range to the sill is approximately 60 meters. This range suggests that the scale of surface cover variability may be less than the typical scale of a flux footprint, which is generally on the order of 50 -500 m (Schmid 2002; Schmid et al. 1999).

Footprint modelling (Kormann & Meixner 2000), was used to determine typical footprint dimensions for Griffin under the observed range of daytime and nighttime conditions. The resulting footprint dimensions (peak distance ~ 150 m, width ~ 100 m) suggest that the footprints adequately represent the observed variability of the canopy. These dimensions, however, belie averaged footprints that were observed to have shorter peak distances and smaller averaged widths (both ~ 20 m). These smaller values represent the effect of unstable, low wind speed conditions and under such circumstances the small-scale surface variability at Griffin may not be adequately represented, with biases towards surface characteristics immediately surrounding the flux tower. Under all but the most stable conditions, the observed footprint characteristics fell within the 200-300 m topographic scale for which variability showed more spatial consistency.

Averaged footprint analyses, using the footprint model of Fan et. al. (1992), indicates that the average flux footprints for the five years of data collection were centred on two main regions to the northwest and southeast of the tower, with a minor region to the south southwest. The two main regions reflect the bimodal wind direction forcing of the valley with the minor region reflecting the effect of topography on wind flow from the south under higher wind speed conditions, see section 4.13. The footprints show that the predominant regions of flux sources do contain some contamination by breaks in the forest cover associated with roads, rides and streams.
Referring to figure 4.62, we recall that under light winds upslope flow predominated during the day while down slope flow predominated at night. Therefore any variability in fluxes or storage within the two footprints and correlated with upslope and down slope flow directions may have resulted in a bias in the resulting flux values. Verification of such a bias would require more detailed spatial measurements of fluxes and storage than were available in this experiment.

The only further implications of the theoretical derivation of equation 5.11 lies in the processing of raw signal samples to achieve the closest possible representation of ecosystem exchange. This procedure requires close scrutiny of the applied signal calculation and correction methods, and a justified selection of the most appropriate signal processing methods.

![Flux footprints](image)

**Figure 5.5** Flux footprint (Bakwin et al. 1992) probability distributions using available data for the years 1997 through 2001. Contours represent average probability levels of 0.2 percent as accumulated in 50x50 m grid squares.
5.5 Data acquisition and processing

5.5.1 Data management
Much of the data management employed in the processing and analysing of raw eddy covariance and logger data for the calculation of ecosystem exchanges is described in the sections detailing microclimate measurements in appendix A. Additional management of raw data involved transferral of all raw data to a single, external hard disk drive to allow unimpeded access to all raw data files. Files were excluded if file length fell below an arbitrary threshold limit of approximately 20 minutes in length, or if manual quality control contraindicated its inclusion.

5.5.2 Data processing
For the processing of raw data a software package, EdiRe, was developed to allow simplified alteration of analysis procedures while still maintaining the rapid processing rates required when processing large data sets. Considerable effort went into the development of this package, and it has been made available free of charge to other users in need of such software. No effort will be made to describe the structure and operation of EdiRe as the help files describing the program stretch to over 400 pages, a volume of text thought inappropriate for this thesis. The associated help files provide an adequate description of the operation and capabilities of this package and a current version of the program can be obtained by contacting the author.

5.5.2.1 Data extraction and conversion
The initial steps in data processing involved extraction of signals from the raw data files and conversion to appropriate units. The conversion of signals involved application of appropriate signal calibration equations.

For the sonic anemometer signals \((u, v, w, \text{ and sonic } T)\) calibration involved application of constant coefficients. The anemometers internal processor, prior to recording of the \(u, v, \) and \(w\) signals, applied anemometer-specific calibrations for wind flow distortion by the anemometer body (Gill Instruments Ltd. 1990). These calibrations accounted for azimuthal effects of wind flow and were valid for wind attack angles of up to \(\pm 30\) deg. Recent investigations (Gash & Dolman 2003; van der Molen et al. 2004) have shown that further correction of anemometer data is required.
for larger attack angles. While the amount of signal occurring beyond these angles is generally small, (figure A.22), it can be determined from uncorrected joint flux distributions, that on the order of 10% of an observed flux may coincide with attack angles outside the valid sensor calibration range. Attack angle dependent correction of velocities following Van der Molen et al. (2004) was applied to the sonic u, v, and w data. During application, this correction was limited to attack angles beyond ±40 degrees because of calculation instabilities in their correction at small attack angles. The effect of this correction upon fluxes was generally less than about 3 percent. Subsequent reanalysis suggest that the applied corrections will have been too small by approximately 2%. A discussion of this reanalysis is covered in Appendix H. It is suggested that further research into this problem should be conducted to establish well-behaved correction factors and to establish if generalized corrections are possible or if sensor specific corrections are required.

The only other calibrations applied to the signals were those associated with the IRGA calibrations. Calibration results given in appendix A.3 and A.4 indicate the range of calibration offsets and gains which were applied to the IRGA CO₂ and H₂O signal voltages. It is apparent from Griffin calibration data (Moncrieff et al. 2004) that random errors in the calibration data likely exist. It was assumed, however, that these calibration coefficients were correct and no attempt was made to adjust for these errors. A statistical comparison of two months of data from the summer of 1998 suggested that incorporation of sensor calibration coefficients did not have a statistically significant effect (p < 0.05, n = 2846) on the flux values of either latent heat or carbon dioxide flux, with the observed effects being much less than 1%. Regardless of the small, observed effect over the test period, IRGA calibrations were applied to the raw data during reprocessing.

5.5.2.2 Run length

The effect of run length on ecosystem exchange determination has, until quite recently, been considered of relatively minor effect, as long as runs were sufficiently long. The commonly accepted run length was normally considered to be from 15 to 60 minutes in length, (Kaimal 1975; Lenschow et al. 1994; Wyngaard 1973). More recently, researchers have begun reconsidering the governing equations and the
possible contribution that low frequency signal variations may make to flux estimates (Finnigan et al. 2003; Lee 2000; Sakai et al. 2001). These considerations have addressed the effect of mesoscale and boundary layer convective activity, as well as data treatment (e.g. high pass filtering, rotation) and how these factors may affect the low frequency flux contributions. Analysis of the effect of run length by simply increasing the length of the run has suggested that additional flux energy on the order of +/- 10 percent can be obtained for some experiments (Malhi et al. 2002; Sakai et al. 2001). For this experiment, thirty minute run lengths were employed. The contribution of low frequency fluxes will be discussed further in section 5.3.2.9.

5.5.2.3 Quality assessment

Quality assessment of flux data was performed on both the raw data and on flux values obtained through data processing. The quality assessment procedures applied to flux values was identical to that performed on run mean values of microclimate variables and is described in section 3.5. Because of the very large amount of raw data, its quality assessment was typically limited to situations under which questionable flux values were obtained. Quality assessment procedures included cospectral examination for noise, despiking of the raw data, and analysis of statistical indicators.

5.5.2.4 Noise and despiking

Noise and spikes, though similar in nature, present decidedly different problems of detection and removal. The difference between noise and spikes in a signal time series is subtle. Noise cannot be differentiated from the local signal in magnitude or in time, while spikes are identifiable because of their localization and/or magnitude. These two characterizations represent opposing ends of the same problem of an unwanted signal imposed upon the desired signal.

Though noise is indistinguishable in a time series view of a signal, it may appear quite prominently in the frequency domain, thus providing an indicator of its presence. While noise can theoretically be removed after it has been recorded, initial prevention of noise is the better option because noise, at frequencies higher than the Nyquist folding frequency, which remains unfiltered before digitisation results in

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contamination of frequencies lower than that of the existing noise (Ifeachor & Jervis 1993).

Unfortunately there was no capacity in the sonic anemometer for filtering of velocity or temperature signals other than the digital signal processing conducted by the anemometer. Similarly the CO₂ and H₂O signals were not filtered, allowing some noise to exist in those signals. Cospectra were routinely inspected to check for the presence of noise. Fortunately any existing noise in velocity and scalar signals was not correlated such that it did not appear in resulting covariances. High frequency noise in scalar signals was also quite prevalent, because of the low signal to noise ratios under some conditions. Similar to velocity noise, scalar signal noise did not appear in the covariances used to calculate fluxes.

In contrast to noise, spikes are localized in time and have a magnitude greater then that of the surrounding signal. This characteristic allows the spikes to be identified and removed. A despiking routine was developed to allow spike identification and removal. This routine was based on the work of Hojstrop (1993) modified to adapt it to the particular task of raw data testing.

Spikes were detected using a the absolute values of the differences in subsequent points in the time series. The standard deviation of this absolute value data set was calculated and values greater than n standard deviations chosen as identifying a ‘leg’ of a spike. Such points could represent either the upward or downward leg of a spike, as shown in figure 5.6. The largest value of the absolute value data set was selected, and if its value was larger than the specified number of standard deviations a search routine was initiated that identified the corresponding leg of the spike. This search was applied over a specified range of times (i.e. spike width) on both sides of the identified spike leg. The search also allowed for a percentage changes in the value of the corresponding leg using a decay parameter. The despiking routine removed wider spikes before identifying and removing narrower spikes. To remove a spike the despiking routine fit a linear trend to the data between the two legs of the spike and subtracted the resulting fit from the original signal, as shown in figure 5.6.
Figure 5.6 Example of signal despiking process. Solid line represents the signal and differenced signal. Grey line represents specified acceptable standard deviation limit of differenced signal. The dotted line represents the despiked signal.

Figure 5.7 Signal spike counts for test done on two month period from the summer of 1998. The same despiking methods were applied to all six signals prior to any other signal calculations. Methods employed either 5 or 7 standard deviations (s5, s7) as lower limit of spike magnitude, spike widths of either 3, 6 or 12 consecutive samples (w3, w6, w12) and spike leg consistencies of either 30% or 60% (c30, c60).
A test of the despiking method was carried out using two months data from the summer of 1998. In this test the differenced signal standard deviation limit was specified as either 5 or 7, spike widths as 3, 6 or 12 samples, and spike consistency as either 30 or 60%. The effect of specifying smaller standard deviations, greater spike width, and decreased consistency was to increase the number of spikes identified, (figure 5.7). The average number of spikes identified per run varied from near zero to greater than 10. The effect of despiking on the fluxes, shown in figure 5.8, was negligible. The maximum effect observed occurred in sensible heat flux, with a reduction of approximately 0.8% for the most restrictive despiking parameters. Regression analysis of the despiked against non-despiked fluxes produced $R^2$ coefficients of 1.0 for all comparisons, suggesting that despiking did not alter variability of the fluxes. Because of the minimal effect of despiking upon the signals in this dataset, and the high computing performance costs required to apply the procedure, it was decided not to employ despiking methods in the final computation of fluxes.

![Despiked flux magnitudes](image-url)

**Figure 5.8** Effect of despiking on flux magnitudes for sensible heat flux $H$, momentum flux, $UW$, latent heat flux, $\lambda E$ and carbon dioxide flux, $F_c$. A description of the methods is given in the caption for figure 5.7.
5.5.2.5 Rotation

The purpose of anemometer signal rotation is to obtain a set of velocities in which the $u$ component represents the mean stream wise wind, $v$ the lateral wind component and $w$ the vertical wind component. It is also required by definition in equation 5.11 that there be no mean vertical velocity. Ideally, a single axis anemometer could be installed to measure only the vertical velocity component, as that is the only velocity signal required for the measurement of scalar fluxes. However measurements over other than flat terrain makes installation of such sensors very difficult. The alternative is to measure three, orthogonal wind velocity components and to rotate these signals after measurement by assuming a zero mean vertical velocity. Two methods of signal rotation were evaluated, planar fit rotation (McMillen 1998), and run period rotation (Wilczak et al. 2001), which, until recently, has been the recommended method of rotation. The planar fit method is currently recommended because it is less susceptible to run-to-run variations resulting from low frequency variations in velocity means (Finnigan et al. 2003). Following is a description of the two methods and their application to the Griffin data set.

Run period rotation

The run period rotation may be employed as either a two or three angle rotation adjustment. The first rotation angle, $\alpha_r$, rotates the coordinate frame about the $z$-axis to place the $u$ velocity in the direction of the stream wise wind velocity. The second rotation angle, $\beta_r$, rotates the $v$ and $w$ velocities about the new $x$-axis so that mean vertical velocity is zero. These two steps can be calculated using equation set 5.12 to determine the rotation angels from the mean unrotated wind velocities and equation set 5.13 to apply the rotation.

$$\alpha_r = \tan^{-1}\left(\frac{v}{u}ight)$$

$$\beta_r = \tan^{-1}\left(\frac{w}{\sqrt{u^2 + v^2}}\right)$$

(5.12)
\[
\begin{align*}
    u &= u_u \cdot \cos(\alpha_r) \cos(\beta_r) + v_u \cdot \sin(\alpha_r) \cos(\beta_r) + w_u \cdot \sin(\beta_r) \\
    v &= -u_u \cdot \sin(\alpha_r) + v_u \cdot \cos(\alpha_r) \\
    w &= -u_u \cdot \cos(\alpha_r) \sin(\beta_r) - v_u \cdot \sin(\alpha_r) \sin(\beta_r) + w_u \cdot \cos(\beta_r)
\end{align*}
\]  

Figure 5.9 Second rotation angle averaged by sensor installation period and wind direction.

The third rotation angle acts to minimize the magnitude of the correlation between lateral and vertical wind velocity fluctuations. It is suggested that over flat homogeneous terrain that this correlation should be zero. Over complex terrain and rough surfaces, however, this correlation is either unstable or does not tend towards a value of zero. For this reason the third rotation angle is normally not applied over rough surfaces or in complex terrain, and was not applied to the Griffin data set when evaluating this rotation method.

Analysis of the second rotation angle for the periods corresponding to different sensor installations, (figure 5.9), suggests different patterns of $\beta_r$ with respect to wind direction but no clear differentiation between installation periods.
Planar fit

The planar fit rotation (Wilczak et al. 2001) requires a period of unrotated run mean $u$, $v$, and $w$ velocities from which a rotation can be determined that sets the period mean of the $w$ velocity to zero. Following Wilczak et al. (2001), the required rotation coefficients, $w_0$, $\alpha_{pf}$, $\beta_{pf}$ may be obtained from the parameters $b_0$, $b_1$ and $b_2$ which are chosen to minimize the summation (equation 5.14) over all runs in the period.

$$\sum_i \left( w_{ui} - b_0 - b_1 u_{ui} - b_2 v_{ui} \right)^2$$

(5.14)

The rotation coefficients were then applied to individual runs from the period used to determine the coefficients, using the equation set 5.15. This rotation places the velocities into the frame for which vertical velocity (adjusted for suspected offset error) is zero when averaged over the period of coefficient determination.

$$u_p = u_u \cdot \cos(\alpha_{pf}) - (w_u - w_0) \cdot \sin(\alpha_{pf})$$

$$v_p = u_u \cdot \sin(\alpha_{pf}) \cdot \sin(\beta_{pf}) + v_u \cdot \cos(\beta_{pf}) - (w_u - w_0) \cdot \cos(\alpha_{pf}) \cdot \sin(\beta_{pf})$$

$$w_p = u_u \cdot \sin(\alpha_{pf}) \cdot \cos(\beta_{pf}) - v_u \cdot \sin(\beta_{pf}) - (w_u - w_0) \cdot \cos(\alpha_{pf}) \cdot \cos(\beta_{pf})$$

(5.15)

A second step is required to rotate the horizontal components through an angle ($\gamma_{pf} = \arctan(u_{u0}/v_{u0})$) such that $u_{u0}$ represents the run mean wind component. The coefficient for this rotation is calculated for each run and not for the period of determination of the coefficients, $w_0$, $\alpha_{pf}$ and $\beta_{pf}$.

$$u = u_p \cdot \cos(\gamma_{pf}) + v_p \cdot \sin(\gamma_{pf})$$

$$v = -u_p \cdot \sin(\gamma_{pf}) + v_p \cdot \cos(\gamma_{pf})$$

$$w = w_p$$

(5.16)

To evaluate this method, the planar fit rotation coefficients were calculated for seven-day periods of run averaged $u$, $v$, and $w$ velocities. The resulting coefficients (figure 5.10) showed distinct patterns associated with the periods of installation for different anemometers. The corresponding vertical velocity offsets did not show the
same distinct pattern. Additional analysis of vertical velocity offset values did not reveal any relationship to mean wind speed or Monin-Obukhov stability.

Figure 5.10 Planar fit vertical velocity offset and rotation coefficients obtained from one week periods weekly and corresponding data averaged for sensor installation periods.

A comparison of the effects of planar fit rotation approach with that of the standard 2 axis rotation, (figure 5.11 and table 5.1), suggest that the net effect on fluxes was small. Although the net effect on the fluxes is two percent or less for the data examined it is also apparent from figure 5.11 that the effect on individual runs can vary quite significantly depending upon the method employed.

For final calculation of the data set, the planar fit method was employed. This method was selected as it appeared to best represent the actual sensor installation periods, and because its coefficients are less susceptible to alteration by atmospheric effects with time scales similar to that of the desired run length.
Figure 5.11 Comparison of fluxes calculated using the standard 2-axis rotation with those calculated using the planar fit rotation employing installation period averaged rotation coefficients. Data used in the figures are from two months in the summer of 1998.

Table 5.1 Regression coefficients for the relationship of fluxes calculated using 2-axis rotations (independent variable) and planar fit rotation (dependent variable), n=2847.

<table>
<thead>
<tr>
<th>Flux</th>
<th>offset</th>
<th>slope</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$H$</td>
<td>0.41</td>
<td>0.98</td>
<td>0.99</td>
</tr>
<tr>
<td>$u'w'$</td>
<td>0.0027</td>
<td>0.99</td>
<td>0.99</td>
</tr>
<tr>
<td>$\lambda E$</td>
<td>1.23</td>
<td>1.01</td>
<td>0.90</td>
</tr>
<tr>
<td>$F_c$</td>
<td>0.038</td>
<td>1.02</td>
<td>0.97</td>
</tr>
</tbody>
</table>

5.5.2.6 Lag removal

When calculating fluxes, lag removal is employed with scalar signals that are sampled at a physically separate location from that of the vertical velocity. Employing Taylor’s
hypothesis of frozen turbulence \( x = u \cdot t \), a physical separation, \( x \), of measurements can be translated into a temporal separation, \( t \), when the transition velocity, \( u \), is known. Cross correlation of signals can then be employed to determine the time lag associated with the physical separation (Kristensen et al. 1997; Lee et al. 1994). For open path sensors this separation is generally small and dependent upon wind direction. For such sensors the error associated with physical separation may be corrected by employing frequency response correction techniques (Irwin 1979; Kristensen & Jensen 1979; Moore 1986). For closed path sensors the separation between the vertical velocity and scalar sensor can be considerably larger. In this situation the lag time will depend not only upon the physical separation of the sensor’s intake and the vertical velocity sensor but also upon the sample flow rate and gas sample line dimensions. For closed path sensors the lag time is generally removed as a distinct process and corrections for signal loss caused by attenuation of fluctuations for gas travelling down the sample tube are treated using frequency response techniques (Lenschow & Raupach 1990; Massman 1991; Philip 1963; Rannik et al. 1997).

![Sample cross correlation curves](image)

**Figure 5.12** Sample cross correlation curves used for the determination of signal lag time. The top panel shows a curve with from which a lag time can be easily determined. The bottom panel shows an example of a poorly formed cross correlation curve. The dashed line represents the linear curve fitted to the correlation curve endpoints and the arrow represents the point of maximum magnitude difference.
The magnitude of the circular cross correlation of vertical velocity with a scalar will obtain a maximum (or minimum depending upon the sign of the flux) corresponding to the lag time. This lag time was identified by determining a linear fit between the end points of the circular cross correlation curve over the range of expected lag. The difference between the magnitude of this fit line and the correlation curve is then searched for the maximum deviation, which under most circumstances corresponds to the desired lag time as shown in the top panel of figure 5.12.

Incorrect determination of a signal's lag time will bias fluxes because it will always act to reduce the magnitude of the associated flux. For fluxes of reasonable magnitude there will be a small error in the lag determined from the time of peak correlation because of the temporal resolution of the sampling frequency. For example, sensors sampled at 20 Hz will have lag time resolution of ±0.05 s. Based on lag curves for sonic temperature and vertical velocity a resolution of 0.05 seconds will induce a reduction in the flux magnitude of 1% or less. Examination of the frequency distributions of sonic temperature lags indicates that only 1.3% of runs have lags greater than 1 sample. This suggests that for well formed cross correlation curves, accurate lag times can be determined under most circumstances. However, when flux magnitudes are low, cross correlation curves may be ill formed. Such curves, (eg. bottom panel in figure 5.12), may produce inaccurate lag time determinations. Fortunately, because of the corresponding small flux magnitudes such inaccuracy will cause only small absolute errors.

Statistical tests of the effect of lag removal on the fluxes of carbon dioxide and latent heat flux were also performed. The test compared fluxes for which lag times were determined for each run with fluxes using constant lag times. For CO₂, the constant lag time was set to the annual mean lag time, while for water vapour the lag time was set either to the annual mean or to the annual mean lag used for CO₂. The results of the comparison indicate that employing a constant lag time for CO₂ did not have a significant effect on the fluxes (p=0.05, n= 2486), with less than a 1% effect on the regression between fluxes. The use of an annual mean lag for H₂O did not significantly reduce the fluxes of λE (slope 0.94 R² 0.98), but the use of the annual mean CO₂ lag with H₂O did cause a significant reduction (slope 0.87, R² 0.97) in λE.
A test of the effect of determining lag either before or after coordinate rotation again revealed no significant effect (p=0.05, n= 2486), though it was noted that a very few runs exhibited large differences between the two methods.

For final calculation of the fluxes, lags were determined after velocity signal rotation and were removed for signals obtained from closed path sensors. For CO₂, these lags were limited to the range of 6 to 8 seconds to allow for small variation in pump operation. For H₂O, lags were limited to the range of 6 to 25 seconds. The reason for the large range of lag times is discussed below.

5.5.2.7 Intake tube adsorption/desorption
The lag of a trace gas signal because of its transport through an intake tube is caused primarily by the (slug) transport of the gas, though it is known that an additional lag is induced by attenuation of the signal through mixing of gas during transport (Massman 1991; Philip 1963). This additional lag corresponds roughly to the time constant of the associated high frequency attenuation. For carbon dioxide this attenuation has been observed to be nearly constant. Analysis of water vapour lag times for the Griffin data set suggests an additional attenuation that is not constant but which is a function of environmental conditions. Similar observations have been made by other researchers (Lenschow et al. 1990; Massman 1991). To understand the cause of this variable lag time requires a reference to the chemistry of the interaction of gaseous and solid materials.

For flow through a tube, the species gaseous concentration will attempt to obtain equilibrium with its wall bound concentration, \( C_w \) (mol cm\(^{-2}\)). This equilibrium is determined by the difference in rates of adsorption onto the wall and desorption from the wall’s surface. The rate of molecular desorption can be defined using the relationship (Do 1998; Gregg & Sing 1967):

\[
\frac{dC_w}{dt} = \frac{2}{r} \cdot C_w \cdot A_p \cdot e^{-E_d / r} \tag{5.17}
\]

in which the ratio 2/r represents the surface area to volume ratio of a tube with radius \( r \), \( A_p \) is the Arrhenius prefactor describing the molecular exchange rate (\( \sim 10^{12} \text{ s}^{-1} \)), \( E_d \)
is the activation energy of desorption for the species (~20 kJ mol\(^{-1}\) for H\(_2\)O), \(R_g\) is the ideal gas constant (8.314 J mol\(^{-1}\) K\(^{-1}\)), and \(T\) (K) is the temperature of the system.

The rate of adsorption of a gas onto the wall may be described using a Langmuir adsorption isotherm (LeVan et al. 1997) if the adsorbing gas forms only a single molecular layer on the wall of the tube. However, for water vapour it is known that multiple layers of molecules may exist on the adsorbing surface. In such circumstances an alternate isotherm is required to describe the characteristics of the adsorption. While many different functions have been put forward as explanations of the adsorption behaviour of multiplayer adsorption, the BET isotherm equation provides a useful general function to describe multiplayer adsorption (LeVan et al. 1997):

\[
N_w = \frac{c_e \cdot e \cdot N^*}{(e_s - e) \left[1 + \left(\frac{c_e}{e_s} - 1\right)\frac{e}{e_s}\right]} \quad (5.18)
\]

in which \(N_w\) is the amount of gas adhering to the wall, \(N^*_w\) is the amount of gas constituting a mono-layer coverage of the wall, \(e\) and \(e_s\) are the partial pressure and saturation pressure of the gas species and \(c_e\) is a species specific constant. We first rearrange equation 5.18 to obtain the ratio of partial to saturated pressures (i.e. \(h_i\)). The resulting equation was then differentiated with respect to RH and multiplying both sides with the time derivative of \(h_i\) we obtain the rate of adsorption of the gas species to the wall.

\[
\frac{dN_w}{dt} = \frac{dh_i}{dt} \cdot \frac{c_e \cdot N^*}{\left(h_i - 1\right)^2 \left(1 + h_i^2 \left(\frac{c_e}{e_s} - 1\right)\right)} \quad (5.19)
\]

We see from equations 5.18 and 5.19 that the gas exchanged with the intake tube surface will be a balance between a temperature controlled process (desorption) and a saturation fraction controlled process (adsorption). Because of the denominator of the second term on the RHS of equation 5.19, the amount of transfer of water vapour to the tube wall will tend to infinity as the gas species approaches saturation (i.e. condensation occurs). In contrast, experience tells us that, for the range of observed temperatures, that desorption caused by temperature increases does not have as large
an effect. This implies for gases that are near saturation, i.e. water vapour, that adsorption will occur more rapidly than desorption. Based on this knowledge, a relation between lag time and relative humidity was identified, (see figure 5.13). Assuming that this additional attenuation process was behaving as a first order system, an exponential relation was fitted, providing the relationship,

\[ t_{\text{lag}} = 7.15 + 0.0085 \cdot e^{13.8} \]

Although it can be determined from equation 5.20 that the relationship should behave more closely as \((1-h_r)^{-1}\) and should also have a temperature dependency, the above relationship was employed because it provided a simple means to relate to a first order filter function. Further analysis of the signal lag times did suggest a temperature dependency, however, autocorrelation of temperature and relative humidity made determination of a representative function impractical. Further investigation into this topic is needed to define an appropriate physical model fully describing this relationship.

As observed in figure 5.13 there is considerably more scatter associated with water vapour lag time determination under conditions of high humidity. It was assumed that much of this scatter results from errors in the determinations of lag peaks, though some scatter is also likely the result of errors in the determination of relative humidity values. Therefore, lag was determined for each run and limited to range of lags suggested by the model \(i.e. 6\) to \(25\) seconds).

The lag time of carbon dioxide increased only slightly with relative humidity. This slight variation may be the result of increased scatter or may be the result of a greater affinity of \(CO_2\) molecules to water covered tube walls under high humidity. The observed lag time variation was not significant and run determined lag times were limited to the range of \(6\) to \(7\) seconds to allow for small variation in flow rate.
5.5.2.8 Filtering/detrending

In the calculation of eddy covariance fluxes, highpass signal filtering is used to remove signal drift and effects of non-stationarity of environmental conditions, while lowpass filters are used to remove high frequency noise. Highpass filters, historically, have been used to remove the effects of instrument drift (Shuttleworth 1988). Over the past few decades, instrumentation electronics and construction have improved to the point where signal drift may often be considered insignificant in relation to the observed variability of signals. However, the use of filtering and detrending to remove the effects of non-stationarity is still common practise, and is often a recommended signal processing technique (Aubinet et al. 2000). The theoretical basis of this practice has come under question based on the reasoning that there should be no innate correlation between the vertical velocity and other signals used to construct flux values as a result of diel or annual patterns, and that other patterns may in fact be contributing to the true flux of the variable concerned (Finnigan et al. 2003). For this reason no high pass filtering was applied to this data set, other than that inherent in the block averaging associated with the run length.
5.5.2.9 Low frequency flux information

The use of half hour run lengths has the effect of eliminating low frequency turbulent flux information. Ideally, this information can be recovered by including mean transport term described by equation 5.10.

\[ \overline{wC} = \overline{wC'} + \overline{w'c'} \]

The mean transport term, \( \overline{wC} \), cannot be determined if anemometer signals are rotated over the run length. Similarly, if anemometer signals are rotated using planar fit methods then the resulting mean transport term may be subject to errors caused by sensor inaccuracy and long-term instability.

An appropriate method would include information from periods long enough to include most low frequency flux but short enough to minimize the effects of instrument instability. The flux of this long time period may also be subdivided into two shorter time periods \((t = t_A + t_B)\). The average turbulent flux of these two shorter time periods will contribute to the turbulent flux of the longer time period in proportion to their time fraction relative to that of the long time period:

\[ \overline{wC} = \overline{wC'} + \overline{w'c'} = \overline{wC'} + \left( \overline{w'c'} \right)_A \frac{t_A}{t_A + t_B} + \left( \overline{w'c'} \right)_B \frac{t_B}{t_A + t_B} \]

(5.22)

We may also express this proportion in terms of the ratio of period \( A \) to the total period \((f = t_A/(t_A + t_B))\), giving a variable that ranges between a value of 0 and 1:

\[ \overline{wC} = \overline{wC'} = \overline{wC'} + \left( \overline{w'c'} \right)_A f + \left( \overline{w'c'} \right)_B (1 - f) \]

(5.23)

If we assume that the total period is sufficiently long to provide a stream wise flow representative of the upper surface of the imaginary box in figure 5.1 we may also rotate the velocity signals so that \( \overline{w} = 0 \) using methods described in sections 5.3.2.5.

\[ \overline{wC} = \overline{w'c'} = \left( \overline{w'c'} \right)_A f + \left( \overline{w'c'} \right)_B (1 - f) \]

(5.24)
Although rotation will set the mean vertical velocity of the long period to zero, it must be remembered that the residual mean vertical velocity and mean scalar value of the sub-periods A and B are not guaranteed to be zero. These residual mean transport terms, \( \bar{w}\bar{c} \), represent a low frequency flux component not captured by calculation of only the turbulent flux component over short time periods, \( i.e. \bar{w}'c'' \).

\[
\bar{w}'c' = (\bar{w}\bar{c} + \bar{w}'c'')_A f + (\bar{w}\bar{c} + \bar{w}'c'')_B (1 - f)
\]  

(5.25)

The flux from one period may then be determined as the difference in the flux from the entire period and the flux from the other period, or directly as the sum of the covariance, advective term and error of the short-term period, \( i.e. \bar{w}'c'' \).

\[
\left( w'c' \right)_B = (\bar{w}\bar{c} + \bar{w}'c'')_B = \frac{\bar{w}'c'' - (\bar{w}'c'')_A f}{(1 - f)}
\]  

(5.26)

Using the run rotation method the value of \( \bar{w} \) for the combined sub-periods will be zero while for planar fit method it will be non-zero. However, for both methods the value of \( \bar{w} \) for both sub-periods for both rotation methods should be similar.

Similarly, the flux value could be obtained directly by calculating \( (\bar{w}\bar{c} + \bar{w}'c'') \) for the desired sub-period.

Calculations using this method produced fluxes with much greater variability. It was observed that this variability behaved randomly over many realizations such that the regression of a suitably large number of short-term fluxes including low frequency information against short-term fluxes without low frequency information obtains the increase in flux by low frequency contribution from the slope of the regression. Caution must be used when assuming random contribution by low frequency flux since it is possible that this contribution may be associated with correlated patterns between the residual mean vertical velocity and the residual mean scalar components. As a result, selection of specific criteria for determining the increase in flux \( i.e. \) day vs night may cause bias in the slope of the regression of the fluxes.
Investigation into low frequency contributions was carried out in which the long-term averaging period employed in equation 5.26 was varied from 1 to 15 hours. This analysis was applied to the 1998 data using both the planar fit rotation and 2-angle run rotation methods. An example regression for an 8-hour long-term period, shown in figure 5.14 demonstrates the large amount of advective term scatter present in these relationships. This scatter increased with long-term averaging period and was considerably larger for latent heat and CO₂ flux, figure 5.15.

The analysis of the increase in fluxes over the 1 to 15 hour long term averaging periods, figure 5.16, suggests negligible contribution to sensible heat flux for any
long-term averaging period using either the planar fit or 2-angle run rotation methods. A similar lack of change is observed for momentum flux using the planar fit method while rather erratic and large changes in momentum are observed for the 2-angle rotation method. The behaviour of the momentum flux for the 2-angle rotation method points to possible weaknesses in this approach and should be investigated further.

Figure 5.16 Changes in net flux with increased low frequency flux information obtained by linear regression of flux with advective component against flux without advective component. The data in the left panel are obtained from planar fit rotation of the raw data while a run length 2-angle rotation was applied to the data in the right panel. Data are for the year 1998.

By increasing the contributions of low frequency flux corresponding to periods of from 1 to 15 hours, a linear increase in carbon dioxide flux was observed and a similar decrease in water vapour flux. For these fluxes, the 2-angle rotation method showed similar behaviour, though the flux contribution of the 2-angle method was consistently smaller for \( F_c \) and more negative for \( \lambda E \). Initially, we may discount that the low frequency trends are a fortuitous coincidence of non-stationary patterns of scalars and vertical velocity mean. If this were true a similar trend would be observed in the contribution to sensible heat flux. Therefore, the observed trends indicate that the low frequency flux contributions may be acting to transport drier, CO\(_2\) rich air away from the surface.

Further examination of these results by dividing the data into classes corresponding to environmental conditions revealed no consistent patterns between fluxes or between the two rotation methods. It may be possible that correlation between the residual
mean vertical velocity and mean scalar may result from arbitrary division of the data set.

It should also be noted that the scatter observed in the half hour flux values calculated using this methodology renders those flux values impracticable for direct use because of the rather large error, $\varepsilon$, associated with those short-term fluxes. Therefore this method may be appropriate for identifying low frequency flux components but is not capable of dividing and assigning those contributions to short-term flux periods.

5.5.2.10 Density effects

When applying the methods described in section 5.2.1, we assume a zero vertical wind velocity. Webb et al. (1980), hereafter WPL, have pointed out that a small residual mean vertical velocity does exist, and is caused by the density variations associated with the fluxes of heat and water vapour. More properly, equation 5.11 assumes, instead of a zero mean vertical velocity, that there is no mean vertical transport of dry air under steady state conditions. Under experimental conditions it is not possible to stipulate such steady state conditions. Instead, it is hypothesized that the correct assumption should be that there is no mean change in the number of molecules in an underlying column of air under conditions of temperature and pressure corresponding to initial conditions. A derivation of the resulting relationship is given in appendix B.2. This equation provides a framework to assess the sensitivity of this correction for relevant environmental factors.

\[
\frac{w}{\Delta t} = \frac{h}{\Delta t} \left[ \left( 1 + \frac{\Delta T}{T_i} \right) \left( 1 - \frac{\Delta P}{P_i} \right) - 1 \right] + \frac{w'n'}{n} + \frac{w'T'}{T} - \frac{w'P'}{P}
\]

Equation 5.27 generalizes the results of WPL to include the effect of storage on density changes in a volume of air underlying the flux measurement apparatus. The $\Delta T$ and $\Delta P$ terms represent the change in temperature and pressure in the volume and are straightforward in their interpretation. A term for storage of molecules within the column, $\Delta n$, does not appear because this term is zero by definition (i.e. there can be no mean change of molecules from the initial condition). This assumption is the
equivalent to the assumption of no transport of dry air under steady state conditions assumed by WPL and others.

The $\overline{w'T'}$, $\overline{w'P'}$ and $\overline{w'n'}$ terms represent the loss of heat, pressure or molecules from the volume by turbulent advective or diffusive transport. This loss of heat may be converted to an equivalent volume temperature change given the initial volume, pressure, etc. It can be assumed that the transport loss of pressure is negligible so that the $\overline{w'P'}$ term is excluded. The loss of molecules can be measured directly by trace gas flux measurement methods. It is likely that only the flux of water vapour molecules occurs in sufficient volume to have a measurable effect on the vertical velocity. The flux of CO$_2$ is measured and may be taken into account though the effect of CO$_2$ will be near zero because its transport will be balanced by the transport of oxygen molecules in the opposite direction (Hall & Rao 1999).

The effects of this modification to the density corrections are shown in figure 5.17. The sensitivity to individual equation components, excepting measurement height and averaging period, show increases in vertical velocity for increases in $H$, $\lambda E$, and $\Delta T$ and decreases in $\Delta P$. The effect of $H$ on $\overline{w}$ is more than four times greater than the effect of $\lambda E$ for an equivalent flux magnitude. The effects of storage at a mean column temperature change of (a rather large) 5 degrees is approximately a tenth of the effect of a sensible heat flux of 300 W m$^{-2}$ while the effect of a column pressure change is smaller still by an order of magnitude.

The interpretation of the revised density corrections with respect to measurement height and measurement period is more interesting. With respect to changes in measurement height, the resulting mean vertical velocity will decrease logarithmically with height when flux magnitudes are constant until reaching equivalent values obtained by WPL (1980) at a height of approximately 30 m. When storage values ($\Delta T = 1$ C and $\Delta P = 1$ kPa) are added into the relationship of mean $w$ to height, a minimum in mean $w$ is observed to correspond with a height of 10 m. This minimum is nearly twice that of the mean $w$ calculated using WPL equation 14.
Figure 5.17 In the top panel the curves show the sensitivity effects of model components. Initial conditions were set for a measurement height, \( h \), of 15 m, measurement period of 1800 seconds, initial temperature of 300 K, initial pressure of 100 kPa with no flux or storage terms. The variables were changed linearly over their specified range. In the bottom panel the sensitivity to measurement height and period are shown. For this panel the initial conditions were changed so that both \( H \) and \( E \) flux were 300 W m\(^{-2}\), column temperature change was 1 C and pressure change was –0.1 Kpa. The variables \( h \) and \( \Delta t \) were changed linearly over their specified range.

The relationship with run length exhibits no increase with increasing run length if flux magnitudes and storage are held constant. If storage terms are varied, a greater vertical velocity is observed for shorter run lengths. However, it should be noted that the storage terms are held constant for all run lengths. This may be unrealistic in that it is more likely that the storage terms would increase with run length from very small initial values.

The effect described by equation 5.27 and its analysis is generally considered specific to open path sensors. When employing closed path sensors we are typically measuring concentration values with respect to dry air and assuming that fluctuations of temperature and pressure have been reduced to negligible magnitudes. If these...
concentrations, in the process of calculation of fluxes, are converted to densities employing run mean temperature, and pressure there is no need for this correction to be applied. It must be remembered that if instantaneous values of temperature, pressure or molecular fluctuations are employed in converting concentrations it will be necessary to apply these corrections.

5.5.2.11 Frequency response

As mentioned in section 5.2.2.1, inaccuracies in measured fluxes can result from the inability of sensors to respond adequately to both very long and very short period signal fluctuations. This frequency response error can be quantified as the ratio of the measured signal’s spectra to the spectra of the true signal. This frequency dependent ratio is referred to as a transfer function and generally varies between values of 0 and 1, though values less than 0 and greater than 1 can occur. The attenuation of eddy covariance signals was noted early in the development of eddy covariance flux estimation techniques (McIlroy 1961). As different measurement systems have developed, so too have the theoretical methods for estimating the effect of frequency response attenuation.

Well-established models of transfer functions for frequency response of sensor time constant as well as analogue and digital filters exist because of their common use in other fields (Horowitz & Hill 1989; Ifeachor et al. 1993). More specialized transfer functions required by eddy covariance measurements have generally been developed in response to instrument advances. Transfer functions for the effect of signal path averaging and volume averaging have been derived in response to the development of sonic anemometers and other open path instruments (Horst 1973; Kristensen & Fitzjarrald 1984; Silverman 1968). Transfer functions for sensor separation have existed for some time (Kristensen 1979; Kristensen et al. 1979), though the complexity of atmospheric structure has made obtaining a definitive 3-D transfer function elusive (Kristensen et al. 1997). More recently, with the widespread use of closed path sensors, transfer functions have been developed describing the attenuation of trace gas fluctuations caused by transport down sampling tubes (Lenschow et al. 1990; Massman 1991).
The correction of fluxes for frequency response attenuation requires knowledge of both the true and attenuated cospectra. True cospectra are not measurable; therefore it is necessary to make assumptions about their form. The most common method of obtaining true cospectra is to assume a model cospectral shape based on measurements taken under ideal conditions. Generally, the models of spectral shape obtained from the Kansas experiment have been accepted as the good representation of true spectral shapes (Kaimal et al. 1972). In contrast to true cospectral shapes, attenuated cospectra are readily available by calculation from measurements.

For the purpose of frequency response corrections, it is more common to obtain attenuated cospectra by multiplying the model true cospectra with the frequency response transfer functions described above. The flux correction factor may then be obtained as the ratio of the integrated true model cospectra to the integrated attenuated model cospectra. Moore (1986) describes this procedure in an article that provides simplifications of some of the more complicated transfer functions for ease of application. More recently, further simplified methods have been developed which are based on analytical solutions to the integration process (Horst 2000; Massman 2000).

Although frequency response techniques are necessary for the appropriate estimation of fluxes, the analysis of the Griffin data set has led to the realization of the need for further development of the frequency response correction. Additional corrections have been developed to address the enhanced attenuation of gases in closed path systems, the effect of temporal averaging of signals, and the non-linearity of the correction process. The first two of these topics will be discussed in more detail in chapters 6 and 7, and will be only briefly discussed later in this chapter. The remainder of this section will describe the application of the frequency response corrections and address problems associated with non-linear averaging.

The procedures outlined by Moore (1986) formed the basis of the correction of the Griffin, flux data set. Because cospectra from Griffin did not show evidence of frequency shift under stable conditions (see sections 7.3), cospectral shapes based on the model presented by Massman (2000) were employed as representations of true
cospectra. The model cospectra were determined for 100 frequencies logarithmically spaced over the range of 0.0001 to 832 Hz.

The transfer functions employed in the frequency response corrections are given in table 5.2. The given transfer function variables refer to the following: $T_B$ - high pass filtering resulting from run length (Massman 2000), $T_{PV}$ - vector path averaging (Haugen et al. 1968), $T_{PS}$ - scalar path averaging (Silverman 1968), $T_1$ - sensor time constant (Moore 1986), $T_{PS}$ - temporal averaging (described in chapter 6), $T_{S}$ - path separation (Irwin 1979; Kristensen et al. 1979), $T_{TA}$ - tube attenuation (Leuning & Moncrieff 1990; Massman 1991), and $T_a$ - tube attenuation resulting from adsorption/desorption (described in section 5.3.2.7). The equations for transfer functions $T_B$, $T_{PV}$, $T_S$ and $T_{TA}$ are given in appendix B.14. For the correction calculations these transfer functions were evaluated for the same spectral frequencies as used in calculating the model cospectra.

**Table 5.2 Frequency response correction transfer functions applied to fluxes.**

<table>
<thead>
<tr>
<th>Flux</th>
<th>Transfer functions</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\bar{u}'w'$</td>
<td>$T_B T_{PV}^2 T_1^2$</td>
</tr>
<tr>
<td>$H$</td>
<td>$T_B T_{PV} T_{PS} T_1^2$</td>
</tr>
<tr>
<td>$H_{bc}$</td>
<td>$T_B T_{PV} T_1 T_i$</td>
</tr>
<tr>
<td>$F_c$</td>
<td>$T_B T_{PV} T_S T_{TA} T_i^2$</td>
</tr>
<tr>
<td>$\lambda E$</td>
<td>$T_B T_{PV} T_S T_{TA} T_a T_i^2$</td>
</tr>
</tbody>
</table>

For both path averaging transfer functions $T_{PV}$, $T_{PS}$, it was assumed that the paths of the sensors to which they are applied were oriented vertically and perpendicular to the mean flow. This is a commonly applied assumption, which may either over or under estimate because of the multiple measurement paths with inclined angles that are combined to obtain the vertical velocity (Silverman 1968). The path separation transfer function, $T_{PS}$, assumes similarity in the attenuation effects for both longitudinal and lateral separation of sensors relative to the mean wind. This assumption will be in error because the lag removal procedure eliminates the longitudinal separation while the magnitude of the lateral separation will vary with changing wind directions. Fortunately, this error will be minimized as the primary
separation is vertical with the IRGA intake located below the sonic anemometer (Kristensen et al. 1997). An assumption by the temporal integration transfer function, \( T_i \), that the sub-sampling rate of the sensors is sufficient to capture much of the signal at the corresponding higher frequencies, is discussed in chapter 6.

The combined transfer functions were applied to the modelled cospectra to obtain a shape for the attenuated cospectra. Both the model cospectra and attenuated cospectra were integrated using Simpson's rule. The correction factor was then obtained as the ratio of the integrated unattenuated cospectra to that of the integrated attenuated cospectra.

**Effect of water vapour transfer function**

The transfer function for tube attenuation of a fluctuating signal, (Lenschow et al. 1990; Leuning et al. 1990; Massman 1991; Philip 1963) describes the attenuation of gas concentration variations caused by diffusive and turbulent mixing when the gas sample flows through a tube. Although, this transfer function applies well to carbon dioxide, it has been pointed out that there is an additional attenuation for water vapour (Lee et al. 1994; Peters et al. 2001). Several approaches have been taken in trying to account for the additional attenuation of water vapour (Blanken et al. 1998; Villalobos 1997), most of which are either empirical or can be accounted for with the mean deposition velocity which Massman (1991) has included in his transfer function derivation for turbulent flow down a tube.

It is normally assumed that the attenuation of water vapour is similar to that of carbon dioxide. The only difference in the formulation of the transfer functions for these two gases in the attenuation coefficient, as a result of differing diffusive characteristics of the two gas species. It is commonly observed that the attenuation of water vapour occurs to lower frequencies than that of carbon dioxide. The analysis of water vapour lag in section 5.3.2.7 explains the rational behind the lag of water vapour as it travels down an intake tube. If we continue with our assumption that the adsorption/desorption process is a first order linear system we may further deduce the water vapour signal’s attenuation time constant from the increase in lag relative to a non-adsorbing species. From this assumption, a first order linear system’s signal lag time, \( \Delta t \), may be expressed in terms of frequency, \( f \), and its phase lag, \( \theta \), as
\[ \Delta t = \frac{\theta_i}{2\pi f} \]  

(5.28)

We may then employ the relationship of the phase lag to system time constant, \( \tau \), as:

\[ \tau = \frac{\tan(\theta_i)}{2\pi f} \]  

(5.29)

Combining these equations with equation 5.20 for the relationship of signal lag to relative humidity we may determine the signal attenuation time constant associated with the adsorption of water onto the walls of the intake tube.

\[ \tau = \tan \left( \frac{2\pi f}{0.0085 \cdot e^{0.38}} \right) \]  

(5.30)

At low frequencies, this function describes an exponential increase in signal attenuation time constant that varies between near zero to approximately 12 seconds at 100% humidity.

![Figure 5.18 Frequency response correction as a function of associated flux \((\lambda E\) filled circles, \(F_c\) open circles). All available data included, error bars represent one standard error.](image)

At low frequencies, this function describes an exponential increase in signal attenuation time constant that varies between near zero to approximately 12 seconds at 100% humidity.

Application of this time constant to the frequency response correction method results in sizable corrections in the water vapour flux signal. However, it should be noted
that large corrections are typically limited to nocturnal periods and rainy periods when relative humidity is high and fluxes of water vapour are naturally smaller, as shown in figure 5.18. A similar effect is not seen in $F_c$ because atmospheric CO$_2$ is not close to saturation (the slight rise in $F_c$ correction for negative fluxes is a result of the cospectral model shapes, see chapter 7).

**Effect of time averaging transfer function**

The effect of the temporal averaging by subsampling of signals was tested by comparing fluxes corrected for frequency response effects including and excluding this transfer function. The effect on the fluxes is presented as the change in flux observed in response to including the transfer function, figure 5.19. The effect is presented as a function of wind speed because it is the main determinant of this transfer function (via its presence in the model cospectra). It is observed that the resulting effect on fluxes is very small for the experimental setup employed at Griffin. Because other transfer functions were employed in this comparison some of the effects of time averaging will have been masked by the attenuation effects of those transfer functions.

![Figure 5.19](image)

*Figure 5.19 Change in fluxes of carbon dioxide, $F_c$, momentum, sensible heat $H$, and latent heat $\lambda E$, as a result of including the transfer function for the temporal averaging of signals in the frequency response correction.*
Effect of using neutral instead of stability dependent cospectra

Because cospectra did not exhibit the expected frequency shift under stable conditions, the effect upon frequency response corrections was examined by comparing corrections determined using only neutral Kaimal model cospectra with stable Kaimal model cospectra under stable conditions. The effect of this test, shown in figure 5.20, was to make the fluxes of $\lambda E$ and momentum more negative, $H$ more positive and to reduce the fluxes of $F_c$. The effect on the fluxes of momentum, carbon dioxide and sensible heat was quite small, generally less than 10% of the flux. However the effect for latent heat was much larger, on the order of 30% of the flux. The larger effect on latent heat results from a reduction in high frequency correction because of the lack of frequency shift when using the neutral cospectral model. (These results were obtained before the cospectra used in final analysis had been developed and are meant only as a qualitative example of the effects of modelled cospectra used in the final analysis).

![Figure 5.20 Change in fluxes of carbon dioxide, $F_c$, momentum, sensible heat $H$, and latent heat $\lambda E$, as a result of employing neutral model spectra under stable conditions in the frequency response correction. The observed changes presented in the figure are only those for stable conditions.](image-url)
Effect of non-linearity of frequency response correction

Another problem in the application of the frequency response correction technique is the assumption of linearity of the correction factor in response to controlling variables. Under unstable and neutral conditions, the frequency response correction factor is a function of wind speed (ignoring the effect of adsorption for \( \text{H}_2\text{O} \)), while under stable conditions the correction should be a function of both wind speed and stability. When applying the frequency response correction we assume that its response to controlling variables \( U \), and \( z/L \) over the period of application is linear. This assumption conflicts with our knowledge of the non-linearity of its relationship to wind speed and stability.

A simple test of this concept was carried out using a hypothetical run period with constant heat flux of 100 W m\(^{-2}\) and mean wind speed of 1.29 m s\(^{-1}\). A similar period with the same mean wind speed, but divided into 100 sub periods and with a distribution of wind speeds as shown in figure 5.2, the sensible heat flux was assumed to remain at 100 W m\(^{-2}\) for all sub-periods. It was also assumed that the low frequency behaviour of both the longer period and sub-periods corresponded to that described by the model cospectra used by Moore (1986).

![Figure 5.21](image)

Figure 5.21 Frequency distribution of wind speeds for sample demonstration of effect of non-linear averaging on frequency response correction. Comparison of model cospectra for average conditions, solid line, with average of model cospectra for varying conditions.

The flux for the entire period was corrected using Moore's (1986) frequency correction approach to obtain a corrected flux of 101.9 W m\(^{-2}\), consistent with our expectations of frequency response corrections of sensible heat flux. However, when the flux from each of the 100 sub periods was corrected and averaged to obtain an
equivalent average flux for the entire period. The magnitude of this corrected flux (108.4 W m^{-2}) was more than 5% higher.

The cause of this difference in correction factors is apparent from the difference in assumed shapes of the cospectra employed in the correction, (figure 5.21 right panel). If we use averaged conditions (wind speed, stability) in determining model cospectra we obtain a cospectral shape which is decidedly different than that which we obtain by averaging model cospectra evaluated for varying conditions within the longer period of interest. The resulting average of multiple model cospectra is flatter with proportionally more energy at both higher and lower frequencies. A measured true cospectra should exhibit similar flattening as a result of within run variability, as has been observed by Laubach and McNaughton (Laubach et al. 1998) and is suggested in the cospectral model of Massman (2000).

There are three possible solutions to this problem. The ideal solution would be to apply the inverse of the transfer functions directly to the attenuated cospectra, thus obtaining the corrected covariance. Unfortunately this solution is susceptible to calculation instabilities at high frequencies because of the large transfer function values required to account for the missing cospectral information.

A second solution would be to employ more realistic cospectral models that are representative of the true cospectra under all conditions. This option is ideal but is dependent upon the development of a theoretical spectral model capable of explaining spectral structure of the surface layer and ABL, and of incorporating temporal variability.

Alternately, a third, statistical, approach to the problem was developed. In this approach, an adjusted frequency response correction factor was obtained for neutral and unstable conditions by convolution of the frequency response correction factor’s relationship to wind speed with the normalized wind speed distribution curve. A similar approach was taken for stable conditions in which both wind speed and stability distribution curves were employed. For both neutral and stable conditions normal probability distributions were assumed with the distribution mean corresponding to the run mean wind speed or stability; no account was taken for
skewed or kurtotic distributions though they would also affect the effective shape of the resulting cospectra.

The validity of this approach depends upon the standard deviation of the distributions appropriately representing the within run non-stationarity. This information was obtained by comparing statistics from one-minute interval runs with those from one-hour runs. One-minute runs were employed because this length roughly corresponded to the low frequency end of the inertial subrange (Kaimal et al. 1972; Wyngaard 1990) and above which energy is primarily the result of the dissipation of energy instead of being subject to the variety of forcings that can cause non-stationarity.

The analysis of the one and sixty minute runs suggested that the standard deviation of the one minute run wind speeds $\sigma_{ui}$ accounted for approximately 60% of the standard deviations of wind speed in the sixty minute runs, $\sigma_u$, (equation 5.31). This result is consistent with the amount of spectral energy below the spectral peak frequency of wind speed.

$$\frac{\sum \sigma_{ui}}{n} = 0.6 \cdot \sigma_u$$ (5.31)

Based on this result, for final processing, the $\sigma_u$ associated with spectral variation was assumed to be 60% of run $\sigma_u$. The effect of incorporation of variability of wind speeds into the frequency response correction, (figure 5.22), suggests an increase in correction factors for $F_c$ (and $\lambda E$) with increasing wind non-stationarity at low wind speeds. For sensible heat flux (and momentum) there should be an increased correction with increasing wind non-stationarity at moderate to high wind speeds. At very low wind speeds, corrections reach a maximum at moderate non-stationarity before declining at high levels of wind non-stationarity.

Analysis of mean wind speeds also indicates an approximately 3% increase in the mean wind speeds for shorter runs. This increase is the result of rotation of the $u$ component velocity shorter periods into the mean wind direction for that period. While no attempt was made to adjust for this increase, it is acknowledged that this increase was greater for mean wind speeds closer to zero, a region at which the
correction factors are very responsive to the magnitude of the mean wind so that not accounting for this affect will cause some overestimate in the observed correction effect.

Figure 5.22 Frequency response correction factors for sensible heat flux and carbon dioxide flux as a function of wind speed for different values of velocity standard deviations. The curve with zero standard deviation (filled circles) represents the response curve evaluated for run mean wind speeds.

For stable conditions it was also necessary to determine a standard deviation value for the distribution of stability. Analysis of 1-minute and 30-minute period calculations of Monin-Obukhov stability, (figure 5.23), suggests different patterns for stable and unstable periods. During unstable diurnal periods, one-minute calculations exhibited less variability and were always of the same sign but with a negative skewness. In contrast, during stable nocturnal periods, one-minute stabilities exhibited much greater variability and less skewness. Analysis of this comparison revealed that the variability of the one-minute calculations of stability could be represented with a multi-linear regression against sensible heat flux and the natural logarithm of friction velocity.

\[
\sigma_{\varepsilon} \approx -0.09 - 0.9 \cdot H + \frac{\ln\left(u_{*}\right)}{350}
\]

(5.32)

For calculation expediency, it was assumed that positive deviations in stability corresponded linearly to decreases in mean velocity so that the same probability distribution convolution would apply for both velocity and stability.
Only the effects of wind speed and stability were accounted for in this correction factor adjustment. In the analysis of the Griffin data, it was assumed that relative humidity and latent heat fluxes were stationary over each run period. Considering the large errors associated with the effect of relative humidity on water vapour tube attenuation it is suggested to investigate the effect of within run variability of humidity upon the frequency response correction factor.

Similarly, within run variability of fluxes could also be considered by convoluting the resulting frequency response correction factor with the distribution of fluxes. Analysis of the deviations of sixty one-minute fluxes from the corresponding sixty-minute flux suggested that there was no relationship between the deviation of short period fluxes with the deviations of short period mean velocities and stability from their corresponding sixty-minute values (all relations exhibited $R^2$ values of less than 0.15). Such poor correlations indicate that convolution may not be an appropriate method of correction.

Functional responses of the frequency response correction to wind speed and stability, figure 5.24, indicate that modification of the correction for the effect of wind and stability distribution affects the frequency response correction factor primarily at wind speeds below 5 m s$^{-1}$. The modified correction factor for momentum is enhanced

---

Figure 5.23 Example of Monin-Obukhov stability calculations for 1 minute and 30 minute periods (May 22, 1998).
under stable conditions and reduced under unstable conditions. For carbon dioxide flux the correction factor is very similar under unstable conditions but is enhanced at low wind speeds and reduced at moderate wind speeds under stable conditions. (Note that the behaviour of the sensible heat flux and latent heat flux correction factors are similar to the modified momentum and carbon dioxide flux corrections, respectively).

![Image of frequency response correction factors](image)

**Figure 5.24** Frequency response correction factors in relation to wind speed and stability. Left panels represent correction factors for Griffin determined following Moore (1986). Right panels are same correction factors adjusted for the frequency distributions of low frequency wind speed and stability variations. Distributions of wind speed and stability used in determination of the modified correction factors were determined as approximate functions of wind speed.

The effect of the application of equations 5.31 and 5.32 to the frequency response correction calculations of momentum and $F_c$, based on measured means and modelled variability, are shown in figure 5.25. As expected, only slight increases in the correction factor are observed under unstable conditions and larger corrections observed under stable conditions.

When applied to their corresponding fluxes, the effect of incorporating $U$ and $z/L$ variability do not appear as dramatic. The effect on $\lambda E$ and $F_c$ is shown in figure 5.26, and it is observed that the effect appears primarily to introduce variability into the relationship with fluxes corrected using standard approach. Large biases in $\lambda E$
and \( F_c \) were not observed and the effects on \( H \) and momentum were even smaller. More in depth analysis is required if the variability results in more realistic flux values.

Figure 5.25 Frequency response correction factors for fluxes of momentum, (left panels) and carbon dioxide (right panels) under unstable (top panels) and stable (bottom panels) conditions. The filled circles represent correction factors following Moore (1986) while the open circles represent the correction factors adjusted for the low frequency distributions of wind speed and stability. Distributions of wind speed and stability were obtained as functions of the run standard deviation of wind speed and as a function of friction velocity and sensible heat flux, respectively.

Figure 5.26 Comparison of latent heat flux and carbon dioxide flux. The values on the x-axis are corrected following Moore (1991) while those on the y-axis employ the frequency response correction, which incorporates wind speed and stability variations.
Comparison of methods frequency response correction methods

The examples given in the preceding analysis of frequency response correction effects were based on the procedures presented by Moore (1986). However, as pointed out there are several assumptions incorporated into Moore’s method that suggest inaccuracies in the resulting correction factors. In order to determine what errors may be associated with this method, and to determine if other existing methods of frequency response correction may be more appropriate, a more in depth analysis of various frequency response correction methods was carried out. This analysis is presented in chapter 7. The conclusion of that chapter is that the numerical integration method is sufficient but that the cospectral models employed were inappropriate such that a greater sensitivity associated with high frequency attenuation of some scalar fluxes was observed.

Conclusions on frequency response corrections method

For the processing of the Griffin flux data the integration method presented by Moore (1986) was employed because it represents a commonly accepted approach. However, it is acknowledged that the results of the comparison of methods suggest that this approach may not provide a true representation of the correction effect. To minimize possible misrepresentation, cospectra models based on Massman’s (2000) model were employed. The temporal averaging transfer function was included on a theoretical basis even though the effect on fluxes was negligible. While no transfer function was developed for the adsorption/desorption of sample gases, an empirical representation using the sensor time constant transfer function was employed. The inclusion of this empirical transfer function was justified on the basis of its significant effect on fluxes. A further adjustment to the frequency response correction factors was applied based on the variability of wind and stability conditions. This correction factor adjustment was also of considerable magnitude under some conditions, though the effect on fluxes was not as large.

5.5.2.12 Flux Storage

The theoretical concepts of the storage term presented in section 5.2.1.5 are described in terms of the actual storage value calculations in the following section.
In all of the storage calculations employed, there is an implicit assumption that a single measurement is representative of a larger ensemble average. For storage terms involving storage within the air column we assume that the air is well mixed such that the storage calculations can be applied on a layer basis using a small number of sensors without loss of accuracy. For storage (of heat) within soil and biomass, the slow movement of temperature through these media make the assumption that they are well mixed less valid. However, the small diameter of much of the canopy biomass makes it possible to approximate its temperature variation using ambient air temperatures (Rinne et al. 1999; Saxton & McCaughey 1988), though the inherent temperature difference between the surface and air under conditions of sensible heat flux (Mahrt et al. 1997; Monteith 1973; Rosenberg et al. 1983) will inevitably result in storage term errors. For soil and bole storage it was assumed that radiation loading of these components within the canopy was low, see section 4.7, so that spatial variability was small and direct temperature measurements of these components was acceptably accurate.

The time period over which storage terms are calculated, $\Delta t$, should specify the period between two instantaneous measurements employed to calculate the difference variable required by the storage calculations. Instead of the preferred measurements at the beginning and end of the run period, only half hour averaged values were available from the Griffin data set. Therefore, beginning and ending values for each run were obtained by interpolation from these run average values. This procedure imposes an undesired filtering process on the storage calculations (J. Finnigan, per. com.); unfortunately appropriate measurements were not available and it was assumed that this filtering process imposed negligible effects.

The following set of equations describe the storage of sensible heat, $S_{Kin-air}$, latent heat, $S_{Pot-water}$, and carbon dioxide, $S_{CO_2}$, in the air column below the level of flux measurement. All three storage calculations were determined for a minimum of eight air layers, $\Delta z_k$, comprising the entire column depth, (figure 5.27). Each storage calculation also employed layer values of air density $\rho_A$ for conversion to flux values. The storage of sensible heat required measurements of the change in layer air
temperature, $\Delta T_a$, as well as the specific heat of air at constant pressure, $c_p$. The storage of latent heat and carbon dioxide were obtained from measurements of the change in concentrations of water vapour, $\Delta C_q$, and carbon dioxide, $\Delta C_C$. The latent heat storage also required layer values of the latent heat of vaporization, $\lambda$.

$$S_{\text{Kin-air}} = \sum_{i=1}^{n} \rho_{a_i} c_p \left( \frac{\Delta T_a}{\Delta t} \right) \Delta z_i$$  \hspace{1cm} (5.33)

$$S_{\text{Pot-water}} = \sum_{i=1}^{n} \lambda_{i} \rho_{a_i} \left( \frac{\Delta C_q}{\Delta t} \right) \Delta z_i$$  \hspace{1cm} (5.34)

$$S_{C_C} = \sum_{i=1}^{n} \rho_{a_i} \left( \frac{\Delta C_C}{\Delta t} \right) \Delta z_i$$  \hspace{1cm} (5.35)

**Profile heights and layers**

<table>
<thead>
<tr>
<th>Measurement Heights</th>
<th>Temperature</th>
<th>CO2 and H2O</th>
</tr>
</thead>
<tbody>
<tr>
<td>14.63 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>12.18 m</td>
<td></td>
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</tr>
<tr>
<td>9.46 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>7.91 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6.75 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3.38 m</td>
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</tr>
<tr>
<td>1.70 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.75 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.35 m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.05 m</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.27 Representation of profile measurement heights for $T_a$, CO$_2$ and H$_2$O over the period of the experiment.

For calculations of energy storage not associated with the air column layers were not defined but instead masses, $M$, associated with specific temperature change
measurements, $dT$, were defined. The following three equations describe the storage of heat in soil, water and biomass.

\[
S_{\text{Kin-soil}} = c_{\text{soil}} \rho_{\text{soil}} \frac{dT_s}{dt}
\]

\[
S_{\text{Kin-water}} = c_w \left( \frac{\theta_{wc} M_{\text{canopy}}}{1 - \theta_{wc}} \frac{dT_s}{dt} + \frac{\theta_{wb} M_{\text{bole}}}{1 - \theta_{wb}} \frac{dT_b}{dt} + \theta_p \rho_w \Delta z \frac{dT_s}{dt} \right)
\]

\[
S_{\text{Kin-biomass}} = c_b \left( \frac{M_{\text{canopy}}}{dt} + M_{\text{bole}} \frac{dT_b}{dt} \right)
\]

In these equations the variable $c_w$ represents the specific heat of water (4188 J kg$^{-1}$ K$^{-1}$) while the specific heat of biomass, $c_b$, is the sum of the specific heat of wood, $c_0$, (4.85 T + 1113 J kg$^{-1}$ K$^{-1}$) and the specific heat of wetting of hygroscopic materials, $\Delta c_h$ (335 J kg$^{-1}$ K$^{-1}$). The specific heat of soil, $c_{\text{soil}}$ (1920 J kg$^{-1}$ K$^{-1}$), was selected to be appropriate for an organic soil.

The canopy and bole dry matter masses were assumed to account for 60% and 40%, respectively, of total above ground biomass. The proportion of mass associated with boles was checked by calculating bole volume from canopy height and basal area, section 2.6.1, using standard mensuration tables (Hamilton 1988). This volume was then converted to biomass assuming a wood density of 380 kg m$^{-3}$. Daily values of above ground biomass, $W_a$, were determined using beginning of year above ground biomass, $W_{a0}$, annual biomass increment, $W_{ay}$, and day of year, $d_y$, assuming a sigmoid increase over the months of May through October (see section 8.3.3.3 and Appendix B.9).

\[
W_a = W_{a0} + W_{ay} - \frac{W_{ay}}{1 + \left( \frac{d_y}{182} \right)^3}
\]

Mass of water in the soil was obtained from the TDR determined volumetric water content, $\theta_v$, density of water, $\rho_w$, and depth of the soil heat flux plate, $\Delta z$. Volumetric soil moisture was measured only in the year 2001, section 4.16, and was assumed to have a similar pattern in other years. The gravimetric moisture content of the biomass was assumed to be a constant value of 0.6 for canopy biomass, $\theta_{wc}$, while the assumed
value of 0.3 for bole biomass, $\theta_{wb}$, was probably an underestimate (Agee et al. 2002; Hamilton 1988). Biomass water masses were then determined from relevant moisture content and associated dry biomass weight.

For soil heat storage, it was assumed that the soil heat flux plates accurately measured heat transfer to the soil below a depth of 0.05 m. Therefore, it was necessary to apply equations 5.36 and 5.37 to the top 0.05 m soil layer using soil mass, water content and temperature representative of that layer.

As discussed in section 5.2.1.5, the determinations of potential energy stored in biomass and soil have been combined because distinct measurements of photosynthesis and respiration for these two components were not available. Instead, the net exchange of carbon dioxide, $F_c$, was used as a surrogate for these mass changes. To convert this flux to appropriate heat units it was multiplied by the metabolic energy stored during photosynthesis and released in respiration, $E_m = -0.525 \text{ J \mu Mol}^{-1}$ (Jones 1992),

$$S_{\text{pot-biomass+soil}} = E_m F_c$$

(5.40)

The resulting storage values obtained for equations presented in this section and their patterns are presented in chapter 8.

5.5.2.13 Correction iteration

A final analysis examined the effect of recursion of correction procedures on flux values. In this analysis the frequency response and sonic temperature compensation corrections were iterated until the sum of the incremental change in all flux values became negligibly small; up to twelve iterations of the correction procedure were allowed.

The mean and range of the change in flux as a result of iteration are shown in figure 5.28. The results indicate that the net effect on sensible heat, momentum and carbon dioxide fluxes is very small but that values do exist for which the effect of iteration can be a few percent of typical daytime flux magnitudes. Latent heat flux exhibited
somewhat greater response to iteration, which may be a result of its greater sensitivity to the flux frequency response correction.

![Graph showing mean and range of change in flux as a result of iteration of flux correction procedures. Iterations beyond six are not shown as they contained one or no data point.](image1)

**Figure 5.28** Mean and range of change in flux as a result of iteration of flux correction procedures. Iterations beyond six are not shown as they contained one or no data point.

![Graph showing number of correction iterations required for all flux values to converge on constant values.](image2)

**Figure 5.29** Number of correction iterations required for all flux values to converge on constant values.

In general, less iteration was required for periods when corrections had smaller magnitude. Examination of the distribution of runs by iteration cycle, figure 5.29, further indicates that only very few runs required more than 5 iterations. Based on this observation it is recommended that for the processing procedures carried out as described in this chapter that iteration of the correction procedure is not necessary.
However, in final processing of results, a single iteration of the calculation of friction velocity, stability, and frequency response corrections was performed.

### 5.5.2.14 Final correction procedures

The final set of flux calculation procedures are outlined in table 5.3. All procedures listed, except application of storage corrections, were applied using the EdiRe software package. It should be noted that not all of the procedures described in this chapter are listed explicitly in this table, as some of the topics will be implicit components of the procedures listed in the table or were not capable of being applied as a correction to individual flux values. A more complete list of the correction procedure inputs relevant to EdiRe are provided in appendix F.3.

**Table 5.3 Signal processing procedure order used in determining corrected flux values.**

<table>
<thead>
<tr>
<th>Signals</th>
<th>Fluxes</th>
<th>( u )</th>
<th>( v )</th>
<th>( w )</th>
<th>( T )</th>
<th>( C_c )</th>
<th>( C_q )</th>
<th>( uw )</th>
<th>( H )</th>
<th>( \dot{\lambda}_E )</th>
<th>( F_c )</th>
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<td>Calibration</td>
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<td>( x )</td>
<td>( x )</td>
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<td>( x )</td>
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<td>( x )</td>
<td>( x )</td>
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<tr>
<td>Sonic attack angle correction</td>
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<td>( x )</td>
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<td>( x )</td>
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<td>Sonic temperature path correction</td>
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<td>Lag removal</td>
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<td>( x )</td>
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<tr>
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<td>( x )</td>
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<tr>
<td>Friction velocity/stability calculation</td>
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<td>( x )</td>
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<td>Frequency response correction</td>
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<td>( x )</td>
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<td>Repeat previous 2 steps</td>
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</tr>
<tr>
<td>Storage correction</td>
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<td>( x )</td>
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<td>( x )</td>
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</tbody>
</table>
5.6 Summary

A review of eddy covariance theory touches upon the critical assumptions employed and their implications for practical conduct of this experiment. The implications of the reduced transport equation for ecosystem exchange determination point to problems with the representativity of the storage term and frequency response of the flux term. In addition, signal quality of some sensors is addressed. Correction of sonic anemometer signals for angle of attack dependent calibration errors resulted in increases in fluxes of generally less than about 3%, though subsequent reanalysis suggests that the correction may have been larger. Despiking methods were applied to signals but were found to have an insignificant effect on the resulting fluxes. Anemometer coordinate rotations were applied using 2-angle and planar fit rotation methods. The difference in the effect on fluxes between these two methods was found to be quite small. The planar fit method was employed because of consistency of fit coefficients with sensor installation periods. Signal lag for CO2 and H2O was removed for each run. The resulting improvement in fluxes, as compared to assuming a constant lag time was small for CO2 but larger for H2O. The observed variability of the H2O lag was explained in terms of surface chemistry. The only filtering applied to signals was that implicit in block averaging for run length. Analysis of low frequency flux losses associated with this block averaging suggests significant flux exists at lower frequencies but that it can not be apportioned to particular runs. The use of closed path IRGAs for the measurement of H2O and CO2 made application of density corrections unnecessary, though analysis of this correction suggested small additional effects associated with storage terms. Several characteristics of frequency response corrections were examined. Empirical functions defining an attenuation term associated with surface chemistry was developed to account for the greater signal loss of latent heat flux. A response function was also developed to address temporal averaging by sensors. The non-linearity of the frequency response correction is investigated in terms of within run variability and its application to fluxes is presented. A qualitative examination of the effects of cospectral shapes, which show a smaller shift with stability then expected, is detailed. Iteration of frequency response corrections is examined and if found to be small for most situations though it may be occasionally large. Explicit storage equations are developed and related to their application forms.
6 The effect of time averaged sampling on fluxes

6.3 Introduction

The simplicity and directness of the eddy covariance method of determining atmospheric fluxes of mass and energy is dependent upon a few key assumptions. The eddy covariance method states that the instantaneous flux of an atmospheric property $p$ at a point in space $(x, y, z)$ at time $t$ is the product of the fluctuations of that property with the fluctuations in vertical wind velocity, $w'(x, y, z, t) \ p'(x, y, z, t)$. Because the resulting flux is itself a fluctuating quantity, it is usual to obtain an average over some time interval and to write this averaged flux as, $\bar{w}'\bar{p}'$. To implement this method exactly, we assume that a true 'vertical' velocity can be determined and that continuous and instantaneous measurements of $w'$ and $p'$ at exactly the same point in the flow are possible, as described in chapter 5.

Common use of three-dimensional sonic anemometers (Coppin & Taylor 1983; Fujitani et al. 1982; Kaimal & Businger 1963; Wyngaard 1981; Wyngaard et al. 1986) combined with techniques for coordinate rotation of velocity vectors (McMillen 1998; Wilczak et al. 2001) have addressed the problem of determination of a true vertical wind velocity. Solution of rotation has allowed for the proper determination of the fluctuating $w$ component.

Methods for measuring vertical velocity and other atmospheric properties, however, are characterized by finite temporal extent and resolution, as well as finite spatial positioning and resolution. These limitations compromise the assumptions underpinning the eddy covariance technique and result in the frequency response attenuations described in chapter 5.

Past studies have addressed most of the associated attenuation problems. For example, theoretical functions describing the attenuation of signal fluctuations caused by sensor spatial averaging have been addressed by Gurvich (1962), Silverman
(1968), Kaimal et. al. (1968), Kirstensen(1984), and Andreas (1981), while attenuation caused by sensor separation have been addressed by Kaimal et. al. (1968), Moore (1986), Kirstensen (1997), and Irwin (1979).

Temporal averaging of signals has historically been considered to occur as a result of the capacitive characteristics of sensors. Such characteristics result in the attenuation of a signal’s high frequencies, although the behaviour of such sensors is well known and the resulting signal losses have been well characterized, (Horst 2000; Moore 1986).

**Table 6.1 Sampling rates and sample output rates for some common micrometeorological sensors**

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Output rate</th>
<th>Sampling rate</th>
<th>Points averaged</th>
</tr>
</thead>
<tbody>
<tr>
<td>ATI</td>
<td>10 Hz</td>
<td>200 Hz</td>
<td>20, variable</td>
</tr>
<tr>
<td>Gill solent R3</td>
<td>10 - 50 Hz</td>
<td>100 Hz</td>
<td>6-30</td>
</tr>
<tr>
<td>Gill solent R2</td>
<td>20.8333 Hz</td>
<td>166.66 Hz</td>
<td>24</td>
</tr>
<tr>
<td>Campbell CSAT</td>
<td>10 or 20 Hz</td>
<td>60 Hz or</td>
<td>1-18</td>
</tr>
<tr>
<td>averaged output</td>
<td></td>
<td>matching</td>
<td></td>
</tr>
<tr>
<td>Metek</td>
<td>25 Hz max</td>
<td>25 Hz max</td>
<td>variable</td>
</tr>
<tr>
<td>RM Young</td>
<td>4 – 32 Hz</td>
<td>160 Hz</td>
<td>15-120</td>
</tr>
<tr>
<td>Licor 6252</td>
<td>10 Hz</td>
<td>500 Hz</td>
<td>50</td>
</tr>
<tr>
<td>Licor 7500</td>
<td>0 – 20 Hz</td>
<td>&gt; 300 Hz</td>
<td>&gt;15</td>
</tr>
<tr>
<td>TGA100</td>
<td>10 Hz</td>
<td>&gt;10 Hz</td>
<td>?</td>
</tr>
</tbody>
</table>

For other sensors, such as sonic anemometers, and radiation-absorption based instruments, such as IRGAs, the sensor’s time constant is a function of processing electronics and can be considered negligible. Many of these sensors employ sub-sampling to reduce noise in the sensor’s output signal. Some examples of current sensors’ signal output rate and sub-sampling rates are given in table 6.1. When subsampling is employed, a sensor’s output signal may be considered to correspond to a signal that has been averaged over the period between signal outputs. Such time averaging can be regarded as path averaging across a field of frozen turbulence. The unique feature of this averaging is that the length of the averaging path, \( l_x \) will depend
on the mean wind speed $u_1$ and subsampling time interval, $\Delta t$, (corresponding to the measurement system's sampling frequency $n_s = 1/\Delta t$).

$$I_\xi = u_1 \cdot \Delta t \tag{6.1}$$

As a consequence, the amount of spectral or cospectral attenuation observed by such sensors has the potential to be large at high wind speeds.

In this chapter, after first reviewing relevant spectral theory, temporal attenuation will be addressed by applying this theory to the problem of temporal averaging of a point sensor. This will be followed by application of the theory to more complicated measurement scenarios and a demonstration of the effect using data from the Griffin experiment.

### 6.4 Spectral theory

To address the signal attenuation that would result from the area averaging described above, spectral notation following (Andreas 1981; Haugen et al. 1968) will be employed.

An atmospheric property, $p_1$, which exhibits random fluctuations in both space and time may be described by a function $p_1(x,t)$. In this function, $t$ represents time, the subscript $i$ indicates the atmospheric property of interest, and the three orthogonal directions are represented by the vector $x$. An equivalent function describing the atmospheric quantity in wave number space and time is described by a function, $Z_1(k,t)$, where the vector, $k$, describes wave number space. The wave number vector $k$ has components $k_1$, $k_2$, and $k_3$ coinciding with the spatial coordinates $x_1$, $x_2$, and $x_3$. The spatial and wave number functions can be related through the Fourier-Stieltjes integral following (Lumley & Panofsky 1964).

$$p_1(x,t) = \int e^{ik \cdot x} dZ_1(k,t) \tag{6.2}$$

The three dimensional spectral density tensor, $\Phi_{ij}(k)$, is then obtained from the average of the complex conjugate, indicated by the asterisk in equation 6.3, of the
differential of the function $Z_i$. Note that the subscript $j$ indicates either the same or different atmospheric property and allows, $\Phi_{ij}$, to describe either the spectral or cospectral tensor, and that the time dependence of spectra will not be expressed explicitly in further equations.

$$\frac{dZ_i(k,t)dZ_j(k',t)}{dZ_i(k,t)dz} = \begin{cases} 0 & k' \neq k \\
\Phi_{ij}(k)d\kappa & k' = k 
\end{cases}$$ (6.3)

The one dimensional spectral density tensor, $F_{ij}(k_1)$, is obtained from the three dimensional tensor by integration over the two orthogonal wave number indices. It is the one-dimensional spectra that are commonly presented in the micrometeorological literature.

$$F_{ij}(k_1) = \int \Phi_{ij}(k)dk_2dk_3$$ (6.4)

Assuming homogeneous turbulence, the spectra in the $k_1$ wave number space may be considered representative of the one-dimensional spectra in the inertial sub-ranges of the two corresponding, orthogonal directions. It is therefore assumed that the attenuation associated with the transfer function to be derived will be limited to the inertial sub range.

### 6.5 Spatial equivalent of time averaging for a point sensor

For measurements obtained from a point sensor averaged over the period between signal outputs, and assuming Reynold’s frozen turbulence hypothesis, the measured signal time series can be related to the true signal by integration over the line characterized by the dimension, $l_x$ (see equation 6.1).

$$p_i''(x,t) = \frac{1}{l_x^2} \int_{-l_x/2}^{l_x/2} p_i(x + l_x, t)dl_x$$ (6.5)
Again, as for equation 6.2, we assume there exists a function $Z_m$ representing the measured variable $p_i^m$ in wave number space, which can be related to the measured variable $p_i^m$ through the Fourier-Stieltjes integral.

$$p_i^m (x,t) = \int_{-\infty}^{\infty} e^{ik \cdot x} dZ_i^m (k,t)$$ (6.6)

Employing the concept of equations 6.5 and 6.6 and integrating over the path length on both sides of equation 6.2, we obtain a representation of the Fourier-Stieltjes integral for the measured signal in terms of the true wave number function $Z_1$:

$$p_i^m (x,t) = \frac{1}{l_x} \int_{l_x/2}^{l_x} \int_{-\infty}^{\infty} e^{ik \cdot x} dZ_i (k,t)$$ (6.7)

Expansion and separation of the exponent provides the equation in the form

$$p_i^m (x,t) = \frac{1}{l_x} \int_{l_x/2}^{l_x} e^{ik \cdot x} \frac{dl_x}{l_x} \cdot \int_{-\infty}^{\infty} e^{ik \cdot y} dZ_i (k,t)$$ (6.8)

If the first term on the right hand side of equation 6.8 is expressed as the function $G_x$, then the equation may be restated as:

$$p_i^m (x,t) = \int_{-\infty}^{\infty} e^{ik \cdot x} \cdot G_x (k/l_x) \cdot dZ_i (k,t)$$ (6.9)

Using equations 6.6 and 6.9, the function $G_x$ defines the relation between the true and measured wave number functions:

$$dZ_i^m (k,t) = G_x (k/l_x) \cdot dZ_i (k,t)$$ (6.10)

Employing this relationship and equation 6.3, we may describe the relation between the true and measured three dimensional spectral representation as:
\[ \Phi_{ijm}(k) \, dk = G_x^2(k, l_x) \cdot \Phi_{ij}(k) \, dk \]  

(6.11)

As with equation 6.4, the one dimensional wave number spectra are defined in terms of the true three-dimensional spectra and the averaging function \( G_x \) as:

\[ F_{ij}^m(k_1) = \int \int G_x^2(k, l_x) \cdot \Phi_{ij}(k) \, dk_2 \, dk_3 \]  

(6.12)

Which when expressed in terms of only the one-dimensional spectra becomes,

\[ F_{ij}^m(k_1) = G_x^2(k_1, l_x) \cdot F_{ij}(k_1) \]  

(6.13)

The frequency dependent function describing the attenuation of the experimentally measured signal, commonly referred to as the transfer function, \( T(k_1) \), can be described using the ratio of the measured to true one dimensional spectra:

\[ T_{ij}(k_1) = G_x^2(k_1, l_x) = \frac{F_{ij}^m(k_1)}{F_{ij}(k_1)} \]  

(6.14)

The function derived by Gurvich (1962) (see also Andreas 1981; Kaimal 1968; Mitsuta 1966) describes the linear averaging of a signal, defined by equation 6.14, as a function of path length, \( l_s \), and path elevation angle, \( \beta_p \), with respect to a unit direction.

\[ G^2 = 0.142 \int_0^\infty \left[ 1 + \left( \frac{l_s / \pi f - \cos \beta_p}{\sin \beta_p} \right)^2 \right]^{4/3} \cdot \sin^2 \frac{l_s}{l_s} \, dl_s \]  

(6.15)

The solution of this equation (Mitsuta 1966; Silverman 1968) at \( \alpha_p = 0 \) provides a solution for the transfer function \( T_{xx} \).

\[ G_x^2(k, l_x) = T_{xx}(k_1) = \left( \frac{\sin(a_x)}{a_x} \right)^2 \]  

(6.16)
In this equation \( a_k = \pi n l_k / U_k \) and \( U_k \) is the velocity component corresponding to vector \( k_1 \). The results of a sensitivity analysis of this transfer function are presented in figure 6.1.

Figure 6.1 Sensitivity relations of signal attenuation correction factors from equation 6.16. Temperature \((T = 15)\) and roughness length \((z_0 = 0.1 \text{ m})\) were held constant in all plots. When not used as a variable sampling frequency was held at 10 Hz, and measurement height was set to 10 m. Wind speed was held constant at 3 m s\(^{-1}\) \((u_r = 0.16 \text{ m s}^{-1})\) in the sampling frequency and stability relation plots; it was allowed to vary logarithmically with height in the height relationship plot. Stability was determined by setting sensible heat flux to either \(-25 \text{ W m}^{-2}\) (stable) or \(250 \text{ W m}^{-2}\) (unstable) in all plots except the stability relationship in which stability was determined by varying \( H \) from \(-205\) to \(255 \text{ W m}^{-2}\). Correction factors were determined from the ratio of numerically integrated unattenuated to attenuated Kaimal (1972) scalar cospectra.

For this analysis, parameters were held constant unless there were known to vary along with the sensitivity variable. From this analysis it is apparent, for sampling frequencies higher than 10 Hz, and wind speeds less than 1 m s\(^{-1}\), that the effect of temporal averaging is relatively small (< 10%). At higher wind speeds, attenuation is only increased dramatically under stable conditions, as a result of the shift in cospectra to higher frequencies. The variation with height lies primarily in the relationship of height and wind speed (i.e. if wind speed is held constant and height is changed there is only slight variation in the resulting correction factor). Similarly, the
effect of stability (through sensible heat flux) exhibits a step change as a result of changes in cospectral models but does not exhibit significant changes as a direct result of stability changes. It is clear from this analysis that maintaining sampling frequencies above 10 Hz, will minimize this error to a few percent for most situations.

6.6 Caveats and complicating factors

While the solutions for temporal averaging described by equation 6.16 suggests that significant improvements can be obtained by its application, it is unlikely that any practical, field deployable sensor will consist of such a simple temporally averaged point measurement. The following derivations explore some of the possible complications resulting from variations in instrument dimensions and combinations.

6.6.1 Time averaged closed path sensors

Although a closed path sensor has a well-defined sampling volume, if the sensor is subsampled a cylinder swept past the inlet by the wind will more closely represent its sampling volume. The diameter of this cylinder is determined by the ratio of sensor’s inlet flow rate, $Q_t$, to the ambient wind speed.

$$l_y = \frac{Q_t}{u_1 \pi}$$  \hspace{1cm} (6.17),

For a typical inlet flow rate ($Q_t = 10$ LPM) the dimensions of the cylinder of averaging, for a range of wind speeds, are shown in figure 6.2. At very low wind speeds, the dimensions of this cylinder will be similar to those defined by the sensors sampling volume, or as an equivalently sized sphere as suggested by Massman (2000). It is clear, however, that for all wind speeds except those $\ll 1$ m s$^{-1}$ the cylindrical diameter becomes insignificant compared to the length of the cylinder. It is therefore suggested that a subsampled closed path sensors may be treated as a horizontal line-averaged sensor ($l_x = u_1 \Delta t$, $l_y \approx 0$), unless the ratio of sample flow rate to wind speed (in units of LPM/ m s$^{-1}$) is greater than $\sim 100$. Based on this assumption, equation 6.16 may be employed as the path averaging transfer function relevant to time averaged closed path sensors.
Figure 6.2 Comparison of length and diameter dimensions of a cylindrical area of sampling resulting from extraction of air at a constant flow rate, from a point in the atmosphere. Conditions assume a sub-sampling period of 0.1 s, though higher or lower sampling rates will not affect the proportion of dimensions, only their magnitudes.

6.6.2 Sensors which are both time and path averaged

The development of transfer functions for path averaging sensors has already been carried out by several authors (Gurvich 1962; Kaimal 1968; Mitsuta 1966). However, in their derivation of path averaging effects, these authors have assumed that measurements over the path of the sensor occur instantaneously. From table 6.1 we know that sonic anemometers and open path spectrometers are both subject to coincident time and path averaging. The following section develops the spectral theory relevant to a spectral attenuation function combining both path and time averaging.

Following the logic used, in section 6.3, to describe a subsampled point sensor as a line averaging sensor when exposed to advection, we may similarly describe a subsampled, line-averaging sensor as an area averaging sensor if it is exposed to advection. It is only under the rather unusual conditions of zero mean wind speed or the sensing path lying parallel to the mean wind, that such a sensor will behave as a linear averaging sensor. If such a sensors sampling volume lies perpendicular to the mean flow, the area swept by the wind will describe a rectangle, whereas if it is inclined with respect to the mean flow the area swept in time $\Delta t$ will describe a rhombus. For the purposes of this derivation we will treat such rhombi as their equivalent rectangular areas defined by the height and length of the rhombus.
We have already defined the streamwise extent of the rectangular area, \( l_x \), in equation 6.1. The corresponding cross-stream extent, \( l_y \), may be obtained from the elevation angle of the sensor's measurement path, \( l_s \), relative to the mean horizontal wind, \( \beta_p \), and the horizontal angle between the mean wind direction and the vertical plane dissecting a sensor's inclined measurement path, \( \alpha_p \):

\[
l_y = l_s \left[ \sin(\beta_p) + |\sin(\alpha_p)| - \sin(\beta_p) \cdot |\sin(\alpha_p)| \right]
\]  

(6.18)

For measurements comprising a rectangular plane surface, the measured signal time series is related to the true signal by its integration over a plane characterized by the dimensions, \( l_x \) and \( l_y \):

\[
p_t^m(x, t) = \frac{1}{l_x l_y} \int_{-l_y/2}^{l_y/2} \int_{-l_x/2}^{l_x/2} p_t(x + l_x + l_y, t) \, dl_x \, dl_y
\]  

(6.19)

Employing the concept of equation 6.7 by integrating equation 6.2 over both sides of the rectangle, we obtain a representation of the Fourier-Stieltjes integral for the measured signal in terms of the true wave number function \( Z_t \):

\[
p_t^m(x, t) = \frac{1}{l_x l_y} \int_{-l_y/2}^{l_y/2} \int_{-l_x/2}^{l_x/2} e^{i(x + l_x + l_y)k} \, dl_x \, dl_y
\]  

(6.20)

which may be separated similar to equation 6.8 to give:

\[
p_t^m(x, t) = \frac{1}{l_x} \int_{-l_y/2}^{l_y/2} e^{i(x + l_y)k} \, dl_y \cdot \frac{1}{l_y} \int_{-l_x/2}^{l_x/2} e^{i(x + l_x)k} \, dl_x \cdot \int_{-\infty}^{\infty} e^{i\kappa \cdot x} \, dZ_t(k, t)
\]  

(6.21)
If, following equation 6.9, the first two terms on the right hand side of equation 6.21 are expressed as the functions \( G_x \) and \( G_y \), equation 6.21 may then be restated as:

\[
P_{i}^{m}(x,t) = \int_{-\infty}^{\infty} e^{i(\mathbf{k} \cdot \mathbf{x})} G_x(k_x, l_x) G_y(k_y, l_y) dZ_1(k, t)
\]  

(6.22)

Similar to a time averaged point sensor, the relation between the true and measured wave number functions of an area-averaged sensor is defined by a product of the functions \( G_x \) and \( G_y \). When expressed in terms of one dimensional spectra equation 6.22 becomes,

\[
F_{ij}^{m}(k_1) = G_x^2(k_1, l_x) G_y^2(k_1, l_y) F_{ij}(k_1)
\]

(6.23)

The frequency dependent transfer function may again be described using the ratio of the measured to true one-dimensional spectra:

\[
T_{ij}(k_1) = G_x^2(k_1, l_x) G_y^2(k_1, l_y) \frac{F_{ij}^{m}(k_1)}{F_{ij}(k_1)}
\]

(6.24)

Moore (1986): has given approximate functions describing the transfer function for a vertical sensing path (\( \beta_y = 90 \)) of an anemometer (equation 6.25) and scalar sensor (equation 6.26), in which \( a_y = 2\pi n l_y \ U^{-1} \).

\[
G_y^2(k, l_y) = \frac{2}{a_y} \left( 2 + e^{-a_y} - \frac{3}{a_y} (1 - e^{-a_y}) \right)
\]

(6.25)

\[
G_y^2(k, l_y) = \frac{1}{a_y} \left( 3 + e^{-a_y} - \frac{4}{a_y} (1 - e^{-a_y}) \right)
\]

(6.26)

It is acknowledged that these functions are approximations of the cross-stream extent of the rectangular averaging area because the value \( l_y \) defined by equation 6.18 does not necessarily lie in the vertical plane. For simplicity, it is assumed that the non-vertical cross-stream spectral behaviour is similar to the vertical cross-stream
behaviour so that equation 6.25 may be used to describe the attenuation of both conditions.

Having already defined the function for \( G_x \) in equation 6.16 we may apply it and either equation 6.25 (or 6.26) to equation 6.24 to obtain a transfer function describing the attenuation caused by the rectangular averaging of a subsampled, linear-averaged signal subject to advection.

\[
T_{xy}(k_1) = \frac{2 \cdot \sin^2(a_x)}{a_y \cdot a_x^2} \left( 2 + e^{-a_y} - \frac{3}{a_y} \left( 1 - e^{-a_y} \right) \right) \tag{6.27}
\]

### 6.7 Transfer function equation for symmetric arrays

Although equation 6.27 should appropriately define the area averaging of a single-path, subsampled sensor, the configuration of many sonic anemometers do not adhere to this simple scenario. A currently popular configuration for 3-D sonic transducer arrays consists of three pairs of transducers, each of which has a sampling path zenith angle of \( \beta_p \) and a 120-degree azimuth separation from the other transducer pairs. For such configurations the velocity components are described in terms of the transducer path velocities, \( V_a, V_b, \) and \( V_c \), by the equations:

\[
u = \frac{1}{3} \left( \frac{V_a - V_b - V_c}{\cos(\beta_p)} + \frac{-V_b - V_c}{\cos(\beta_p) \cdot \cos(60)} \right) \tag{6.28}
\]

\[
\begin{align*}
\nu &= \frac{V_b - V_c}{2 \cdot \cos(\beta_p) \cdot \cos(30)} \tag{6.29} \\
w &= \frac{V_a + V_b + V_c}{3 \cdot \sin(\beta_p)} \tag{6.30}
\end{align*}
\]

For simplicity, we assume that the anemometers installation is such that it is mounted perpendicular to a horizontal surface and that the mean wind is aligned with the sensors \( u \) component. Using the vertical velocity component as an example, we may describe the measured vertical velocity as a combination of the measured transducer pair velocities:
\[ w^m = \frac{1}{3} \left( \frac{V_a^m}{\sin(\beta_p)} + \frac{V_b^m}{\sin(\beta_p)} + \frac{V_c^m}{\sin(\beta_p)} \right) \]  

(6.31)

By examining only the vertical component of these velocities we may express the measured vertical velocity as an average of the three transducer-pair velocities', vertical velocity components:

\[ w^m = \frac{1}{3} \left( w_a^m + w_b^m + w_c^m \right) \]  

(6.32)

Relating the measured vertical velocity to the wave number component, we express its dependence on the horizontal and vertical dimensions involved in the attenuation of each of the component velocities.

\[ w^m (x, l_x, l_y, l_z) = \int_{-\infty}^{\infty} e^{ik \cdot x} dZ_w^m (k, l_x, l_y, l_z) \]  

(6.33)

Following equations 6.21 and 6.22, each of the vertical velocity components of the transducer pair velocities may be written in terms of the unattenuated wave number function and the functions \( G_x \) and \( G_y \), corresponding to each vertical velocity measurement.

\[ w_a^m (x, l_x, l_y) = \int_{-\infty}^{\infty} e^{ik \cdot x} \cdot G_{xa} \cdot G_{ya} \cdot dZ_w (k) \]  

(6.34)

\[ w_b^m (x, l_x, l_y) = \int_{-\infty}^{\infty} e^{ik \cdot x} \cdot G_{xb} \cdot G_{yb} \cdot dZ_w (k) \]  

(6.35)

\[ w_c^m (x, l_x, l_y) = \int_{-\infty}^{\infty} e^{ik \cdot x} \cdot G_{xc} \cdot G_{yc} \cdot dZ_w (k) \]  

(6.36)

The measured wave number function then becomes a function of the averages of the attenuated vertical components of the wave number functions. For compactness, we have replaced the product of \( G_x \) and \( G_y \) with the function \( G_{xy} \).

\[ dZ_w^m = \frac{1}{3} \left( G_{xya} dZ_{wa} + G_{xyb} dZ_{wb} + G_{xyc} dZ_{wc} \right) \]  

(6.37)
Expressing the measured spectra as the complex conjugate of the true wave number functions and corresponding transfer functions,

\[
\Phi_{ww}^m = \left[ \left( \frac{G_{xya} dZ_{wa} + G_{xyb} dZ_{wb} + G_{xyc} dZ_{wc}}{3} \right) \right]^* \left[ \left( \frac{G_{xya} dZ_{wa} + G_{xyb} dZ_{wb} + G_{xyc} dZ_{wc}}{3} \right) \right]
\]

(6.38)

With no complex components to the attenuation factors, the vertical velocity spectra may be expressed as a combination of attenuation terms and component three-dimensional spectral and cospectral tensors.

\[
\Phi_{ww}^m = \frac{1}{9} \left( T_{xya} \Phi_{wwa} + T_{xyb} \Phi_{wwb} + T_{xyc} \Phi_{wwc} + 2G_{xya}G_{xyb} \Phi_{wawb} + 2G_{xyb}G_{xyc} \Phi_{wawc} + 2G_{xya}G_{xyc} \Phi_{wbwc} \right)
\]

(6.39)

If we now apply the simplifying assumption that the attenuation caused by the separation between transducer paths as a result of their azimuth orientations is negligible, we may then assume that the spectral and cospectral tensors can be considered identical and we may then extract them to produce the relationship

\[
\Phi_{ww}^m = \frac{1}{9} \left( G_{xya} + G_{xyb} + G_{xyc} \right)^2 \Phi_{ww}
\]

(6.40)

which gives the transfer function for w velocity component

\[
T_{xyw} = \frac{1}{9} \left( G_{xya} + G_{xyb} + G_{xyc} \right)^2
\]

(6.41)

When expressed using equations 6.16 and 6.25 this transfer function can be expressed as:

\[
T_{ww}(k) = \frac{\sin(a_x)}{9 \cdot a_x} \left( \sum_i \left[ \frac{2}{a_{yi}} \left( 2 + e^{-a_{yi}} - \frac{3}{a_{yi}}(1-e^{-a_{yi}}) \right) \right] \right)^2
\]

(6.42)
in which the summation over index \( i \) represents iteration over the three cross stream paths of the sonic’s three transducer pairs.

Figure 6.3 Comparison of vector path averaging transfer function (filled circles) with subsampled vector sensor transfer functions for a vertically oriented sensor, equ. 6.27 (open circles) and an array of three sensing paths, equ. 6.42 (filled inverted triangles). For equ 6.42, the angle \( \beta_p \) was assumed to be 45 degrees, and \( \alpha_p \) was \(-120\), \(0\), and \(120\) for the three transducer pairs.

Comparison of the transfer functions for a three-dimensional array (equ. 6.42) a vertical velocity sensor (equ. 6.27) and Moore’s (1986) vector path averaging attenuation approximation (equ. 6.25) is presented in figure 6.3. The transfer functions were evaluated for two wind speeds (1 and 5 m s\(^{-1}\)). Most apparent is the difference between the vector path averaging transfer function and the two approaches that include area averaging. Closer analysis of equation 6.42 indicated that the slightly higher cutoff frequency of the sonic array area averaging transfer function vertical and relative to the vertical area averaging transfer function was a result of the simplification of assuming negligible interaction between the array averaging paths used in obtaining equation 6.40. It is recommended that the vertical area averaging transfer function (equation 6.27) provides a good estimate of the more complicated array area averaging function. Similar logic can be used to suggest that equation 6.27 can be employed for U and V components in a three-dimensional sonic array, and equation 6.27 (incorporating equ. 6.26) may be used for scalar sensors.
6.8 Summary

Evaluation of the sensitivity of the frequency response correction resulting from the application of equation 6.27 was done using Kaimals (1972) model cospectral shapes. The procedure outlined by Moore (1986) for evaluating the correction factor was employed in which the transfer function for run length was incorporated as well the path averaging correction for both non-subsampled or subsampled sensors at both 10 and 20 Hz signal output frequencies. The difference in flux magnitudes resulting from assuming non-subsampled or subsampled sensors is plotted against flux magnitude in figure 6.4. Results for the effect of temporal averaging of other fluxes are presented in chapter 5.

Figure 6.4 Effect of inclusion of temporal averaging modified path averaging transfer function transfer on fluxes of momentum (left panel) and sensible heat flux (right panel). Changes in flux are plotted against corresponding flux values. Effect of temporal averaging was evaluated for both 10 Hz (open circles) and 20.83 Hz (filled circles) sampling frequencies.

It is apparent from figure 6.4 and from the sensitivity analysis shown in figure 6.1 that, under conditions used in this analysis, the effect of temporal averaging is quite small. However, it is noted that under some circumstances this error has the potential to become a significant proportion of the associated flux. It is recommended, when calculating flux frequency response corrections, that this additional attenuation term be included for the sake of completeness.
7 Comparison of frequency response correction methods

7.3 Introduction

The determination of frequency dependent flux loss requires either knowledge of the proportion of flux attenuation at all frequencies, the spectral behaviour of the true flux, or the attenuated (i.e. measured) flux at corresponding frequencies. Hypothetically, any two of these three quantities may be combined to obtain the correction to a measured flux. Unfortunately, attenuation often reduces fluxes at high frequencies to such low magnitudes that corresponding true flux values cannot be recovered, with accuracy, using knowledge of the spectral characteristics of attenuation. Thus frequency response correction methods are forced to presume knowledge of the spectral behaviour of true fluxes, employing either the spectral characteristics of attenuation or the spectral behaviour of measured fluxes to determine the amount of flux loss.

Currently, the most commonly accepted method for the correction of fluxes for frequency attenuation losses requires numerical integration of modelled true cospectra and attenuated model cospectra (Aubinet et al. 2000; Moore 1986). As mentioned, this method of correction requires implicit knowledge of both cospectral shape and transfer function behaviour. To more fully investigate the ability of this currently accepted method, a series of comparisons with similar correction methodologies was conducted. The purpose of this comparison was to identify any possible failings of the current numerical integration method and to see if alternate methods are capable of providing a truer representation of frequency response correction factors.

7.4 Calculation methods and description of comparison scheme nomenclature

Although almost all frequency response correction methods are conceptually identical, the details of their implementation are often quite different. For convenience we have divided the techniques into three categories: integration, analytical, and pass band ratio.
The integration method (see Moore (1986)) employs a cospectral shape for an assumed unattenuated flux and a frequency dependent transfer function describing the proportion of true flux lost at different frequencies. The true cospectra and transfer function are multiplied to obtain an attenuated cospectral shape. The integrated true cospectra are divided by the attenuated cospectra and multiplied by the attenuated flux to obtain a true flux, equation 7.1. In this chapter, ten variations of this method were obtained by employing differing sources of ‘true’ cospectra and transfer functions. A summary of these method variations is listed in table 7.1.

\[
\frac{w'x'}{w'x'_{\text{atten}}} = \sum \frac{C_{wx}}{T \cdot C_{wx_{\text{atten}}}}
\]

(7.1)

The analytical method represents a more recent version of the integration method (Massman 2000). This approach provides an analytical solution to the integration method, resulting in a reduction of the number of calculations required to obtain a flux correction value. A single variation of this method was employed. The reader is referred to Massman’s paper (Massman 2000) for a mathematical description of this method.

The pass band ratio method (Billesbach et al. 1992; Hicks & McMillen 1988; Mestayer et al. 1990) does not use a transfer function but instead employs the true and attenuated cospectra directly. It is assumed that a spectral region exists over which the attenuated cospectra suffers no attenuation. The ratio of the attenuated to true fluxes within this spectral region is multiplied by the true flux to obtain the corrected attenuated flux, equation 7.2.

\[
\frac{w'x'}{w'x'_{\text{atten}}} = \sum \frac{C_{wx}}{\sum_{a}^{b} C_{wx_{\text{atten}}}}
\]

(7.2)

In all, fourteen variations of methods of frequency dependent flux correction were applied to the Griffin data for the period from July – September 2000. This period was selected because it included coincident open and closed path CO₂/H₂O measurements. The comparison scheme and identification of sources for the components employed for the variations on the methods employed are given in table 7.1. Further description of the
sources and methods for determining the correction components are described below. A quantitative description of the comparison results will follow.

**Table 7.1 Frequency response correction methodologies. Each method consists of a calculation method, reference spectra source and transfer function source.**

<table>
<thead>
<tr>
<th>Label</th>
<th>Calculation Method</th>
<th>Reference Cospectra</th>
<th>Transfer function</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Integration</td>
<td>Moore*</td>
<td>Model*</td>
</tr>
<tr>
<td>B</td>
<td>Integration</td>
<td>Horst/Massman fitted**</td>
<td>Model</td>
</tr>
<tr>
<td>C</td>
<td>Integration</td>
<td>WT cospectra</td>
<td>Model</td>
</tr>
<tr>
<td>D</td>
<td>Integration</td>
<td>WT cospectra (5 yr avg)</td>
<td>Model</td>
</tr>
<tr>
<td>E</td>
<td>Integration</td>
<td>Moore</td>
<td>Average of cospectral ratios</td>
</tr>
<tr>
<td>F</td>
<td>Integration</td>
<td>Moore</td>
<td>Ratio of average cospectra</td>
</tr>
<tr>
<td>G</td>
<td>Integration</td>
<td>Moore</td>
<td>Calibration time constant</td>
</tr>
<tr>
<td>H</td>
<td>Integration</td>
<td>Moore</td>
<td>Time constant from filtered sonic temperature, by run</td>
</tr>
<tr>
<td>I</td>
<td>Integration</td>
<td>Moore</td>
<td>Time constant from filtered sonic temperature, average</td>
</tr>
<tr>
<td>J</td>
<td>Analytical</td>
<td>Horst/Massman fitted</td>
<td>Model</td>
</tr>
<tr>
<td>K</td>
<td>Integration</td>
<td>Open path C or Q cospectra</td>
<td>Model</td>
</tr>
<tr>
<td>L</td>
<td>Pass Band Ratio</td>
<td>WT cospectra</td>
<td>N/A</td>
</tr>
<tr>
<td>M</td>
<td>Pass Band Ratio</td>
<td>WT cospectra (5 yr avg)</td>
<td>N/A</td>
</tr>
<tr>
<td>N</td>
<td>Pass Band Ratio</td>
<td>Kaimal Model***</td>
<td>N/A</td>
</tr>
</tbody>
</table>

* “Moore” and “Model” obtained from Moore (1986)
** “Horst/Massman fitted” obtained from Massman (2000)
*** “Kaimal Model” obtained from Kaimal et al. (1972)

### 7.5 Reference cospectra determination

Cospectra employed in the frequency response correction process were either obtained from spectral models or calculated using fast Fourier transform techniques; when specified, averaged cospectral shapes were employed. A description of the model cospectra, cospectral calculations and cospectral averaging processes are given below.
7.5.1 Model cospectra

Two cospectral models were employed. The primary cospectral model was that of Kaimal et al. (1972). A model based on flat terrain cospectra averaged by stability conditions. As such, this model cospectrum suffers from the weakness that it was developed for experimental situations closer to the hypothetical ideal than the experimental site for which this model was applied. The version of these models employed by Moore (1986) was used for the analysis in this chapter. Slight differences between the Moore and Kaimal models were observed, as is shown in figures 7.1 and 7.2. The model equations for scalar flux as applied to neutral and unstable conditions, and for stable conditions are described in appendix B.

As an alternative to the Kaimal model, we employed the simplified model spectral model developed by Horst (2000) and Massman (2000). The equation defining this cospectral model is given in equation 7.3.

\[
C_{uu}(f) = N \frac{f / f_s}{1 + (f / f_s)^{2 \mu}}^{m+1/2 \mu m}
\]

where \(f_s\) is the spectral peak frequency, \(\mu\) is a parameter controlling spectral broadness, \(m\) is a parameter defining the slope of the inertial subrange and \(N\) is a normalization coefficient. Empirical values of \(f_s\), \(\mu\) and \(N\) were obtained by non-linear fit (SAS) of equation 7.3 to sensible heat flux cospectra averaged by stability classes (see next section). The resulting empirical relations for spectral peak frequency, broadness and gain factors are given in table 7.2.

<table>
<thead>
<tr>
<th>Stability</th>
<th>(f_s) (\mu)</th>
<th>(N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(\overline{w'T'}) Unstable</td>
<td>0.08</td>
<td>0.33</td>
</tr>
<tr>
<td>Stable</td>
<td>0.08 + 0.13*(z/L)</td>
<td>0.4</td>
</tr>
<tr>
<td>(\overline{u'w'}) Unstable</td>
<td>0.04</td>
<td>0.65</td>
</tr>
<tr>
<td>Stable</td>
<td>0.08 - 0.01*(z/L) (0.04 \text{ if } z/L &gt; 4)</td>
<td>0.65 - 0.29*(z/L) (0.36 \text{ if } z/L &gt; 1)</td>
</tr>
</tbody>
</table>
Figure 7.1 Comparison of Kaimal, Moore and Massman scalar cospectra models for stable and unstable conditions. Stable conditions are evaluated for Monin Obukhov stability of 1.0.

Figure 7.2 Same as figure 7.1 but for UW cospectral models
Examples of the relation of cospectral peak frequency to stability for sensible heat and momentum cospectra are shown in figure 7.3. At very high stabilities (>2) the spectral peak frequencies of sensible heat flux continued to increase while those for momentum cospectra exhibited a decreasing trend, thus accounting for the empirical response to stability as shown in figures 7.1 and 7.2.

7.5.2 Calculated cospectra

While the cospectral models may be used to represent spectra under the conditions for which they were developed, it was considered necessary to create cospectral shapes that applied more appropriately to Griffin. To address possible differences in cospectra, either run-specific or long-term averaged cospectra were employed in determining frequency response flux correction. Calculations employed in the determination of cospectra and of the averaging process are described below.

Cospectra were calculated (EdiRe) for each 30-minute data set using smoothed and spliced fast Fourier transforms (Kaimal & Finnigan 1994). To obtain high frequency spectra, run data were divided into \( n \) sequential segments, where \( n-1 \) is the number of complete 512-point data segments that can be obtained from the data set. These data segments were detrended using a linear least squares fit and tapered using a Hamming
window (Kaimal & Kristensen 1991). The final segment was zero buffered if insufficient data points were available to fill the segment. The FFT was applied to each segment and the transformed results were averaged prior to logarithmic bin averaging of the spectral results. A low frequency spectrum for the same run was obtained by averaging the data by \( n \) sequential data points to obtain a single set of 512 averaged data points. This data segment was detrended, tapered, transformed, and logarithmically bin averaged using the same methods as were used with the individual high frequency spectra. The resulting two spectral curves were then merged and sorted by frequency to obtain a single spectral curve.

![Graph showing cospectra](image)

**Figure 7.4** Example of average sensible heat cospectra. Moore model cospectra are plotted for comparison.

Cospectra of closed path and open path CO\(_2\)/H\(_2\)O sensors were averaged over the period under investigation (July to September of 2000). However, to obtain averaged \( \overline{w'T'} \) cospectral shapes at a finer separation of stability classes, all available cospectra over the period 1997 to 2001 were employed.
Prior to averaging, cospectra frequencies, \( f \), were normalized \( f = n \cdot (z - d) / U \), using canopy height, \( z \), zero plane displacement, \( d \), and run mean wind speed, \( U \). Cospectra were separated into 18 stability classes with the low end of the class bin specified by the values: -3, -2, -1, -0.5, -0.1, -0.05, 0, 0.05, 0.1, 0.2, 0.4, 0.6, 0.8, 1.0, 1.5, 2 and 3. The measured cospectral powers were summed into an array with normalized frequencies corresponding to a spacing of \( \log(0.08) \) between the values of 0.00001 and 831.8. To adapt the measured cospectral powers to the summing array, the cospectral powers were linearly interpolated based on their normalized frequency. Similarly, the covariance obtained from the integrated cospectral powers was summed for corresponding frequencies. Averaged cospectra were obtained as the ratio of the summed cospectral powers and their corresponding summed covariances; this method reduced the effect on cospectral shapes associated with small covariances; examples of the averaged cospectral curves are shown in figure 7.4.

### 7.6 Response function determination

The response functions employed were theoretical, empirical or based on a combination of theoretical and empirical components. Empirical response functions were limited in their application to \( \text{CO}_2 \) and \( \text{H}_2\text{O} \) results because no alternative velocity or temperature sensors were available in the data set from which alternate, more accurate, sensible heat response functions could be obtained.

#### 7.6.1 Modelled response functions

Theoretical sensor response functions employed in the integration method were obtained from Moore (1986) and Massman (2000). The equations describing these transfer functions are given in appendix B. For the analytical method, values of equivalent time constants were obtained following Massman (2000); the defining equations are also given in appendix B. The model response function combinations, and equivalent time constants used in the integration and analytical methods, respectively, are specified in table 7.3.

#### 7.6.2 Empirical response functions

As an alternative to the theoretical frequency response models, estimations of frequency response have been determined empirically using three approaches; cospectral ratios, sensor calibration step change decay, and filtered sensible heat cospectra time constant. Each of these methods is described below.
Table 7.3 Definition of transfer function and equivalent time constant variables. Equations defining the transfer functions are available in Moore (1986) unless otherwise indicated while those defining equivalent time constants are defined in Massman (2000)

<table>
<thead>
<tr>
<th>Transfer function</th>
<th>Equivalent time response</th>
<th>Meaning</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_B$</td>
<td>$\tau_B$</td>
<td>Block averaging high-pass</td>
</tr>
<tr>
<td>$T_{PV}$</td>
<td>$\tau_{PV1}$</td>
<td>Vector path averaging (vertical) used for momentum flux</td>
</tr>
<tr>
<td></td>
<td>$\tau_{PV2}$</td>
<td>Vector path averaging (vertical) used for scalar flux</td>
</tr>
<tr>
<td>$T_{PH}$</td>
<td></td>
<td>Vector path averaging (horizontal) used for momentum flux</td>
</tr>
<tr>
<td>$T_{PS}$</td>
<td>$\tau_{PS}$</td>
<td>Scalar path averaging</td>
</tr>
<tr>
<td>$T_t$</td>
<td>$\tau_t$</td>
<td>Sensor time constant</td>
</tr>
<tr>
<td>$T_S$</td>
<td>$\tau_S$</td>
<td>Sensor separation</td>
</tr>
<tr>
<td>$T_{TA}$</td>
<td>$\tau_{TA}$</td>
<td>Tube attenuation (Massman 1991; Leuning Moncrieff 1990)</td>
</tr>
<tr>
<td>$\tau_{Cyl}$</td>
<td></td>
<td>Scalar cylindrical averaging</td>
</tr>
<tr>
<td>$\tau_{Sph}$</td>
<td></td>
<td>Scalar spherical averaging</td>
</tr>
</tbody>
</table>

Table 7.4 Modelled response function frequency response transform components applied to the specified covariances. Equivalent time constants were evaluated at a wind speed of 1 m s$^{-1}$

<table>
<thead>
<tr>
<th>Covariance</th>
<th>Integrated</th>
<th>Analytical</th>
<th>Equivalent time constant</th>
</tr>
</thead>
<tbody>
<tr>
<td>$u'u'$</td>
<td>$T_B T_{PV}^2$</td>
<td>$\tau_B \tau_{PV1} \tau_{PH}$</td>
<td>0.0422 s</td>
</tr>
<tr>
<td>$w'T_z'$</td>
<td>$T_B T_{PV} T_{PS}$</td>
<td>$\tau_B \tau_{PV2} \tau_{PS}$</td>
<td>0.0415 s</td>
</tr>
<tr>
<td>$w'c'$ closed path</td>
<td>$T_B T_{PV} T_{PS} T_t T_S T_{TA}$</td>
<td>$\tau_B \tau_{PV2} \tau_{Sph} \tau_t \tau_S \tau_{TA}$</td>
<td>0.386 s</td>
</tr>
<tr>
<td>$w'q'$ closed path</td>
<td>$T_B T_{PV} T_{PS} T_t T_S T_{TA}$</td>
<td>$\tau_B \tau_{PV2} \tau_{Sph} \tau_t \tau_S \tau_{TA}$</td>
<td>0.244 s</td>
</tr>
<tr>
<td>$w'c'$, $w'q'$ open path</td>
<td>$T_B T_{PV} T_{PS} T_t T_S$</td>
<td>$\tau_B \tau_{PV2} \tau_{Cyl} \tau_t \tau_S$</td>
<td>0.296 s</td>
</tr>
</tbody>
</table>

### 7.6.2.1 Response time constant determined by filtering sensible heat flux

Following Goulden (1996), a response function appropriate for a sensor, located near to a sonic anemometer measuring sensible heat flux, can be obtained by low-pass filtering the sonic’s temperature signal until the shape of the sensible heat flux cospectra is...
identical to that of the scalar signal being measured. In this approach a 0.05 second recursive filter was applied to the temperature signal and sensible heat cospectra calculated using the filtered temperature signal and vertical velocity. A spectral ratio was then calculated with the scalar cospectra in the numerator and the sensible heat cospectra in the denominator. Prior to calculating the ratio the cospectra were normalized over the spectral frequency range of 0.002 to 0.05 as it was assumed that the two cospectra behaved similarly in the low frequency regions. This cospectral ratio was then averaged over the frequency range of 0.1 to 10. If this average was greater than 1 then the low pass filter time constant was increased (15% for CO$_2$ and 17% for H$_2$O) and the process was repeated until the average over the frequency range of 0.1 to 10 was less than 1. The low pass filter time constant coinciding with an average of less than one was considered to be an equivalent system time constant for the scalar sensor. If an average of less than one had not been achieved within 25 iterations, the maximum time constant was assumed appropriate (1.43 s for CO$_2$ and 2.16 s for H$_2$O). This time constant was then applied to the theoretical transfer function approach as a sensor time constant while other transfer function parameters were set to those appropriate for the sensible heat flux measurements. In general, the observed time constants (figure 7.5) were similar to those obtained from modelled transfer functions.

![Figure 7.5 Distribution of time constants obtained from the filter applied to sensible heat flux cospectra to match the shape of IRGA cospectra.](image)

This method was subject to errors caused by high frequency noise in the CO$_2$ and H$_2$O cospectra. Such high frequency energy resulted in premature termination of the filtering process and equivalent time constants that were too small. Conversely, for some runs the time constant failed to obtain a value before the limit of 25 iterations. This may have
resulted from difference in cospectral shapes between the sensible heat flux cospectra and that of the other scalars.

### 7.6.2.2 Response function time constant determined from calibration decay

Following a suggestion by Goulden (Ameriflux meeting May 2000), the sensor time constant of the closed path CO₂ and H₂O signals was determined from the signal step change associated with sensor calibration. The step change associated with the initiation of calibration was inappropriate for our system because the source of calibration gas was not located at the intake tube inlet. The calibration cycle consisted of two repetitions of two gases (zero and span gas) with each gas flowing through the system for two minutes. The first minute allowed the system to be purged while sensor readings were recorded during the second minute. The flow rate of calibration gas was such that the excess gas was sufficient to also purge the inlet tube to the intake. Therefore, the resulting step change was sufficient to characterise the response of both the sensor and inlet tube system. Therefore, the step change associated with the termination of calibration was employed. The time constant was determined from the time required to achieve a 63.2% change in the signal associated with a step change. To more accurately identify the step change, the signal was averaged for periods of 10 s prior to and 20 s following the step change. To allow for signal noise, a sensitivity corresponding to +/- 5 standard deviations of the pre and post step change signal variability was imposed. Because calibrations were intermittent (every 48 hrs), individual step change time constants were not employed in the correction. Instead, a distribution of the resulting time constants was plotted and the peak of the lognormal distribution was taken as the representative time constant for the system, (figure 7.6).

The time constant obtained from the signal step change method were applied to the theoretical transfer function approach as a sensor time constant while other transfer function parameters were set to those appropriate for the sensible heat flux measurements. Because this method does not take into account frequency loss caused by path separation, an additional theoretical term was added to the frequency response to take the sensor separation into account.
Response function time constant determined from cospectral ratios

In this approach, a response function was calculated from the ratio of scalar flux cospectra to that of sensible heat flux cospectra \( \frac{C_{wx}}{C_{WT}} \). Two variations of this approach were tried; the ratio of average cospectra and the average of cospectral ratios. These two approaches should produce the same response function, because cospectra were grouped by stability classes before calculations were performed. Similar calculations using ratios of scalar and temperature spectra were attempted but proved problematic because of the greater noise levels present in the scalar spectra. In the calculation of average ratios, the cospectral ratios at each frequency were limited to values between 0 and 2 prior to averaging. For both variations, an equivalent time response was obtained by matching the cospectral ratios to the response curve of a low pass recursive filter. In this approach, it was common to observe a frequency response drop off at low frequencies for both CO\(_2\) and H\(_2\)O. The cause of the low frequency drop off was not investigated and only the time constant associated with the low pass filter was employed in the frequency response correction. Figure 7.7 shows a typical cospectral ratio and the fitted recursive filter response curves.

Four stability classes were used for the estimation of equivalent sensor time constants from cospectral ratios. For the ratio of average cospectra only the 5-year averaged
cospectral shapes were employed. For the average of cospectral ratios, the effect of using one year versus one-month averages was tested. For the open path sensors only one month’s data was available and only the average of cospectral ratios was calculated. The results of these comparisons are shown in Figure 7.8; the calibration step change time response is presented in the figure for comparison. The results suggest higher values of time constant under near neutral conditions for the closed path H₂O sensor. There also appears to be a slight decrease in time constant with increasing stability for both CO₂ sensors and for the open path H₂O sensor. It was decided to use two values of sensor equivalent time constant based on these methods, one for unstable and one for stable conditions. The values selected are presented in Table 7.5.

![Figure 7.7 Example of ratio of averaged CO₂ flux cospectra to sensible heat flux cospectra (points). Solid curve represents least squares fitted combined high and low pass recursive filters.](image)

**Table 7.5** Equivalent sensor time constants employed in frequency response corrections. Two variations of cospectral ratios and the calibration step change method were used to obtain time constants for the closed path system and the average of cospectral ratios was used for the open path sensors. Single values were chosen for stable and unstable conditions.

<table>
<thead>
<tr>
<th>Method</th>
<th>CO₂</th>
<th></th>
<th>H₂O</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Unstable</td>
<td>Stable</td>
<td>Unstable</td>
<td>Stable</td>
</tr>
<tr>
<td>Closed path, average of cospectral ratios</td>
<td>0.18</td>
<td>0.11</td>
<td>1.05</td>
<td>1.99</td>
</tr>
<tr>
<td>Closed path, ratio of cospectral averages</td>
<td>0.18</td>
<td>0.23</td>
<td>0.59</td>
<td>1.38</td>
</tr>
<tr>
<td>Closed path, calibration response</td>
<td>0.12</td>
<td>0.12</td>
<td>3.91</td>
<td>3.91</td>
</tr>
<tr>
<td>Open path, average of cospectral ratios</td>
<td>0.09</td>
<td>0.07</td>
<td>0.10</td>
<td>0.08</td>
</tr>
</tbody>
</table>
As expected for the open path sensor, the time constants are similar for both CO₂ and H₂O under both stable and unstable conditions. For the closed path sensor, the time constant estimations are consistent for CO₂ but exhibit large variability and larger values. The reasons for the larger time constants for water vapour are discussed in chapter 5.

7.7 Comparison results

The effect of different methods of frequency response correction was determined by examining both the resulting correction factors as well as the effect of the correction factors on associated flux values. The results are presented by flux type.

The spectral band pass ratio methods (eg. L, M, and N) exhibited large amounts of scatter in correction factors and correspondingly large scatter in the net effects for different fluxes. It is believed that large variability of cospectral powers in the region of band agreement, and the lack of similarity of different fluxes in that region, was the cause of the large observed scatter for these methods. It was therefore decided that discussion of the effect of these methods would not be included in the comparison of methods. Although the band pass ratio method proved impractical for long-term application in this
analysis, it may still be considered applicable for situations under which more intensive analysis is possible.

### 7.7.1 Sensible heat flux

In calculation of the frequency response correction of sensible heat flux only three distinct methods were available, methods A, B, and J. Other methods depend upon sensible heat flux cospectra as a reference, which precluded their use as a correction method for sensible heat flux.

Sensible heat flux corrections behaved similarly with respect to stability, for the three methods employed, exhibiting larger correction effects under unstable conditions, figure 7.9. Comparison between methods indicates the correction effect of method A to be approximately 1% less than that of methods B and J under both unstable and stable conditions.

![Figure 7.9 Comparison of net correction effect on sensible heat fluxes measured under unstable, gray bars, and stable conditions, black bars. The correction effect is the ratio of the average corrected flux to the average uncorrected flux.](image1)

Example plots of model cospectra and transfer functions, (figure 7.10), help visualization of how high and low frequency attenuation of cospectra behave under different wind speeds and stabilities. It can be seen that high frequency losses under unstable conditions should be larger for methods B and J because of broader cospectral shapes of the fitted Massman curves. Because the spectral shift with wind speed of the high frequency transfer function cutoff is identical to that experienced by cospectral peak frequency, this attenuation effect will remain constant with changes in wind speed. However, the low
frequency transfer function cutoff frequency is not a function of wind speed so that the low frequency cospectral attenuation will increase as wind speeds decrease.

Figure 7.10 Relationship of cospectral models to frequency response attenuation curves for sensible heat flux under two wind speed conditions. Cospectral models are Moore (solid lines) and fitted Massman (dashed lines) for both unstable (heavy curves) and stable \((z/L = 1.0)\) conditions. The transfer functions (top two curves in both graphs) are based on Moore's model (solid line) and Massman's equivalent time constant (dashed line).

Figure 7.11 Sensible heat flux correction factors obtained from methods A, B, and J plotted in relation to wind speed, Filled circles are for unstable conditions and open circles are for stable conditions.
Under stable conditions, similar attenuation characteristics are observed, with the added complications of spectral shifts of cospectral peak frequency as a function of stability and changes in cospectral shapes. From figure 7.10 it is observed that the greater cospectral shift in the cospectral model used in method A will act to reduce the effect of low frequency cospectral attenuation while making it more susceptible to high frequency attenuation losses.

Figure 7.12 Relationship of values to stability. The top panel contains the frequency distribution of data (n=1393, approximately 1 month of data) and average wind speed corresponding to each stability class. The second panel shows the corresponding average, uncorrected sensible heat flux. The third panel contains the cumulative correction effect in units of W m\(^{-2}\) (sum of corrected minus uncorrected for corresponding stability class) for each of the three methods investigated (A - filled circle, J - open circle, B - crosses). The bottom panel shows the corresponding average correction factors for each method. The x-axis is a log representation of the Monin-Obukhov stability, which has been truncated at stability magnitudes of less than 10\(^{-3}\).
The relationships to wind speed and stability suggested by figure 7.10 are shown more explicitly in a plot of correction factors against wind speed, figure 7.11. This figure demonstrates that for unstable conditions it is only wind speed that controls the magnitude of signal attenuation. Under stable conditions, the correction also increases with decreasing wind speed but also varies with stability – which appears as increased scatter for the stable correction factors.

Comparison between methods, (figure 7.11), shows, in agreement with figure 7.10, that under unstable conditions the increase in the correction factor of method B is smaller than that of method A with decreases in wind speed, as can be expected from the lower cospectral energy at low frequencies in the method B’s cospectral model. Under stable conditions the correction of method B is greater than that of method A, as can be expected from the smaller cospectral shift of the cospectral model employed by method B.

While the cospectral models implicit in method J should be similar to that used in method B it is apparent, in figure 7.11, that method J has greater correction factors for both unstable and stable conditions. Based on the larger unstable correction values and wider range of stable corrections it is suggested that the analytical solution results in effectively increased low frequency cospectral energy.

To understand the differences in the observed net effect on fluxes, a frequency distribution of data, average wind speeds, fluxes, and correction factors, as well as the cumulative correction effect have been plotted as a function of stability in figure 7.12. It is obvious that large net correction effects result from a combination of large flux magnitudes and a larger frequency of occurrence, and is not necessarily linked to large correction factors. The larger differences in correction factors at extreme stable and unstable conditions, as well as the greater similarity between correction factors at near neutral conditions have insignificant effects because of both the low frequency of data corresponding to these conditions as well as the correspondingly small magnitudes of sensible heat flux.
7.7.2 Momentum flux

For momentum flux, as with sensible heat flux, no alternate methods of estimating momentum reference cospectra were available. Therefore, as with sensible heat flux, only three distinct correction methods were available (A, B, and J). As with sensible heat flux corrections, the average correction effect was small (<4%) with corrections under unstable conditions 1 to 2% higher than under stable conditions for all methods, (figure 7.13). However, unlike sensible heat flux, methods A and B were more similar in their net effect while method J had a 2% greater effect under unstable conditions and 1% greater under stable conditions.

The similarity of methods A and B under unstable conditions is apparent from comparison of cospectral models and transfer functions, figure 7.14, as the cospectral shapes are very similar at both low and high frequency ends of the cospectra. The greater correction effects of method J, again suggest that its implicit cospectral shape has more low frequency cospectral energy than suggested by the cospectral model of method B (from which method J obtained it's cospectral parameters).

Under stable conditions the larger spectral shift of method A is clearly discernable as a flatter response to wind speed (figure 7.15), whereas both methods B and J exhibit the increase in correction factor at low wind speeds characteristic of low frequency attenuation.
Figure 7.14 Relationship of cospectral models to frequency response attenuation curves for sensible heat flux under two wind speed conditions. See figure 7.9 for description.

Figure 7.15 Momentum flux correction factors obtained from methods A, B, and J plotted in relation to wind speed. Filled circles are for unstable conditions and open circles are for stable conditions.

Examination of the data frequency distribution, average values and cumulative values, figure 7.16, show that, similar to sensible heat flux, the corrections to momentum flux are largest as a result of a combination of data distribution and flux magnitudes. What is different for the correction of momentum is that flux magnitudes are largest for near neutral conditions, in contrast to sensible heat fluxes that were larger for moderate
stabilities. This combination has resulted in a shift of the greatest momentum correction effect to periods when the magnitude of stability is about an order of magnitude smaller than observed for sensible heat flux. The large correction effects observed under extreme stabilities become insignificant because of the negligible momentum flux under these conditions.

Figure 7.16 Momentum frequency response correction relationships with stability. See figure 7.12 for description.

Also apparent are the larger correction effects of method J compared to methods A and B. This difference increases at extreme stabilities, as a result of decreasing wind speeds, implying that the amount of low frequency flux correction of method J is greater than suggested by the fitted cospectral model and transfer functions shown in figure 7.14.
7.7.3 Carbon dioxide flux

For CO₂ fluxes, all correction factor estimation methods are compared; however, because most methods are simply variations on the numerical integration method A, they will be compared to that method for the purpose of analysis. Comparisons of both open and closed path sensors are made.

As expected, the frequency response corrections for carbon dioxide flux are greater in magnitude than those observed for either sensible heat or momentum flux (figure 7.17). The average net correction applied to the closed path CO₂ sensor was 23% ± 9% and 28% ± 18% for unstable and stable conditions, respectively. For the open path sensor the corresponding values were 15% ± 9% and 15% ± 19%. These results indicate an approximately 10% larger correction for the closed path sensor as compared to the open path sensor under both stable and unstable conditions. However, the variability observed between methods is large compared to the average correction effect, suggesting a great potential for erroneous adjustment of fluxes.

![Figure 7.17 Frequency response correction effect for carbon dioxide flux measurements. See figure 7.9 for description.](image)

In contrast to the sensible heat and momentum fluxes, corrections to carbon dioxide fluxes are generally larger under stable conditions as compared to unstable conditions. The greater correction for the closed path sensor arises from its high frequency attenuation, which is not a function of wind speed and does not shift to higher natural frequencies under higher wind speeds, (figure 7.19). The static nature of the high frequency attenuation in relation to wind speed is observed more clearly when correction factors are plotted in relation to wind speed, figure 7.19. This effect is observed as a
correction factor monotonically increasing with wind speed. It is interesting to note that
the methods that show this effect are those which employ the model transfer functions or
filter determined time constant as an equivalent transfer function, all of which had time
constants of more than 0.3 seconds (table 7.4 and figure 7.5). The only method that
employed model transfer functions and did not exhibit the effect was method K; it is
possible that unaccounted for attenuation in the open path cospectra may have affected
the correction factor determined with method K.

![Figure 7.18 Relationship of cospectral models to frequency response attenuation curves for carbon dioxide flux measured with a closed path sensor under two wind speed conditions. See figure 7.10 for description.](image)

Other methods employing empirically determined transfer functions (methods E, F, G)
also showed small correction factors associated with high frequency attenuation,
consistent with their smaller equivalent time constants of between 0.05 and 0.2 s (table
7.5).

For the open path sensors, (figure 7.20) it is observed that methods with modelled
transfer functions have high frequency attenuation that do not increase with wind speed,
while the methods using empirically determined transfer functions do – with the
exception of method E which continued to have small equivalent time constants.
Figure 7.19 Closed path sensor carbon dioxide flux correction in relation to wind speed. Filled circles are for unstable conditions and open circles are for stable conditions. Methods A through N are defined in table 7.1.
Examination of the effects of frequency response correction in relation to stability, figure 7.21, shows that for unstable conditions the effect of correction is associated primarily with the frequency distribution of the data. This is so because both the average flux magnitude and correction effect magnitude are relatively constant with respect to stability. Under stable conditions a similar relationship is observed – except that under extreme stable conditions the effect of correction is larger than would be implied by the frequency distribution, as a result of the greatly increased correction factors. In this figure it is also apparent that sizable and consistent differences between correction methods are persistent for most stabilities. It is only under extreme unstable conditions for the open path sensor that the increasing trend in correction factors related to low frequency attenuation is observed.
7.7.4 Latent heat flux

The frequency response correction effect on latent heat flux is very similar to that observed for carbon dioxide flux for the open path sensor. This similarity suggests good cospectral similarity in the shapes of cospectra for both latent heat and CO₂ fluxes.
For the closed path sensor the correction effects of all methods are similar for those that use modelled transfer functions (A, B, C, D, J, K). Methods employing empirically determined time constants (E, F, G, H, and I) all exhibit larger correction factors. This is expected because the modelled transfer functions do not capture the effect of stronger tube attenuation for water vapour that are captured by the empirical methods, as discussed in chapter 5. That the effect of greater water vapour attenuation is a high frequency effect is emphasised by the strong relationship to wind speed observed in figure 7.23. As stated above, the correction effects for the open path latent heat flux are nearly identical to those observed for open path CO2 flux, this close, expected relationship is repeated again in the similar patterns of correction factors in relationship to wind speed, as shown in figure 7.24.

The relationship of the open and closed path latent heat flux correction factors to stability (for methods A, B, and J), figure 7.25, are likewise identical to those observed for carbon dioxide flux. However, slight differences in the correction effect are observed because of the different distribution of latent heat flux with respect to stability. The largest average latent heat fluxes are observed under moderately unstable conditions, while almost no latent heat flux exists under stable conditions. This pattern means that the very large corrections observed under extreme stable conditions have negligible effect on the net correction effect.
Figure 7.23 Closed path sensor latent heat flux correction in relation to wind speed. Filled circles are for unstable conditions and open circles are for stable conditions. Methods A through N are defined in table 7.1.
Figure 7.24 open path sensor latent heat flux correction factors obtained from methods A, B, and J plotted in relation to wind speed. Filled circles are for unstable conditions and open circles are for stable conditions.
Figure 7.25 Same as figure 7.8 but for latent heat flux. Additional panels have been added for the correction effect on open path $\lambda E$ and open path correction factor for $\lambda E$. 
7.8 Summary

Considerable variability can exist in final flux values as a result of frequency response correction. The problem appears to be primarily related to our inability to define a consistent (and probably appropriate) transfer function for signal attenuation. This problem is compounded by deviations in the shape of true atmospheric cospectra from the model cospectral shapes currently employed to represent ‘true’ cospectra. Despite potential inaccuracies, frequency response correction errors will generally be insignificant in the correction of sensible heat and momentum fluxes for most measurement systems. The results of analysis in this chapter suggest that differences between sensible heat - or momentum - corrected using different frequency response methods for will generally be less than 2%. For the fluxes of carbon dioxide and latent heat (and probably other similarly measured scalars) the effect of different frequency response correction methods can be much larger, with differences in final corrected values being as large as 10% or greater. Improvements in transfer functions and cospectral models are both required before accurate correction of more severely attenuated scalar fluxes can be obtained. It is acknowledged that improvements in experimental design may alleviate many of the problems associated with relying on purely modelled corrections. It is also, however, important to keep in mind that the observed problems will continue to exist even in an experiment designed to reduce their effects. It is suggested that other, carefully planned, experiments be carried out to better identify the differences between correction methods and when they can and cannot be applied successfully to a data set.
8 Ecosystem Exchange

8.1 Introduction

This chapter discusses the measurements of forest-atmosphere exchange at Griffin over the period 1997 to 2001. Terms in the energy balance and the fluxes of momentum and carbon dioxide are described. Patterns of these exchanges are explained in relation to appropriate variables and controlling factors. The storage of mass and energy is presented first, followed by the fluxes of momentum, sensible heat, \( H \), latent heat, \( \lambda E \), and carbon dioxide, \( F_c \), and the energy budget.

8.2 Storage

Storage terms were described in Chapter 5 and the associated measurements are presented in Chapter 5 and Appendix A. In this section, sensible heat storage will be covered in most detail, followed by storage terms for \( \lambda E \) and \( F_c \). A model of storage is developed for the purpose of representing storage on days when there was an incomplete series of associated measurements.

Heat storage terms, averaged by percent of the diurnal/nocturnal period, \( \Pi_d \), are presented in figure 8.1. The temporal changes in net radiation and wind speed have been plotted for comparison because of their importance in determining surface exchange. The Griffin Forest average midday sensible heat storage, shown in figure 8.1, was slightly less than 20 W m\(^{-2}\) (about 10% of the summer, midday sensible heat flux) (McCaughey 1985; McCaughey & Saxton 1988; Saxton & McCaughey 1988). The primary contributors to sensible heat storage, (described as percent of total storage), are warming of water in the soil and biomass, 42%, and photosynthesis, 34%. Storage of heat in the air column, 22%, dry biomass, 16%, and dry soil, 7%, constitute smaller contributions to the total sensible heat storage term. These results are similar to the findings of McCaughey et al. (1997, 1988) who observed heat storage to peak near midday and to account for 10% of midday heat flux.

The diurnal pattern of sensible heat storage is approximately sinusoidal, with a peak slightly before midday. Nocturnal sensible heat storage reaches a maximum
magnitude just after sunset and decays exponentially throughout the night. Comparison of the diurnal patterns of sensible heat storage with the patterns of change of net radiation and wind speed reveals that only the peaks of air and biomass heat storage coincide with the peaks in $\Delta R_n$ and $\Delta U$. The storage of heat by photosynthesis peaks near midday because it is controlled primarily by the magnitude of radiation and not by its temporal change. Other heat storage terms peak later in the day, indicating an effect of an increased component storage capacity.

![Figure 8-1 Average sensible heat storage and components as a function of the percent of the diurnal/nocturnal period, bottom panel. The top panel contains the average run-to-run change of $R_n$ and wind speed. The percent of the diurnal period varies between 0 and 100 and represents the position of a run within the diurnal period. The percent of the nocturnal period varies between 100 and 200 and represents the position of a run within the nocturnal period.](image)

The proportions of sensible heat storage components at night are similar to those observed during the day, although of opposite sign and with a different pattern of change. Without the driving force of radiation to determine the heat storage pattern during the night, heat is lost from the ecosystem following what appears to be an exponential decay rate until the initiation of the diurnal cycle the following morning.
The storage of water vapour and CO$_2$ in the air column (figure 8.2) follows a pattern similar to that observed for the storage of sensible heat in the air column. This pattern is characterized by a strong peak at the start of the diurnal period that decreases in a nearly linear manner to reach a peak of opposite sign at the start of the nocturnal period (Garten et al. 1999; MacPherson 1998). The change in storage of water vapour and CO$_2$ during the nocturnal period is small, but behaves in a manner similar to that observed for sensible heat.

![Figure 8-2: Diel patterns of storage components of $H$, $\lambda E$ and $F_c$ for different ranges of wind speed (left three panels) and net radiation (right three panels).](image)

The magnitude of the latent heat storage term is similar to that observed for sensible heat in the air column, with average values ranging up to about 4 W m$^{-2}$. This amounts to about 3% of the average midday flux during the summer, a smaller proportion than that for the storage of sensible heat. Average carbon dioxide storage reached 3 $\mu$mol m$^{-2}$ s$^{-1}$ or approximately 5% of the summer average midday flux magnitude, although under low wind speed conditions this storage was larger than 10% of the summer midday peak fluxes.
Storage terms were modelled because of poor data coverage, (measurements were occasionally sacrificed in order to maintain sufficient power for the flux measurement systems). Modelling also allowed comparison of point measurements with modelled values that are more representative of spatially averaged canopy storage.

The diel pattern of storage was hypothesized as consisting of two distinct processes. A morning change in storage was assumed to be associated with improved canopy-atmosphere coupling as a result of boundary layer development. Coupling will have provided a sink for mass and energy accumulated in the canopy over night. Similarly, the change in storage beginning at sunset was assumed to result from canopy-atmosphere decoupling. Decoupling allowed within canopy build-up of mass and energy, which was largest immediately after sunset when the energy content of the canopy was still quite large. It was further assumed that each of these processes behaved in a manner that could be represented by a Gaussian distribution. These assumptions lead to an empirical model of the form:

\[
\Delta F = S_0 + a_d \cdot e^{-\frac{1}{2} \left( \frac{|t - \hat{t}_d|}{b_d} \right)^2} + a_n \cdot e^{-\frac{1}{2} \left( \frac{|t - \hat{t}_n|}{b_n} \right)^2}
\]

(8.1)

where the coefficients \(a_d, a_n\) represent the peak magnitude of storage, \(b_d, b_n\) represent the width of the distribution, \(c_d, c_n\) the sharpness of the distribution peak, and \(\hat{t}_d, \hat{t}_n\) represent the time of the storage peak as percent of the diurnal/nocturnal period. The coefficient \(S_0\) represents the net diel storage of energy associated with net photosynthesis. For these coefficients the subscripts \(d\) and \(n\) represent diurnal and nocturnal conditions, respectively.

This model was fitted to the storage distributions obtained by averaging all available storage data using the coefficients given in table 8.1. Figure 8.3 shows that the model provides a good fit to the averaged data (\(R^2 > 0.97\) for all fits).
Figure 8-3 Comparison of average diel curves of storage (filled circles) with the model described by eqn. 8.1 fit to these curves (open circles).

Table 8.1 Static coefficients employed in the storage models for sensible heat, latent heat, and carbon dioxide fluxes.

<table>
<thead>
<tr>
<th>Coefficient</th>
<th>$H$</th>
<th>$\lambda E$</th>
<th>$F_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$b_d$</td>
<td>29.3</td>
<td>6.84</td>
<td>9.92</td>
</tr>
<tr>
<td>$c_d$</td>
<td>2.57</td>
<td>1.36</td>
<td>0.94</td>
</tr>
<tr>
<td>$d_d$</td>
<td>43.4</td>
<td>18.8</td>
<td>15.1</td>
</tr>
<tr>
<td>$b_n$</td>
<td>6.21</td>
<td>2.75</td>
<td>4.73</td>
</tr>
<tr>
<td>$c_n$</td>
<td>0.63</td>
<td>0.49</td>
<td>0.70</td>
</tr>
<tr>
<td>$d_n$</td>
<td>98.8</td>
<td>105.9</td>
<td>100.6</td>
</tr>
</tbody>
</table>

The model given by equation 8.1 and the coefficients given in table 8.1 were assumed to apply for all days. Application of these coefficients allowed the models to integrate to near zero over the diel period and so maintain the criterion of no net storage. In contrast, the coefficients $a_n$, $a_d$, and $S_0$ were assumed to vary in response to conditions that determine the magnitude of canopy storage.

The variables responsible for controlling the magnitude of the storage terms were determined using two approaches. The first approach was to average the storage curves based on the diel range of $R_n$ and $T_a$, and the diel average $U$. The storage peak values were extracted from these averaged curves, (see figure 8.2). The peak values were then stepwise regressed against the averaging conditions to determine which averaging variables were most important in determining storage values and to obtain a multi-linear regression against these variables, equation 8.2.
These regressions provided a moderate level of explanation of the behaviour of storage ($H$, $R^2 = 0.69$; $\Lambda E$, $R^2 = 0.25$; $F_c$, $R^2 = 0.46$). These findings are similar in nature to those of McCaughey and Saxton (McCaughey et al. 1988) though they do not extrapolate their results to the level of modelling of storage terms.

A similar approach was taken to assess the potential controlling effects on $a_d$ of a wider range of variables. The diel ranges of half hour storage values were stepwise regressed against the diel ranges of a larger set of variables ($\ln(U)$, $\Delta U$, $\Delta R_n$, $\Delta T_a$, $\Delta T_s$, $\Delta R_g$, $APcp$, $\Delta R$, $\Delta g_s$, $\Delta D$, $\Delta e$, $\Delta RH$). By this expanded approach, the controlling variables identified for $H$ ($\Delta R_n$, $\Delta T_a$, $\Delta R$) for $\Lambda E$ ($\Delta R$, $\Delta g_s$) and for $F_c$ ($\Delta T_a$, $\Delta T_s$, $\Delta R$, $\ln(U)$), although not identical, were similar to the initial determinations. Because of the larger scatter using the second method ($R^2$ generally less than 0.05) it was decided to employ the first method for determination of $a_d$ given by equation 8.2.

In order to maintain the criterion of zero net storage, the coefficients $a_n$ and $S_0$ were parameterised using the characteristics of the curves presented in figure 8.3, i.e.,

$$
S_0 = \begin{cases}
-0.148 \cdot a_d & H \\
0 & F_c \text{ and } \Lambda E
\end{cases}
$$

$$
a_n = \begin{cases}
-0.59 \cdot a_d & H \text{ and } \Lambda E \\
-1.13 \cdot a_d & F_c
\end{cases}
$$

This approach assumes an implicit control of nocturnal storage by the conditions controlling diurnal storage. This is likely to be a weak assumption, and simple methods of circumventing this weakness should be investigated.

An example of the diel behaviour of all three models of storage is shown in figure 8.4. Some general characteristics of the models can be deduced from this figure. Discontinuity of the model at the transition between days, is most noticeable in the $H$
storage component values. This discontinuity is an inevitable result of trying to
maintain the criterion of zero net storage over the diel period, and the
parameterisation of the curves using averaged values from the diel period to which
they were applied. Also notable is the better correlation of measured and modelled
storage for $H$ than for either $F_c$ or $\lambda E$, with the correlation for $\lambda E$ being the poorest.
This was generally true for all days examined. The large variability in the air storage
component for $F_c$ and $\lambda E$ causes this poor correlation, however, comparison of the 4.5
hour running mean of the measured values shows a better correlation; with the
modelled values – especially for $F_c$ storage – lending support to the model employed.

![Figure 8-4](image)

Figure 8-4 Comparison of measured (filled circles) and modelled (solid
line) storage terms for sensible heat (top panel), latent heat (center panel),
and carbon dioxide flux (bottom panel). A 4.5 hour running mean of the
measured storage (dashed line) is also presented in each panel.

Ideally, there should be no net effect of the storage terms over long time periods, e.g.
annually. However, because of inaccuracies in the formulation of the model some
residual net storage is associated with the models. For $F_c$ this amounted to an
inaccuracy in the final cumulative flux value of approximately 0.5%; the proportional
error in $\lambda E$ should be smaller because of the uni-directional character of this flux. The error could not be evaluated for $H$ because of the energy stored as a result of net photosynthesis.

Although a reasonable representation of sensible heat storage was obtained, the more random nature of storage of $H_2O$ and $CO_2$ in the air column made accurate representation of these point measurements difficult to achieve. However, without further knowledge of the behaviour of spatially averaged canopy storage to verify the model, it was decided to retain measured storage values for days when a complete set of storage values were available and to employ the modelled values on days for which any half hours were missing storage values. Analysis of cumulative fluxes adjusted using either completely or partially modelled storage values revealed that the resultant bias on the fluxes was similar for both approaches, i.e., $\pm 0.5\%$.

8.3 Momentum Flux

Momentum flux describes the energy lost from the kinetic energy of the atmosphere through friction, as a result of its contact with the earth’s surface. This energy transfer may be expressed as friction velocity, $u_*$, which is the square root of the covariance describing momentum flux. Momentum flux will be represented by friction velocity in this section.

8.3.1 Correction effects

Figure 8.5 shows a comparison of friction velocities obtained at run-time from Edisol with final corrected values, table 8.2. A small net effect is apparent and caused by differing coordinate rotations, angle of attack correction (see Appendix H) and frequency response correction. The data in figure 8.5 suggests that correction, or the lack of correction, of friction velocity values can introduce considerable variability. As a measure of this variability, the RMS error of the regression is only about 3% of the entire range of observed values. This variability appears to decrease with increasing flux values.
Table 8.2 Linear regression of corrected against uncorrected friction velocity, n > 10,000 for all years.

<table>
<thead>
<tr>
<th>Year</th>
<th>Intercept</th>
<th>Slope</th>
<th>RMSE</th>
<th>R²</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>0.02</td>
<td>0.99</td>
<td>0.037</td>
<td>0.98</td>
</tr>
<tr>
<td>1998</td>
<td>0.02</td>
<td>0.99</td>
<td>0.040</td>
<td>0.98</td>
</tr>
<tr>
<td>1999</td>
<td>0.02</td>
<td>1.00</td>
<td>0.039</td>
<td>0.98</td>
</tr>
<tr>
<td>2000</td>
<td>0.02</td>
<td>1.00</td>
<td>0.036</td>
<td>0.98</td>
</tr>
<tr>
<td>2001</td>
<td>0.02</td>
<td>1.01</td>
<td>0.035</td>
<td>0.98</td>
</tr>
</tbody>
</table>

Figure 8.5 Correction effect on friction velocity for the year 1998. Uncorrected values are those obtained directly from the Edisol data collection software while corrected values are the results of the procedures described in Chapter 5. The solid line is the linear regression for the 1998 data.

8.3.2 Temporal variations in $u_*$

Figure 8.6 shows that both inter- and intra-annual variation of friction velocity follow a pattern similar to that observed for wind speed, (see figure 4.55), and is consistent with the annual pattern of regional, geostrophic winds (Palutikof et al. 2004). Despite the close comparison of wind speed and friction velocity, the temporal variations in friction velocity show slightly smaller deviations from the five-year mean ($CV = 0.15$) than does mean wind speed ($CV = 0.17$), although not during late autumn and winter, (see figure 8.7).
Figure 8-6 Monthly mean values of friction velocity for each year (open circles). Closed circles represent the monthly average for all five years of data.

Figure 8-7 Comparison of coefficients of variation of mean wind speed (filled circles) and friction velocity (open circles). Values are calculated using all five years data.

The low mean monthly friction velocities, during spring and summer months, shown in figure 8.6, are the result of low nocturnal velocities as shown in figure 8.8. The diel pattern observed in spring/summer is also observed in the months of February,
March and October, with overall larger variation. In contrast, the diel patterns during November, December and January are characterized by lower diurnal $u^*$ with similar nocturnal values.

![Figure 8-8 Diel patterns of friction velocity. Left panel contains autumn and winter monthly average curves while spring and summer curves are in the right panel. Error bars represent ± one standard error, n > 400.](image)

The low nocturnal $u^*$ values during summer probably result from a combination of the short duration of the night and the time required for breakdown of the boundary layer. The energy that would be transferred to the surface as friction velocity at night is instead expended accelerating the decaying atmospheric boundary layer (Stull 2000), resulting in reduced near surface wind and friction velocities. Similarly, the relatively low diel range of $u^*$ during November, December and January may be related to poor boundary layer development during the short day lengths observed during these months. Under this hypothesis, the diel range of observed friction velocities should be related to the amount of boundary layer growth, which may be determined from the amount of energy received at the surface (Seibert et al. 2000; Stull 1988).

Analysis of the diel range of $u^*$ for monthly averaged values indicates that the diel range is indeed larger in summer than in winter, so that the diel range of $u^*$ more closely follows the values of $R_0$ than it does those of wind speed or $u^*$.

The date/time plot of friction velocity, see figure 8.9, shows a combined view of both the annual and diel variations and clearly shows the unusually low $u^*$ values that occurred during the winter of 2000-2001, consistent with similar patterns of wind speed. Low friction velocities characterized the entire diel cycle during this period,
coincident with below average temperatures, above average humidity and average radiation, and may indicate different global circulation patterns over this period. A negative index of the North Atlantic oscillation suggests the unusual conditions for this period are related to global circulation (Bojariu & Gimeno 2003; Osborn et al. 1999) (recent data obtained from http://www.cru.uea.ac.uk/).

Figure 8-9 Date/time contour plot of half hour values of friction velocity, \( u_* \), over five years.

### 8.3.3 Environmental relations

#### 8.3.3.1 Turbulent intensity and stability

Monin-Obukhov similarity theory suggests that the ratio of the standard deviation of a signal to its appropriate scaling parameter, \( i.e. \), the turbulent intensity, \( \sigma_u/x_* \), should be constant under neutral conditions and vary as a function of stability (Haugen et al. 1971; Merry & Panofsky 1976), although the user must be wary of auto-covariance caused by inclusion of \( u_* \) in both turbulent intensity and stability (Hicks 1981). Nevertheless, the variation in vertical velocity turbulent intensity with changes in Monin-Obukhov stability, figure 8.10, agree with the models of Panofsky et al. (1977), and with data from other experiments (Hicks 1985; McBean 1971; Schotanus et al. 1983; Smedman 1988). The turbulent intensities of the vertical velocity showed no significant temporal variation as a result of growth of the forest over five years (not shown).
Turbulent intensity functions of horizontal wind components are expected to scale with the depth of the boundary layer, \( z_i \), or a combination of \( z_i \) and measurement height (Panofsky et al. 1977; Van Den Hurk & De Bruin 1995) because of the effects of larger scale atmospheric motions on variability of horizontal velocity. However, because boundary layer depth was not measured for this experiment, the values in figure 8.10 are presented in relation to values of \( z/L \). The relationship to boundary layer depth is addressed in the following section. Although the values of \( \sigma_{u}/u^* \) and \( \sigma_{v}/u^* \) may not be properly scaled, the magnitude with respect to stability are similar to those observed by other researchers (Smedman 1988; Van Den Hurk et al. 1995; Zhang et al. 2001).

![Figure 8-10 Variance similarity relations for velocity components. The models for \( \sigma_{u}/u^* \) are those of Panofsky et al. (1977) (solid line) and Merry and Panosfsky (1976) (dotted line). Error bars represent ± one standard error, \( n > 100 \).](image)

Figure 8.11 shows turbulent intensities for near neutral conditions by wind direction and indicates a slight wind direction dependence, with higher values for westerly winds. This finding differs from the observed directional dependence of other turbulence-based measures, as will be discussed below. The higher turbulent intensity for westerly winds may be caused by increased variability for winds from regions of increased topographic variability, (see section 4.13, figure 4.62). A similar observation has been made by Al-Jiboori et al. (2001). Further analysis is needed to determine if this relationship was associated with variation in the boundary layer.
Figure 8-11 Variation in turbulent intensity by wind direction for unstable (solid line) and stable (dotted line) near neutral ($\vert z/L \vert < 0.01$) conditions. Error bars represent ± one standard error, $n > 300$.

8.3.3.2 Boundary layer depth

The absence of boundary layer depth information precluded the possibility of comparing turbulent intensities of the horizontal wind components with empirical models (Panofsky et al. 1977; Van Den Hurk et al. 1995). If it is assumed that the models are approximately correct we may invert the model of Panofsky et al. (1977) to obtain estimates of boundary layer depth from the values of turbulent intensity.

$$z_i = -2L \left( \frac{\sigma_u}{u_*} \right)^{-12}$$

The calculation of boundary layer depth was carried out only for unstable conditions and for $u_*$ greater than 0.05 m s$^{-1}$. The resulting values were grouped by percent of day to examine the distribution of the statistics shown in figure 8.12.

The boundary layer depth is observed to grow from a mid-morning minimum of about 500 m to a maximum of about 1500 m shortly after sunset, showing only slight decline over the night. A further breakdown of this analysis by month (not shown) indicated that clear boundary layer growth during the day could only be observed in spring and summer months while the boundary layer depth during autumn and winter appeared quite variable and with no obvious diel pattern. For the data shown in figure 8.12, the range of boundary layer heights is consistent with the observations of boundary layer depth measurements (Hicks 1985; Wilczak et al. 1997). A similar
estimate of \( z_i \) could also be obtained by inverting models of horizontal turbulent intensity which depend upon both \( z_i \) and measurement height.

![Graph](image)

**Figure 8-12** Boundary layer depth (median, filled circles, and quartiles, dashed lines) in relation to percent of day. All unstable data with \( u_z > 0.05 \) m s\(^{-1}\) were incorporated. Diurnal periods generally contained more than 2000 data points while nocturnal periods contained about 500. Data were grouped by percent of day (0 to 100%) and night (100 to 200%).

### 8.3.3.3 Drag coefficient

The drag coefficient parameter, \( C_D \), when evaluated at a standard height, provides a metric for describing the roughness of the underlying surface (Stull 2000):

\[
C_D = \left( \frac{u_z}{U} \right)^2
\]

(8.6).

A more appropriate value of mean \( C_D \) can be obtained by integrating vegetation density dependent values of \( C_D \), which have been adjusted for sheltering effects of adjacent vegetation elements, over the canopy height (Landsberg & Jarvis 1973; Thom 1971). Further adjustments are also required to account for the Reynolds number dependency of kinematic pressure and viscous drag forces in the behaviour of \( C_D \) (Finnigan 2000; Thom 1971). The simplified relation given by equation 8.6 will be used for further analysis.

Drag coefficient values varied with wind direction. Values of \( C_D \) from north-easterly and south-westerly directions were more than 50% larger than those from south-easterly and north-westerly directions, (figure 8.13). This pattern was verified by determining values of \( C_D \) using mean wind speeds from cup anemometers, (to remove
doubt about autocorrelation effects or problems associated with potential errors in mean sonic wind velocity caused by probe design (Grelle & Lindroth 1994)). The larger values observed with the cup anemometer are consistent with the lower installation height of the cup anemometer. The lower values of cup anemometer $C_D$ for westerly wind directions coincide with flow through the tower structure.

![Graph showing variation of drag coefficient with wind direction](image)

**Figure 8-13** Variation of the five year average drag coefficient with wind direction. Values were calculated for $u^*$ greater than 0.5 m s$^{-1}$, and for wind speeds higher than 2 m s$^{-1}$. Closed circles are values calculated from sonic anemometer wind speeds. Error bars represent ± one standard error, $n > 200$.

The observed variation in $C_D$ by wind direction is consistent with the directional row planting of trees in the vicinity and its effect on forest structure. Rows in the vicinity of the flux tower were planted along a 90-270 degree (i.e. ~ E-W) axis. The drag coefficient values were larger for wind flow along the rows, suggesting relatively greater momentum transport and indicating a proportionally larger increase in $z_o$ than in $(z-d)$ for wind flow along rows. Similar phenomena have been observed for forest plantations (van der Tol *et al.* 2003) and vineyards (McInnes *et al.* 2003).

Seasonal average drag coefficients, calculated for $u^*$ values higher than 0.5 m s$^{-1}$ and $U$ higher than 2 m s$^{-1}$, ranged from a minimum of 0.022 ±0.0001 in 1997 to a value of 0.037 ±0.0001 in 2001, figure 8.14 (left panel). This trend relies on the inherent dependence of the drag coefficient on the height of measurement above the zero plane displacement (Stull 2000) reflecting the growth of the forest over the period of the experiment. Initial inspection of the five-year trend suggests a nearly constant increase in $C_D$ with date. However, a sigmoid curve fit to the five-year monthly-
averaged values of $C_D$ suggests that the period of most rapid increase in $C_D$, corresponds to forest growth, which occurs between May and November (McWilliam 1972).

Because $C_D$ has been observed, in figure 8.13, to vary with wind direction, it is possible that the seasonal variation in $C_D$ was caused by seasonal changes in wind direction. To reduce the effect of the observed seasonal variation in wind direction probability, (shown in figure 4.64), $C_D$ values were averaged by 60 degree wind direction categories before calculating monthly average values. The resulting monthly variation of $C_D$ (figure 8.14, right panel) suggests that changes of wind direction were not responsible for the observed seasonal changes in $C_D$. It is concluded that the observed seasonal changes were most likely the result of seasonal growth in height of the canopy.

8.3.3.4 Roughness length

A value of canopy roughness length, $z_0$, may be obtained by inverting the log profile relationship for neutral conditions (Seginer 1974; Thom 1971).

$$z_0 = (z - d) e^{(k U_{*} z)}$$

(8.7)
The presence of the ratio of $U$ to $u^*$ in the exponent of equation 8.7 makes it unsurprising that the patterns of $z_0$ are similar to those of $C_D$, (figure 8.14). The values of $z_0$ were obtained using the values of $d$ obtained from Chapter 4. Using this method, the resulting values of $z_0$, shown in figure 8.15 right panel, vary from an annual average of 0.74 m in 1997 to a value of 1.3 m in 2001, ($z_0/h_c$ of 0.11 and 0.12 respectively), which are slightly smaller than the values obtained from wind profiles in Chapter 4. The observed values of $z_0/h_c$ are similar or slightly larger than values found in the literature (Bottema et al. 1998; De Bruin & Moore 1985; Jarvis 1976; Lo 1977; Thom 1971).

The explicit dependence of $z_0$ upon $d$ in equation 8.7 means that errors in the magnitude of $d$ should impact directly on the magnitude of $z_0$, which could be partially responsible for the directional dependence of $z_0$ seen in figure 8.15 (left panel). However, employing directionally dependent values of $d$ in equation 8.7 resulted in only a twenty percent reduction in the magnitude of the directional dependence of $z_0$. It is therefore believed that much of the observed directional dependence of $z_0$ is related to canopy structure and not a calculation artefact.

![Figure 8-15 Roughness length as a function of wind direction for different years, left panel. Data were limited to neutral conditions ($u^* > 0.5$ and $|z/L| < 0.1$); Error bars represent ± one standard error with $n$ ranging from 22 to 1269 (average $n = 305$). The right panel shows annual average roughness lengths, $n > 3400$.](image)

The increase in $z_0$ over the experimental period suggests an increase in roughness with canopy growth – at least up to the point of canopy closure. Because the Griffin canopy had likely not reached maximum vegetation density, the observed pattern of
increase of $z_0$ agrees with the models of Raupach (1980) and Shaw and Pereira (1982). To test further this hypothesis a data set covering the growth of the forest for many years prior to and following canopy closure would be needed. Nevertheless, these results are consistent with the observations of canopy closure found in Chapter 2 and 4.

8.3.3.5 Zero plane displacement and wind speed profile

In the roughness layer, the zero plane displacement ($d$) may be estimated from two levels of wind velocity measurement if the friction velocity is known and values of $d$, $z_0$, and $u^*$ are assumed identical for both levels of measurement. Following Physick and Garratt (1995) and limiting conditions to neutral stability the relationship may be written as:

$$d = z - \frac{u^*}{k} \left( \frac{z_i - z_f}{U_i - U_f} \right) \phi_M$$  \hspace{1cm} (8.8)

in which the variable $\phi_M$ is a function describing the effect of the canopy on the wind profile in the roughness sublayer, and is a stability-independent function of the height of the surface layer that should vary between 0.5 and 1.0, (Physick and Garratt, 1995). Simultaneous solution of 8.8 using non-linear methods was not possible because of singularities, however using values of $d$ from section 4.12.2 and solving $\phi_M$ resulted in an average value of $\phi_M$ of 0.75.

Alternately, two instances of equation 8.7 for two levels of velocity measurement were assumed to have equivalent values of $z_0$ and were re-arranged to give the following equation for $d$:

$$d = \frac{\tilde{z}_2 - \tilde{z}_1 e^{\left( k \frac{U_i - U_f}{u^*} \phi_M \right)}}{1 - e^{\left( k \frac{U_i - U_f}{u^*} \phi_M \right)}}$$  \hspace{1cm} (8.9)

Where the term $\phi_M$ is the integrated value of $\phi_M$ given by Physick and Garratt, (1995), which assuming neutral conditions can be expressed in terms of $z$, $d$, and the depth of the surface layer, $z^*$, as:
Although the resulting values of \( d \) had much scatter, equation 8.9 was instead used to determine the \( \phi_M \) value required to obtain a known value of \( d \) for the site. This calculation also required the assumption of similar \( z_0 \), and \( u^* \) values for both velocity levels as well as a known value of \( d \), which was obtained from wind speed profiles, in section 4.12.2. It also assumed an average value of \( z^* \) for the neutral conditions for which \( \phi_M \) was evaluated. The results, shown in figure 8.16, suggest that the existing wind speed profile would need to be adjusted by a \( \phi_M \) factor of between 0.4 and 0.6 to obtain the correct value of \( d \). This result is similar to the observations of Physick and Garratt (1995), Raupach et al. (1980), and in agreement with concepts of canopy velocity profiles (De Bruin et al. 1985; Koufang Lo 1990; Molion & Moore 1983).

\[
\phi_M = \int_{z-d}^{z} 1 - 0.5e^{-\frac{0.7(z-d)}{z-d}} \frac{dz}{z-d}
\]  

(8.10)

The observed reduction in the \( \phi_M \) is in contrast to its expected increase with increasing \( d \). This difference may reflect the more theoretical nature of equation 8.10, with the observed decrease indicating a different functional relationship than that given by equation 8.10. Although the instrument placement at a height range of 1.5 to 2.0 canopy heights places the eddy covariance measurements above the inhomogeneous wake effects of individual forest elements (Raupach et al. 1980),
these results show an increasingly strong effect of the roughness sublayer throughout the experimental period (De Bruin et al. 1985; Verhoef et al. 1997).

8.3.3.6 Stability adjustment parameter

The parameter $\Psi$ (Dyer 1974; Paulson 1970) accounts for the effect of stability upon velocity profile similarity relationships based on neutral conditions. It may be calculated from values of wind speed, friction velocity, roughness length, measurement height and zero plane displacement height.

$$\psi = \ln \left( \frac{z - d}{z_0} \right) - \frac{u_\ast}{k U} \quad (8.11)$$

![Figure 8-17 Five year averaged stability adjustment parameter as a function of wind direction and Monin Obukhov stability. The filled circles represent the model of Paulson (1970), closed circles are results for Griffin. Error bars represent ± one standard error, n ranging from 2 to 10700 (average n = 500).](image)

The calculated values of the parameter $\Psi$ exhibit deviations from the predicted values for most wind directions, figure 8.17. The disagreement may be associated with neglect of the term incorporating momentum roughness length or because of measurements being made in the roughness sublayer (Nakamura & Mahrt 2001). Because measurements were made in the roughness sublayer the $\phi_M$ term from the
previous section should be added to the right hand side of equation 8.11. However, because $\varphi_M$ is stability independent and because no values of surface layer depth were available inclusion of this term would imply a constant positive offset to the value of $\Psi$ and would not explain the stability dependent deviations from the model. The data suggest closer agreement with the model of Paulson (Paulson 1970) for wind directions falling between east and south and the largest disagreement for winds from west to north. For other directions, data suggest that velocity profiles adhere more closely to Monin-Obukhov similarity theory despite stability conditions. The cause of this variation merits further investigation.

8.4 Sensible heat flux

Sensible heat flux was determined by eddy covariance methods using sonic temperature as well as thermocouple temperature. Sonic temperature sensible heat fluxes, $H$, were used as default values because they spanned the entire length of the experiment; sensible heat fluxes obtained using thermocouple measurements, $H_t$, were only available for three years.

8.4.1 Corrections to sensible heat flux

Corrected and run-time values of $H_t$ are compared in figure 8.18 and table 8.3. The correction factor of about 1.13 is 10% larger than those observed for momentum fluxes. The larger correction for $H$ results from the different cospectral models employed for sensible heat and momentum frequency response corrections, as well as adjustments for conversion of sonic temperature to true temperature and angle of attack corrections. Similar to momentum flux, the RMS error associated with this comparison was approximately 3% of the range of sensible heat flux values. Unlike momentum, scatter associated with the corrections was similar over all values of $H$.

The correction effect on $H_t$ is larger than that for $H$, table 8.3. The larger correction to $H_t$ is an effect of the separation distance between the sonic probe and the thermocouple probes on frequency response corrections. This conclusion is consistent in that the correction was larger for the thermocouple probe with the larger separation
from the sonic probe (0.3 m for TC#1 and 0.45 m for TC#2). The effect of the thermocouple time constant will have been negligible (τ = 0.025 s).

![Graph showing corrected and uncorrected values for sensible heat flux vs. year 1998 data. The solid line is the linear regression for the 1998 data and the dashed line is the 1:1 line.](image)

**Figure 8-18** Corrected, sonic temperature, sensible heat flux vs. uncorrected values obtained directly from Edisol using for the year 1998. The solid line is the linear regression for the 1998 data and the dashed line is the 1:1 line.

Regressions of corrected $H_t$ against $H$ (table 8.4) reveal good agreement between the final values, though fluxes from both thermocouples have a sizable intercept. The intercepts suggest that $H$ values were, on average, 10 W m$^{-2}$ more negative than those of $H_t$.

To check fluxes for sensor gain differences, run mean sonic and thermocouple temperatures were regressed against run mean psychrometer dry bulb temperatures, table 8.5. If the psychrometer $T_a$ is taken as the more accurate temperature, these results suggest that a 4% underestimation by the sonic temperature is more problematic than is the 2% thermocouple overestimation. In total, results indicate a possible 6% larger gain for the thermocouple measurements as compared to the sonic temperature. This is only partially reflected in the 1998 and 1999 data from thermocouple #2. (The poor correlations in 1997 result from a limited range of $H$ values).
Table 8.3 Linear regression of corrected against uncorrected sonic and thermocouple sensible heat flux, n > 10,000 for all years.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Year</th>
<th>Intercept</th>
<th>Slope</th>
<th>RMSE</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sonic</td>
<td>1997</td>
<td>-1.65</td>
<td>1.10</td>
<td>19.13</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>1998</td>
<td>-1.21</td>
<td>1.14</td>
<td>20.23</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>1999</td>
<td>-0.05</td>
<td>1.13</td>
<td>22.05</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>2000</td>
<td>-0.19</td>
<td>1.13</td>
<td>21.88</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>2001</td>
<td>-1.70</td>
<td>1.14</td>
<td>18.94</td>
<td>0.96</td>
</tr>
<tr>
<td>TC#1</td>
<td>1997</td>
<td>-2.4</td>
<td>1.16</td>
<td>13.4</td>
<td>0.91</td>
</tr>
<tr>
<td></td>
<td>1998</td>
<td>0.2</td>
<td>1.37</td>
<td>20.5</td>
<td>0.96</td>
</tr>
<tr>
<td>TC#2</td>
<td>1997</td>
<td>-0.4</td>
<td>1.50</td>
<td>13.2</td>
<td>0.77</td>
</tr>
<tr>
<td></td>
<td>1998</td>
<td>0.6</td>
<td>1.42</td>
<td>21.3</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>1999</td>
<td>-0.2</td>
<td>1.42</td>
<td>24.2</td>
<td>0.96</td>
</tr>
</tbody>
</table>

Table 8.4 Linear regression of corrected $H$ against corrected $H_t$, n > 10,000 for all years.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Year</th>
<th>Intercept</th>
<th>Slope</th>
<th>RMSE</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>TC#1</td>
<td>1997</td>
<td>-4.6</td>
<td>0.48</td>
<td>26.7</td>
<td>0.33</td>
</tr>
<tr>
<td></td>
<td>1998</td>
<td>9.2</td>
<td>1.00</td>
<td>24.5</td>
<td>0.94</td>
</tr>
<tr>
<td>TC#1</td>
<td>1997</td>
<td>-5.3</td>
<td>0.22</td>
<td>27.6</td>
<td>0.07</td>
</tr>
<tr>
<td></td>
<td>1998</td>
<td>11.3</td>
<td>1.02</td>
<td>29.4</td>
<td>0.92</td>
</tr>
<tr>
<td></td>
<td>1999</td>
<td>13.9</td>
<td>1.05</td>
<td>35.1</td>
<td>0.91</td>
</tr>
</tbody>
</table>

Table 8.5 Regression comparisons of run mean thermocouple and sonic temperatures with run mean psychrometer dry bulb temperatures.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Intercept</th>
<th>Slope</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sonic</td>
<td>0.43</td>
<td>0.96</td>
<td>0.96</td>
</tr>
<tr>
<td>TC#1</td>
<td>1.72</td>
<td>1.02</td>
<td>0.95</td>
</tr>
<tr>
<td>TC#2</td>
<td>-0.62</td>
<td>1.02</td>
<td>0.96</td>
</tr>
</tbody>
</table>

Although some of the difference between $H$ and $H_t$ may be explained by sensor gain, this does not address the existence of an offset between the two methods. A more direct analysis was accomplished by observing relationships of the flux difference ($H_t - H$) in relation to environmental variables. Wind speed showed the strongest of what were generally weak relationships. Further analysis revealed an associated wind
direction dependence. For winds from 0 to 240 degrees, strong winds caused $H$ to be more negative than $H_t$ while for winds from 240 to 360 degrees there was closer agreement between $H$ and $H_t$ (figure 8.19). This pattern of difference was caused by the geometry of the sonic anemometer probe and its use of a single transducer pair to measure sonic temperature. The problem is discussed in detail in Appendix H.

![Figure 8-19 Difference in sensible heat fluxes ($H_t - H$) by wind direction and wind speed. Error bars represent ± one standard error, n ranging from 5 to 7500 (average n = 650).](image)

From Appendix H, it can be concluded that head distortion of the Gill Solent R2 anemometer affects derived sensible heat fluxes at all wind speeds and is a function of wind direction. These observations are in general agreement with those of Grelle et al. (1996), although they suggest an effect only for $U$ higher than 7 m s$^{-1}$ and they do not suggest a wind direction dependence. A detailed quantification of this error will require wind tunnel testing of the sonic anemometer. For final flux processing, an empirical adjustment was made to the sonic sensible heat fluxes of the form:

$$\Delta H = \begin{cases} 8.58 \cdot U + 0.8701 \cdot U^2 & \text{for } \theta < 240 \text{ and } 340 < \theta \\ 13.69 \cdot U - 1.7105 \cdot U^2 & \text{for } 240 < \theta < 340 \end{cases}$$  (8.12)

Offsets are not included in equation 8.12 because differences between $H$ and $H_t$ at zero wind speed could not have been caused by sensor head deformation.
8.4.2 Temporal Variability

The annual mean monthly sensible heat flux increases from a December minimum of about \(-45\, \text{W m}^{-2}\) to a May maximum of about \(70\, \text{W m}^{-2}\) (figure 8.20). This pattern of \(H\) roughly follows the pattern of net radiation. Similar patterns have been observed for spruce (Bernhofer et al. 2003), and Douglas fir (Humphreys et al. 2003) forests. In contrast, sensible heat fluxes for broad leaved (Blanken et al. 2001; Schmid et al. 2000), mixed coniferous (Turnipseed et al. 2002) and pine (Berbigier et al. 2001) forests have been observed to be more severely reduced during summer as a result of enhanced latent heat fluxes associated with more continental climates and leaf-out in the case of broad leaved forests.

![Figure 8-20 Monthly mean values of sonic sensible heat flux, \(H\), for each year (open circles) and monthly average for all five years of data (closed circles).](image)

The diel pattern of \(H\), figure 8.21, also follows the general pattern of net radiation. Maximum diurnal values are obtained just after midday while minimum values occur...
just after sunset (\(\text{i.e. } \Pi_d = 100\%\)). The only period which shows significant deviation from the typical diel pattern are the months of January and February, when \(H\) appears to remain high later in the day than in other months. A reduction in \(H\) following midyear is apparent when comparing the diel curves before and after midyear.

![Figure 8-21 Diel plot of sensible heat flux, \(H\), averaged by bi-monthly periods. Values are plotted in relation to percent of the diurnal/nocturnal period. Error bars represent ± one standard error, \(n > 400\).](image)

![Figure 8-22 Sensible heat flux fraction of available energy, midday by month (left panel, \(n > 1000\)) and by percent of day for spring and summer (right panel, \(n > 4000\)). Error bars represent ± one standard error.](image)

The monthly averaged midday (35% < \(\Pi_d < 65\%\)) sensible heat flux fraction (\(H/\left(R_n + G\right)\)) presented in figure 8.22, shows the reduction in \(H\) as a proportion of available energy after the month of August. This reduction occurs despite the lower soil moisture content (figure 4.77) and higher vapour pressure deficit (figure 4.46) after midyear. This asymmetry may be caused by larger energy dissipation by transpiration associated with higher evaporative demand and the formation of new leaves after midyear. Such a pattern is consistent with the more obvious annual patterns of
sensible heat flux observed for broad leaved species (Blanken et al. 2001) and mixed conifers (Turnipseed et al. 2002). Such a hypothesis is also consistent with the diurnal pattern of sensible heat fraction in figure 8.22, which also shows decreases in afternoon $H/(R_n + G)$ associated with increases in $D$ and reductions in canopy conductance (see section 8.5.3.2).

![Figure 8-23 Date/time contour plot of half hour values of sensible heat flux, $H$, over five years.](image)

A date/time plot of sensible heat flux, figure 8.23, shows summer midday values of approximately 200 W m$^{-2}$ and winter midday values of -50 W m$^{-2}$ or less. The annual asymmetry is apparent in mid-day values, and a trend towards more negative nocturnal $H$ is apparent towards the end of the experiment.

### 8.4.3 Environmental Variability

To determine which factors controlled the inter-annual variability of $H$, the differences of annual monthly-mean values from the five-year average monthly means were regressed against similar differences for other environmental variables. The regressions' correlation coefficients were taken as indicators of the relevance of the independent variable to the variability of $H$, table 8.6.

The results presented in table 8.6 indicate that absorbed or global radiation was the dominant factor controlling the inter-annual variability of sensible heat flux at Griffin. Humidity variables are next most important while other variables appear to play insignificant, or at least indirect, roles in affecting inter-annual variability of $H$. 

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Table 8.6 Ranked correlation coefficients for regression of monthly deviation from five year mean sensible heat flux against various environmental variables. The columns labelled Radiation, Humidity, and Other are visual assistants to ease identification of variable groupings.

<table>
<thead>
<tr>
<th>Variable</th>
<th>$R^2$</th>
<th>Radiation</th>
<th>Humidity</th>
<th>Other</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_{of}$</td>
<td>0.16</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_g$</td>
<td>0.13</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.13</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$D_{S}$</td>
<td>0.08</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{ds}$</td>
<td>0.08</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{ps}$</td>
<td>0.07</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$h_r$</td>
<td>0.07</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$D$</td>
<td>0.06</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{or}$</td>
<td>0.04</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$U$</td>
<td>0.03</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_t$</td>
<td>0.01</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Albedo</td>
<td>0.01</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$e$</td>
<td>0.01</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T_r$</td>
<td>0.00</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{of}/Q_{ps}$</td>
<td>0.00</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T_s$</td>
<td>0.00</td>
<td>X</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

8.4.3.1 Relationship to Wind Speed and direction

The general pattern and annual deviations from the five-year mean of sensible heat flux are similar to those observed for net radiation fluxes during spring and summer (section 4.8). However, the autumn and winter minima in $H$ and $H_t$ occurring in the months of October, December, and February, are dissimilar to the observed patterns of radiation.

These minima were examined by comparing the five year average monthly $H$ and $H_t$ in figure 8.24 with corresponding values of wind speed and net radiation. It is apparent that the high winter wind speeds correspond to values of $H$ and $H_t$ that are more negative than for low mean wind speeds.

An effect of $U$ upon $H$ could be expected if the corrections made using equation 8.12 were inappropriate. Such an error may be responsible for the 10 W m$^{-2}$ difference between the two measures of sensible heat flux shown in figure 8.24. This offset could have resulted from errors common to the correction of sensible heat fluxes measured by both methods. However, the only commonly applied corrections were
aspects of frequency response, which would have also been present in other flux corrections – yet similar sensitivities to wind speed to not appear in these fluxes. The observation of reductions in both $H$ and $H_1$ with mean wind speed suggests that the problem is not associated with the temperature measurement techniques, since different temperature measurement methods were employed. A plot of sensible heat flux fraction of available energy by wind direction (figure 8.25) shows dips in the sensible heat flux fraction for winds from east-north-east and south-west directions. This alignment suggests effects related to the row structure of the canopy (see section 8.3.2). It would also suggest reduced values of $H$ for flow along the canopy rows, which may be appropriate if better ventilation of the canopy is associated with high evapotranspiration losses.

![Figure 8-24 Monthly mean values of sensible heat flux from sonic (closed circles) and thermocouple (open circles) temperature measurements. Data used in averages are from periods when both measurements were available. Error bars represent ± one standard error, $n > 1000$.](image)
Temperature profile data were examined to address the possibility of a natural cause of more negative $H$ for along row wind flow. The differences between mean within-canopy and above canopy temperature were averaged by $R_n$ and wind direction classes. The resulting temperature differences, presented in figure 8.26 indicate lower within canopy temperatures for along row winds (north-east and south-west directions). This pattern closely matches that of the sensible heat flux fraction shown above and also the effect of row structure on velocity profiles observed in figure 4.59. Together these observations indicate that along-row winds result in lower within canopy temperatures and a reduction in the loss of available energy as sensible heat.
A further discussion of the observed relationship of sensible heat flux to wind speed is delayed until the discussion of energy budget closure so that effects on latent heat flux can also be taken into account.

8.4.3.2  Relationship to stability

As with momentum, similarity theory suggests turbulent intensity \( \sigma_T/T_* \) should be constant under stable conditions and a function of stability under unstable conditions (Stull 1988).

\[
\frac{\sigma_T}{T_*} = \begin{cases} 
2.0 & \text{stable} \\
1.61 \left(-1 \cdot \frac{z-d}{L}\right)^{-0.281} & \text{stable proposed} \\
0.95 \left(-1 \cdot \frac{z-d}{L}\right)^{1.3} & \text{unstable}
\end{cases}
\]

(8.13)

The stability-averaged values of \( \sigma_T/T_* \), (figure 8.27), indicate that as stability shifts from unstable towards neutrality, turbulent intensities of both sonic and thermocouple temperature increase in magnitude as predicted by equation 8.13. As neutral conditions are approached from the unstable side, both curves begin to rise more rapidly than the model curve, achieving maxima of about 10. For near neutral stable conditions the turbulent intensities exceed the predicted value of 2 (Dias & Brutsaert 1996; Hicks 1981; Wesely 1988) but tend towards this value as stability increases.

The larger than expected values for near neutral conditions are caused either by higher than predicted temperature variance or smaller than predicted fluxes. The most probable cause of the model overestimate is the poor signal to noise ratio observed under near neutral conditions (McBean 1971). The high proportion of noise under near neutral conditions increases temperature variance without affecting \( H \). Such a condition would result in an overestimate of \( \sigma_T/T_* \) for both stable and unstable near-neutral conditions, as observed in figure 8.27. However, if the overestimate under stable conditions is similar to that observed under unstable conditions then the constant value model of \( \sigma_T/T_* \) is inappropriate and the proposed power law model given in equation 8.13 appears to apply. The observed relationships are in agreement.
with other research for unstable conditions (Lee & Black 1993; Wyngaard 1973; Zhang et al. 2001) but few results are available for stable conditions (McBean 1971) making comparisons difficult.

Figure 8-27 Average turbulent intensities for sonic anemometer (filled circles) and thermocouple (open circles) temperature as a function of stability. Dashed lines represent the model relations given in equation 8.13, and error bars represent ± one standard error. The values of the standard deviations of \( T \) were not corrected for frequency response in the values given. Error bars represent ± one standard error; \( n > 100 \).

8.4.3.3 Spatial patterns

Flux footprints were analysed to determine the potential for coincidence of environmental variables and fluxes with undesirable aspects of the landscape at Griffin. Average footprints were obtained by multiplying probability footprints for each half hour by the variable of interest and then averaging over the desired period (one or five years). For aerodynamically measured fluxes this will have produced a time averaged flux footprint similar in form to those commonly available. For other variables, the resulting footprint represents the conditions of the particular variable corresponding to the aerodynamic fluxes. In this analysis, the footprint model employed by Bakwin et al. (1992) was employed. Although it is recognized that more accurate models may be available (Schmid 2002), the purpose of this site analysis was primarily qualitative.

The peaks of the probability footprint, figure 8.28, indicate the strong effect upon the potential flux source distribution of along valley flow. The smaller peak in probability to the south-southwest is associated with conditions of stronger wind flow,
as described in section 4.13. The patterns of both net and short wave radiation are very similar. This suggests that the pattern of nocturnal net radiation is similar to that observed for diurnal $R_n$ and does not selectively diminish diurnal $R_n$ gains. It should be noted that for both $R_n$ and $R_g$ there appear to be higher probabilities of radiation for winds from the north-west. It is likely that the higher radiation from these directions is a result of the association of clearer conditions with passage of high-pressure systems and their associated more northerly flow at this latitude, as can be inferred from the higher precipitation for more southerly flow.

Figure 8-28 Five-year cumulative flux footprint probability as well as cumulative flux footprints of short-wave radiation, net radiation, and precipitation. North is towards top of figure. The grey shading is a representation of an aerial photograph of this site.

For sensible heat flux, (figure 8.29), negative fluxes were associated with higher mean winds from westerly and south-westerly wind directions (see figure 4.63). The broad peaks in the footprints observed for $R_n$ appear for $H$ to be damped and complicated by the combination of nocturnal and diurnal $H$ values associated with flow along the valley axis. There appears to be little or no net positive $H$ associated with flow from the southwest.
8.4.3.4 Gap filling

Although the eddy covariance methods were able to obtain data coverage of 70% for sensible heat flux, it was also desirable to have information on sensible heat flux for the missing 30% of runs. Although standard gap filling techniques can be employed, it is preferable to base estimations of missing fluxes on relevant measured values. To this end, methods of determining the flux of a scalar from the standard deviation of the scalar were attempted. This method, described by Tillman (1972), Lloyd et al. (1991), etc employs similarity relations to estimate the sensible heat flux:

\[ H = \rho \cdot c_p \sqrt{\frac{\sigma_T}{0.95}} \left( \frac{k \cdot g \cdot (z - d)}{T + 273.16} \right) \]  \hspace{1cm} (8.14)

This method proved reasonably accurate when applied to the standard deviations of the sonic anemometer and the fast response thermocouple temperature values (figure 8.30). Some of the observed scatter may have been related to the assumption of an azimuthally constant value of \( d \). Considering the result observed for \( z_0 \) in section 8.3.2.5 this is probably a poor assumption.
Nevertheless, the purpose was to estimate $H$ when these data were not available. Therefore, the only available values of temperature standard deviations were from the psychrometers located on the flux and profile towers. While standard deviations of these values were available, the low sampling rate (0.5 Hz) and the long time constant (50 s) of these sensors introduced unacceptable scatter in the estimation of $H$ via this method, (figure 8.30). Much of this scatter will have resulted from inaccuracies caused by error in the large frequency response correction required for these sensors. However, the results for the sonic temperature standard deviations suggest that much benefit could be gained by measurement of a reasonably fast responding temperature sensor with an alternative logging system, e.g., a data logger operating at 1 Hz.

![Figure 8-30 Comparison of sensible heat flux estimated from temperature standard deviations with eddy covariance sensible heat fluxes obtained using sonic temperature. The left panel contains values estimated from the standard deviation of sonic temperature while those in the right panel were estimated from the standard deviations of temperature measured with the 14.7 m psychrometer.](image)

Because no suitable measurements were available to estimate $H$ when eddy covariance values of $H$ were not available, gap filling of $H$ was calculated as the residual of the energy balance equation, using gap-filled values of $R_n$, $G$, and $\lambda E$.

$$H = R_n + G - \lambda E$$ (8.15)
8.5 Latent heat flux

The following sections describe the measured latent heat flux, \( \lambda E \), at the Griffin site and its patterns in relation to time, space and relevant controlling variables.

8.5.1 Correction effects

A comparison of corrected and runtime values of \( \lambda E \), (figure 8.31 and table 8.7,) indicate a large correction effect. The cause of this rather large correction is discussed in sections 5.3.2.7 and 5.3.2.11. The scatter associated with the correction is larger than that observed for either momentum or sensible heat fluxes with the RMS error about 8% of the entire range of observed values. The correction effects on \( \lambda E \) were larger at the beginning of the experiment and declined by about 40% by the end of the experiment in 2001.

![Figure 8-31 Correction effect on latent heat flux for the year 1998. Uncorrected values are those obtained directly from the Edisol data collection software while corrected values are the results of the procedures described in Chapter 5. The solid line is the linear regression for the 1998 data and the dashed line is the 1:1 line.](image)

The observed reduction in the correction factor over the period of the experiment could have been the result of either a change in the characteristics of frequency response attenuation or, considering that most attenuation was at high frequencies, a shift in cospectra towards lower frequencies. Such a shift may increase low frequency loss while reducing high frequency losses. There were no known changes in the
instrument characteristics that control sensor frequency response so this may be dismissed as the cause of the reduction in corrections. However, a shift to lower frequencies of $\lambda E$ cospectra is consistent with the growth of the canopy over the period of the experiment. Because the height above zero plane displacement decreases faster with canopy growth than does wind speed, the normalized frequency ($f = n (z-d)/U$) will also decrease with growth of the canopy. It is interesting to note that this effect does not noticeably affect the corrections of momentum flux, $H$ or $F_e$.

The corrections of these fluxes are probably not affected because shifts to lower frequency increase low frequency losses in proportion to decreases in high frequency losses. The high frequency attenuation of $\lambda E$ is so large that the increase in low frequency loss is small in comparison to improvements in high frequency attenuation.

### Table 8.7 Linear regression of corrected against uncorrected latent heat flux, n > 10,000 for all years.

<table>
<thead>
<tr>
<th>Year</th>
<th>Intercept</th>
<th>Slope</th>
<th>RMSE</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>2.49</td>
<td>2.11</td>
<td>23.4</td>
<td>0.84</td>
</tr>
<tr>
<td>1998</td>
<td>1.48</td>
<td>2.14</td>
<td>19.0</td>
<td>0.87</td>
</tr>
<tr>
<td>1999</td>
<td>2.52</td>
<td>2.07</td>
<td>25.9</td>
<td>0.81</td>
</tr>
<tr>
<td>2000</td>
<td>2.73</td>
<td>1.90</td>
<td>22.7</td>
<td>0.87</td>
</tr>
<tr>
<td>2001</td>
<td>4.38</td>
<td>1.71</td>
<td>22.1</td>
<td>0.87</td>
</tr>
</tbody>
</table>

### 8.5.2 Temporal variation

The monthly mean latent heat flux rises from near 0 W m$^{-2}$ during December and January to approximately 60 W m$^{-2}$ by May, (figure 8.32) and remain at this magnitude through August before decreasing again in the autumn.

The steady values of $\lambda E$ during the summer suggest that $\lambda E$ may be limited during summer months or that $\lambda E$ is enhanced before and after midyear. Higher values of $\lambda E$ prior to midyear is consistent with the higher overall energy input during May while enhanced $\lambda E$ after midyear is consistent with the observation of enhanced midday $H$ prior to midyear (figure 8.20) and is consistent with higher values of $D$ after midyear and the deployment of new leaves by the canopy, as mentioned in section 8.4.2. Beadle et al. (1985) have also shown that the stomatal conductance (see
section 8.5.3.2 below) of new Sitka spruce leaves is higher than for older leaves. Although many experiments show a lag in annual peak of $\lambda E$ (Berbigier et al. 2001; Black et al. 1996; Grelle et al. 1996; Schmid et al. 2000; Turnipseed et al. 2002), most experiments also show some corresponding limitation to mid-summer $H$. Only Humphreys et al. (2003) showed a lag in $\lambda E$ and value of $H$ that did not show summer limitation. No experiments were found that showed a peak prior to mid-year, which may not be surprising if we consider the unique climatological conditions of the Griffin site.

![Figure 8-32 Monthly mean values of latent heat flux for each year, (open circles). Closed circles represent the monthly average for all five years of data.](image)

The diel patterns of latent heat flux shown in figure 8.33, indicate that $\lambda E$ is enhanced after midday. From this figure it can be determined that mid-afternoon fluxes are larger than mid-morning fluxes by a difference equivalent to 28% of midday fluxes.
Figure 8-33  Diel pattern of latent heat flux averaged for bi-monthly periods. Error bars represent ± one standard error, \( n > 400 \).

Figure 8-34  Evaporative fraction of available energy, midday by month (left panel, \( n > 1000 \)) and by percent of day for spring and summer (right panel, \( n > 4000 \)). Error bars represent ± one standard error.

The relative increases in \( \lambda E \) are more readily apparent for both annual and diurnal time scales when plotted as evaporative fraction (\( i.e. \) the ratio of \( \lambda E \) to available energy), in figure 8.34. It is interesting to note the similar trends in the patterns of annual evaporative fraction and \( D \) (section 4.10) but that the annual pattern of soil moisture follows an opposing trend, figure 4.77. Such a relationship with soil moisture indicates that it is probably not a limiting factor at this site during the summer, despite the fact that there ceases to be a surfeit of soil water during the summer months, (see section 8.5.6).
The delay in peak midday $\lambda E$ to after midyear is identifiable in the date/time plots of $\lambda E$, (figure 8.35). Another notable feature is the similarity of nocturnal $\lambda E$ to nocturnal $H$, with both date/time plots suggesting more negative values during the latter half of the experiment.

![Date/time contour plot of half hourly values of latent heat flux, $\lambda E$, over five years.](image)

**Figure 8-35** Date/time contour plot of half hourly values of latent heat flux, $\lambda E$, over five years.

### 8.5.3 Environmental variability

Table 8.8 Ranked correlation coefficients for regression of monthly deviation from five year mean sensible heat flux against various environmental variables. The columns labelled Radiation, Humidity, and Other are visual assistants to ease identification of variable groupings.

<table>
<thead>
<tr>
<th>Variable</th>
<th>$R^2$</th>
<th>Radiation</th>
<th>Humidity</th>
<th>Other</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_g$</td>
<td>0.11</td>
<td>X</td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>$D$</td>
<td>0.10</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_t$</td>
<td>0.09</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$RH$</td>
<td>0.09</td>
<td></td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>$APAR$</td>
<td>0.09</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pg}$</td>
<td>0.09</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$U$</td>
<td>0.07</td>
<td></td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>$Q_{pr}$</td>
<td>0.04</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.04</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T_a$</td>
<td>0.04</td>
<td></td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>$T_s$</td>
<td>0.03</td>
<td></td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>Albedo</td>
<td>0.03</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$P_g$</td>
<td>0.03</td>
<td></td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>$E$</td>
<td>0.01</td>
<td></td>
<td>X</td>
<td></td>
</tr>
<tr>
<td>$Q_{pf}$</td>
<td>0.00</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pg}$</td>
<td>0.00</td>
<td>X</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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As with sensible heat flux, the inter-annual variability of latent heat flux was analysed by regressing the deviations of the annual monthly means against those for potential controlling variables, (table 8.8). The variability of $\lambda E$ appears to be almost equally controlled by variations in radiation and humidity deficit. Absolute measures of humidity, radiation balance metrics, and variables such as wind speed, temperature and precipitation have poorer correlations. This result is consistent with common models of water vapour exchange, and will be covered further in section 8.5.3.2.

8.5.3.1 Relationship to stability

The relationship of turbulent intensity of water vapour to Monin-Obukhov stability, (figure 8.36), was expected to be identical to that observed for sensible heat in section 8.4.3.2. However, for near neutral unstable conditions, $\sigma_{T}/Q*$ underestimates model predictions whereas $\sigma_{T}/T*$ overestimated the model. The underestimate is a consequence of using values of $\sigma_{Q}$ in the calculation of turbulent intensity that were not frequency response corrected. As a result, the large correction factor of $\lambda E$ may have overwhelmed the possible effect of signal noise in $\sigma_{Q}$ to produce the observed overestimates of turbulent intensity. This hypothesis, is consistent with the revised model for stable conditions, where the behaviour of $\sigma_{Q}/Q*$ also suggests underestimation at near neutral conditions which approaches expected model values under more stable conditions. Under stable conditions both the poor signal to noise ratio observed in the water vapour signal and the potential for large error in small nocturnal fluxes might have combined to produce an underestimate. However, given the potentially large errors in water vapour flux and variance as a result of tube attenuation, it would be inappropriate to speculate on the dominant cause of the non-adherence of $\sigma_{Q}/Q*$ to model predictions under stable conditions. Nevertheless, the decline of $\sigma_{Q}/Q*$ with stability, which is less rapid than that observed for temperature, does agree with the observations of McBean (1971).
Figure 8-36 Average water vapour turbulent intensities as a function of stability. Dashed lines represent model relations given in equation 8.13, and error bars represent one standard error. The values of the standard deviations of water vapour were not frequency response corrected in this calculation. Error bars represent ± one standard error, n > 100.

8.5.3.2 Estimation and modeling

Because the behaviour of latent heat fluxes of vegetated surfaces are well known and well studied, the analysis of the behaviour of $\lambda E$ at Griffin is carried out in the context of a model describing that behaviour. The widely employed Penman-Monteith model (Monteith 1964) relates the amount of water lost from an ecosystem to appropriate controlling variables. In this model, controlling variables are combined into terms that describe the ability of water vapour to flow from its source to its ultimate sink. These terms are referred to as resistances or conductances, one being the inverse of the other. The form of the model, as presented by Montieth (1973) is given as,

$$\lambda E_{PM} = \frac{A}{s + \gamma \frac{r_n}{r_s}}$$

(8.16)

In this model the variable $s$ is the slope of the curve relating saturated water vapour pressure to temperature obtained by rearranging Tetens’ formula of the Clausius-Clapeyron equation (Stull 2000) and can be described in terms of air temperature $T$ (units C) and saturation vapour pressure, $e_s$ (units kPa) as

$$s = \frac{17.269e_s}{T + 237.3} \left( 1 - \frac{T}{T + 237.3} \right)$$

(8.17)
The variable $\gamma$ is the psychrometric constant (units of $\text{Pa} \, \text{C}^{-1}$), which is a function of pressure, $P$ (units $\text{Pa}$), the specific heat of air at constant pressure, $c_p$ (units $\text{J} \, \text{g}^{-1} \, \text{C}^{-1}$), the latent heat of evaporation, $\lambda$ (units $\text{J} \, \text{g}^{-1}$), and $\varepsilon$, the ratio of the molecular weight of water to that of air.

$$\gamma = \frac{P \cdot c_p}{\lambda \cdot \varepsilon} \quad (8.18)$$

The variable $A$ is available energy (units of $\text{W} \, \text{m}^{-2}$), which is defined as net radiation minus the non-atmospheric storage components.

$$A = R_n - (S_{\text{soil}} + S_{\text{canopy}} + S_{\text{water}}) \quad (8.19)$$

The other variables, $r_1$, $r_a$, and $r_c$ are the resistances that describe the time taken for water vapour to transverse a unit distance along its path from source to sink. All resistances may be expressed in terms of conductance, which is the inverse of the resistance $g = (1/r)$. These terms will be used interchangeably as needed. The climatological resistance $r_1$ results from the climatological conditions at a site, specifically the ratio of vapour pressure deficit (units kPa) to available energy.

$$r_1 = \frac{\rho \cdot c_p \cdot D}{\gamma \cdot A} \quad (8.20)$$

The aerodynamic resistance $r_a$ quantifies the resistance to travel of water vapour molecules as a function of wind and friction velocity. This term also includes adjustments for the diffusivities of heat, $D_h$, and the species of interest, in this case water vapour, $D_w$, (Kim and Verma, 1989).

$$r_a = \frac{U}{u^*} + \frac{2}{k \cdot u^*} \left( \frac{D_h}{D_w} \right)^{\frac{2}{3}} \quad (8.21)$$

Although not explicitly shown in equation 8.21, the magnitude of $g_a$ will depend upon site surface characteristics, through the ratio $U/u^*$ (see section 8.3.2). For example, Teklehaimanot and Jarvis (1991) have observed $g_a$ to decrease with increased tree spacing. The wind direction dependency of $C_D$ observed in section 8.3.2 will have affected $g_a$ values, however, variability of this nature has not been taken into account in the estimation of $AE$ for gap filling.
The canopy resistance variable, $r_c$, represents the stomatal resistance and leaf boundary layer resistance of the canopy and surface treated as a single evaporating element. An equation describing canopy conductance can be obtained by inverting the Penman-Montieth (equation 8.16),

$$r_c = r_e \left( \frac{\beta_s}{\gamma} - 1 \right) + r_l (\beta + 1)$$

(8.22)

were $\beta$ is the Bowen ratio, $H/\lambda E$. For the analysis of Griffin data, values of $g_c$ were obtained using equation 8.22 ($g_c = 0.0009 \pm 0.0642$ m s$^{-1}$), and alternately by solving equation 8.16 using non-linear equation methods ($g_c = 0.0095 \pm 0.0864$ m s$^{-1}$). A smaller skewness (i.e. more negative values) of the distribution of $g_c$ obtained from equation 8.22 resulted in the smaller value. Because equation 8.16 provided values of canopy conductance that were more physically realistic, the values obtained by that method were employed in analysis of the data.

**Temporal patterns of conductance**

Perhaps the simplest temporal pattern of conductance to explain is that of aerodynamic conductance, $g_a$. Both the annual (figure 8.37) and diel (figure 8.38) patterns of $g_a$ closely follow the corresponding patterns of wind speed, as can be expected from equation 8.21.

Climatological conductance, $g_l$, exhibits a more distinctive pattern which is similar for both the annual and diel cycles. For both temporal cycles $g_l$ obtains a maximum shortly after the beginning of the cycle (i.e. March for the annual cycle and mid-morning for the diel cycle) and gradually declines to a minimum at the end of the cycle. This pattern is caused by the temporal shift of the pattern of vapour pressure deficit (see figures 4.46), with a peak slightly after the mid-year and mid-day peak in available radiation (see figures 4.30 and 4.32). Slight differences in diel and annual patterns of $g_l$ are caused by the effect of different phase shifts of diel and annual $T_a$ (see section 4.9.1) on patterns of $D$. 
Figure 8-37 Monthly average aerodynamic, $g_a$, climatological, $g_i$, and canopy, $g_c$, conductance. Data from all five years were used but were limited to midday periods (20 to 80% of the diurnal period) and to periods when cumulative net radiation following a precipitation event was greater than 9000 W m$^{-2}$.

Figure 8-38 Diel patterns of aerodynamic, $g_a$, climatological, $g_i$, and canopy, $g_c$, conductance. Data from all five years were included but were limited to the periods when cumulative net radiation following a precipitation event was greater than 16 MJ m$^{-2}$.

The annual and diel patterns of canopy conductance, $g_c$, reflect the patterns of $g_a$ and $g_i$ combined with the corresponding patterns of the $\beta$ (see section 8.6.1). The annual pattern of diurnal $g_c$, figure 8.37, is relatively constant throughout the year. Only
during December and January does $g_c$ exhibit a notable decrease. The constancy of annual $g_c$ suggests that the canopy is not water stressed. The diel pattern is slightly different, having a mid-morning peak with a gradual diurnal decline followed by a more rapid drop at sunset and very low, relatively constant nocturnal values. This pattern appears as a more obvious combination of $g_a$ and $g_i$. Possible causes of this diurnal pattern of $g_c$ may indicate reductions in evaporation losses of surface water over the diurnal period or reductions in stomatal conductance in response to higher afternoon $D$.

$g_c$ relation to precipitation

The presence of liquid water on the canopy surface can result in considerable errors in estimates of $g_c$. In such situations the canopy water vapour conductance tends toward infinity such that equation 8.16 is modified so that it describes the potential water loss, i.e., the latent heat flux loss assuming a continuously saturated surface:

$$\lambda E_{\text{pot}} = A \left( s + \gamma \frac{r_i}{r_a} \right) (s + \gamma^{-1})^{-1}$$  \hspace{1cm} (8.23)

![Graph showing relationship of canopy conductance to $D$ for different levels of accumulated net radiation since last measurable precipitation event. Error bars represent ± one standard error, n ranging from 5 to 4000 (average n = 650).](image)

While the forest canopy was frequently wetted by precipitation, it was observed in figure 4.68 that periods between precipitation events lasted about two hours.
Coincidentally, Teklehaimanot and Jarvis (1991) noted that a Sitka crown took about two hours to dry after wetting on a warm ($T_a \sim 22$ C) summer day. Analysis of the Griffin data suggested that canopy conductance values tended towards values dominated by transpiration after approximately 9 MJ of net radiant energy had accumulated since the last measurable rainfall event, figure 8.39. This threshold was employed as a preliminary estimate of canopy drying while more appropriate methods were employed to account for canopy water in the final estimation of model evapotranspiration.

$g_c$ relation to $D$

As indicated by its presence in the definition of $\gamma$, and as is apparent from the analysis of inter-annual variability (table 8.8), $D$ is the most appropriate atmospheric humidity measure affecting the control of the exchange of water vapour (Aphalo & Jarvis 1991). While the effect of $D$ upon canopy conductance may be affected by wind speed (Aphalo & Jarvis 1993), soil moisture (Turner 1991; Tuzet, Perrier, & Leuning 2003), and plant water status (Beadle et al. 1978; Waring, et al. 1979), it has been shown that canopy conductance declines with an increase in the value of $D$ (Sanford & Jarvis 1986).

The values of $g_c$ from the Griffin data set have a nearly linear, negative relationship to $D$ for all but the lowest levels of $D$, figure 8.39. Also shown in this figure are data for which canopy surface water has not been taken into account. We see that such data have a more non-linear relationship to $D$ and increase more rapidly at low values of $D$.

Re-examination of the annual patterns of $\lambda E$ (figure 8.32) and $D$ (figure 4.46) suggests a similarity in ‘flat topped’ summer patterns for both variables. This pattern suggests that $D$ is acting to enhance $\lambda E$ and not suppress it, as would be suggested by figure 8.40. However, the influence of factors such as plant and soil water status and their interaction with $D$ in affecting canopy conductance have not been examined because of insufficient data.
Figure 8.40 Relationship of canopy conductance to vapour pressure deficit for different levels of net radiation for a dry canopy. The dotted line represents a similar curve for which both dry and wet canopy conditions were considered. Error bars represent ± one standard error, n ranging from 5 to 4000 (average n = 650).

$g_c$ relation to radiation

The relationship of canopy conductance to radiation, figure 8.41, indicates that, irrespective of values of $D$, the canopy conductance only responded to radiation up to a value of between 300 and 400 W m$^{-2}$. This observed relationship is consistent with that of Morison & Jarvis (1983), who employed a rectangular hyperbola to describe the response of $g_c$ to $R_g$ for Sitka spruce.

In addition to the observed variation in $g_c$ caused by radiation and atmospheric humidity deficit, temperature has also been shown to affect stomatal or canopy conductance (Neilson et al. 1972). However the results of Neilson et al. (1972) indicate that the effect of temperature is to increase stomatal conductance appreciably below about −5°C. Because such temperatures were observed infrequently at Griffin, no attempt was made at determining the temperature response of canopy conductance.
Based on the above findings, an empirical model describing the relationship of canopy conductance to $D$ and $R_g$ was established assuming a linear response to $D$ and hyperbolic response to $R_g$.

$$g_c = (0.0143 - 0.0083 \cdot D) \left( 1 - 0.84 \cdot e^{-0.01 \cdot R_g} \right)$$  \hspace{1cm} (8.24)

This model was used in equation 8.16 for the estimation of values of $\Delta E$ when measured values were not available. A comparison of this model will be described below.

### 8.5.4 Model comparisons

Employing the Penman-Monteith model to estimate the ecosystem evapotranspiration required discrimination between conditions of wet and dry canopy. Equation 8.25 was used to model canopy evapotranspiration using the Penman-Monteith formulae for both saturated and dry canopy conditions. In this equation, the two evapotranspiration terms (equations 8.16 and 8.23) were weighted by an exponential relationship of canopy saturation, incorporating the existing ($W$) and maximum ($W_{\text{max}}$) water-holding capacity of the canopy.
The amount of water lost from the canopy was determined using the potential evapotranspiration, the amount of water added to the canopy as precipitation, $E_{\text{pcp}}$, (assuming 25% interception of precipitation, see section 4.14.1.1) and the amount of water pre-existing on the canopy, $W_0$.

$$W = W_0 + E_{\text{pcp}} - \lambda E_{\text{Pot}} \left( \frac{W_0}{W_{\text{max}}} \right) \left( \frac{1.8}{\lambda} \right)$$ (8.26)

A select example of the comparison of measured $\lambda E$ with the models for $\lambda E_{\text{Pot}}$, (equ. 8.23), and $\lambda E_{\text{model}}$ (equ. 8.25, employing the empirical $g_e$ relationship of equation 8.24 in the value of $\lambda E_{\text{PM}}$), is shown in figure 8.42. It can be seen, for this example of a drying canopy in June of 1998, that the modelled values of $\lambda E_{\text{model}}$ provide a reasonable representation of the measured values of latent heat flux.

![Figure 8-42 Comparison of measured and modelled latent heat flux during a drying event. Circles are measured values of $\lambda E$. The dotted line is $\lambda E_{\text{Pot}}$, the dashed line is $\lambda E_{\text{PM}}$ and the solid line is $\lambda E_{\text{model}}$. The top panel is canopy water storage.](image-url)
Further, regression comparisons of the modelled and measured $\lambda E$ values shown in figure 8.43 indicated that $\lambda E_{\text{model}}$ overestimated measured $\lambda E$, by about 50%, under conditions of low $D$. This overestimation was the result of an approximate 15 W m$^{-2}$ bias in $\lambda E_{\text{model}}$ at all levels of $D$. This discrepancy may have been caused by excess evaporation associated with incorporating canopy wetness into equation 8.25 or may have resulted from unaccounted for underestimations in measured $\lambda E$. Lack of energy budget closure after flux correction (see section 8.6.2 below) suggests that measured $\lambda E$ may be at fault, though there is no way of checking model values to verify this conjecture.

8.5.5 Spatial distribution

8.5.5.1 Relationship to wind speed and direction

Although the Penman-Monteith model does not contain a term that would imply a relationship of $\lambda E$ to wind direction, the results obtained for the dependence of sensible heat flux upon wind direction indicated the possibility of a similar, coincidental dependence of $\lambda E$ upon wind direction. The values of evaporative fraction ($\lambda E/(R_n-G)$) were therefore analysed in a manner similar to those for sensible heat flux fraction and are presented in figure 8.44.
The evaporative fraction is constant with wind direction with the exception of a doubling of evaporative fraction for wind directions between 180 and 300 degrees. This pattern is inconsistent with the bimodal pattern of sensible heat flux fraction with reductions observed at 60 and 210 degree wind directions. To investigate possible causes, similar analysis was done for climate variables. The pattern of evaporative fraction corresponded to the directional pattern of $D$ (figure 8.45) which existed for all seasons of the year. This pattern of high $D$ for westerly winds corresponded to sunnier conditions, lower precipitation, and surprisingly, lower atmospheric pressure; there did not appear to be an associated directional relationship of temperature, (figure
It is suggested that air masses reaching the site from the west were dried as they travelled over the higher elevations in western Scotland. It is believed that $\lambda E$ is enhanced for wind flow along the rows as a result of greater penetration of wind into the canopy, but that the resulting bimodal pattern of evaporative fraction is masked by the pattern of $D$, which enhances the evaporative fraction peak at 230 degrees and diminishes the peak at 60 degrees. A more complex model of $\lambda E$ would be required to recreate the effects of canopy structure and water retention upon energy exchange. A further discussion of the directional effects upon energy budgets is presented in section 8.6.2.

8.5.5.2 Flux footprint maps

Figure 8-46 Cumulative flux footprint of latent heat flux by year.

The spatial distribution of observed sources of latent heat flux shown in figure 8.46 are more similar to the corresponding patterns of $R_n$ and $R_g$ (figure 8.28) than to that of sensible heat flux (figure 8.29). Relative to the footprints of $R_n$, there appears to be a slight preference towards larger values of $\lambda E$ from north-westerly wind directions. This preference is a result of site climatology through the influence of $D$ on $\lambda E$. As
shown in figure 8.45, higher vapour pressure deficits are associated with westerly wind directions throughout the year.

### 8.5.6 Water budget

Annual evapotranspiration losses were obtained by gap filling measured values of $\lambda E$ with modelled values using equation 8.16. When missing $\lambda E$ could not be modelled because of insufficient data, gaps were filled using monthly averaged diel values of $\lambda E$ (Falge et al. 2001). A comparison of gap filled measurements with purely modelled evapotranspiration indicates that modelled evapotranspiration is between 10 to 25% higher than gap filled measurements as shown in figure 8.47. The monthly differences in cumulative evapotranspiration are observed to be larger during the more humid autumn and winter months, consistent with the differences between model and measured $\lambda E$ observed in figure 8.43.

![Figure 8-47 Annual cumulative evapotranspiration. The solid line represents the modelled values determined using equation 8.25. The dotted line shows the measured data with gaps filled using the modelled values. The figure in the lower right corner is the average cumulative monthly difference between modelled and measured (gap filled) evapotranspiration.](image)

Despite differences between modelled and measured values, annual evapotranspiration losses were obtained employing the combined modelled and
measured values of evapotranspiration for each year. Annual evapotranspiration ranged between 402 and 463 mm. The smaller value in 2001 appears to have been the result of conditions resulting in lower evapotranspiration during the post summer period and not a result of the shorter data period.

### Table 8.9 Annual evapotranspiration for each of the five experiment years.

<table>
<thead>
<tr>
<th>Year</th>
<th>1997</th>
<th>1998</th>
<th>1999</th>
<th>2000</th>
<th>2001</th>
</tr>
</thead>
<tbody>
<tr>
<td>Annual evapotranspiration, mm</td>
<td>426</td>
<td>431</td>
<td>460</td>
<td>463</td>
<td>402</td>
</tr>
</tbody>
</table>

The annual curve of cumulative evapotranspiration losses showed a maximum rate of increase over the period of April through September, (figure 8.48). Because precipitation is nearly uniform throughout the year (figure 4.66) this higher rate of evapotranspiration is balanced by a reduction in stream flow losses over a similar period. Indeed, figure 8.48 suggests that nearly all precipitation falling between June and September is lost as evapotranspiration, with moisture also being extracted from the soil (figure 4.77)

![Graph showing cumulative stream flow and evapotranspiration for the last 8 months of the year 2000.](image)

**Figure 8-48** Cumulative stream flow and evapotranspiration for the last 8 months of the year 2000.

A comparison of cumulative ecosystem water gains (precipitation) and losses (evapotranspiration and stream flow) for periods in the year 2000 when all three component measures were available are shown in figure 8.49 indicates closure of the hydrologic budget to within 10%. Periods of missing data may have biased results because ecosystem storage capacity was not included in the budget. However, the largest gap, in July, corresponds to a period of small soil moisture change (see figure 4.77) thus minimizing this error. The better water budget closure, 3%, obtained by
Wilson et al. (2003) may have been associated with their more complete set of measurements. Because, hydrologic budget closure does not rule out significant errors in the accuracy of any of the budget components, the accuracy of the values of $\lambda E$ as a term in the energy budget cannot be verified, although this agreement does lend credence to their values.

![Cumulative water losses and gains for the last 8 months of 2000](image)

**Figure 8-49** Cumulative water losses (stream flow and evapotranspiration) and gains (precipitation) for the last 8 months of 2000. Only when all three contributing data were available were the losses and gains incremented.

### 8.6 Bowen Ratio and Energy Budget

#### 8.6.1 Bowen Ratio

The Bowen ratio, $\beta$, the ratio of sensible to latent heat, is controlled by those factors that affect evapotranspiration as described in section 8.5.3.2. This response can best be described using the Penman-Monteith model of $\lambda E$, equation 8.16, rearranged to describe $\beta$, in a format using resistances. This format is more accessible and will simplify explanation of the temporal and spatial patterns of $\beta$.

$$\beta = \frac{r_a + r_c - r_i}{\gamma r_a + r_i}$$

(8.27)

An example of the response of $\beta$ to environmental conditions can be exemplified using the average July diel curve of $\beta$ in relation to $D$ shown in figure 8.50. It is seen that, from sunrise to midday, $\beta$ remains constant at about 1.5 irrespective of the value of $D$. Because both $r_1$ and $r_c$ are proportional to the ratio of available energy to $D$, the constancy of the resistance terms and hence $\beta$ is a result of a near constant ratio of
available energy to $D$. After midday, $\beta$ decreases as $r_i$ and $r_c$ increase in response to available energy decreasing more rapidly than $D$. Examples of the ratios of $R_n$ to $D$ over diurnal and annual periods are also shown in figure 8.50, and should roughly predict the expected patterns for these periods, which are discussed further below.

Figure 8-50 Diel pattern of response of Bowen ratio (top left) and resistances (top right) to $D$. Data are averaged by percent of diel period, (note that in this figure percent of diel period ranges from 0% at 0:00 hrs to 100% at 24:00 hrs). Error bars represent ± one standard error, $n > 20$. The bottom two figures show the diel (for July), left, and annual (midday), right, patterns of the ratio of $R_n$ to $D$.

The diurnal average $\beta$ values, determined using $H$ and $H_t$ and shown in figure 8.51, increase rapidly after sunrise and reach a value of 1.5 by mid-morning and maintain that value until early afternoon. At about 70% of the diurnal period the value of $\beta$ begins a decline and becomes negative just before nightfall. The pattern of measured $\beta$ follows the predicted pattern of the ratio of $R_n$ to $D$ and falls midway between the corresponding patterns of $\beta$ obtained using equations 8.16 and 8.23. This again points to the overestimation of the model of $\lambda E$ used in this thesis and the underestimation of the Penman-Monteith estimation of $\lambda E$. 

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Fritschen et al. (1992) have observed a similar pattern for grassland. In some ecosystems, limitation of afternoon $\lambda E$ results in higher values of $\beta$ similar to those observed by Stewart and Thom (1973). Although the canopy conductance values, (figure 8.38), indicate a restriction of transpiration losses in the afternoon, the diel patterns of $\beta$ in figure 8.51 suggests that limitation of transpiration by the canopy is not sufficient to limit increased partitioning of available energy into $\lambda E$ in response to increasing $D$.

![Figure 8-51](image)

**Figure 8-51** Diurnal patterns of measured Bowen Ratio (filled and open circles) compared with scaled $R_{\nu}/D$ (gray curve), (left graph) and compared with model derived Bowen ratios (right graph) Data were averaged to bi-weekly values and error bars represent ± one standard error, $n > 1000$.

![Figure 8-52](image)

**Figure 8-52** Annual midday (30 To 70% of diurnal period) measured Bowen Ratio (filled circles) compared with scaled $R_{\nu}/D$ (gray curve), (left graph) and compared with model derived Bowen ratios (right graph) Data were averaged to bi-weekly values and error bars represent ± one standard error, $n > 4000$. 
Over the annual cycle, the midday values of $\beta$ increase to a maximum of about 1.5 before gradually decreasing to a near zero value in December, figure 8.52. The seasonal decrease in $\beta$ is again primarily the result of different annual patterns of available energy and $D$, similar to the diel patterns. This is reflected in the good agreement ($R^2$ 0.53) of the pattern of $\beta$ with the scaled ratio of $R_n$ to $D$. Patterns of model derived $\beta$ was similar ($R^2$ 0.49) but failed to capture the stronger variations at the end of winter while the pattern of Penman-Monteith estimates of $\beta$ did not adequately capture the observed variations ($R^2$ 0.09).

Variation of midday $\beta$ with wind direction in figure 8.53 have distinct minima for winds from 60 and 210 degrees; directions corresponding to the tree row orientation. The scaled pattern of the ratio of $R_n$ to $D$ agree closely with this pattern while the Penman-Monteith estimates of $\beta$ instead show a pattern that is more closely related to the directional pattern of $D$ while the pattern of model-derived $\beta$ is of intermediate agreement. This may suggest that equation 8.25 used for determining $\lambda E$ may require a canopy roughness factor in determining the balance between potential and Penman-Monteith estimates of $\lambda E$.

![Figure 8-53 Variation of average midday Bowen ratio (30 to 70% of diurnal period) with wind direction. Error bars represent ± one standard error, n > 300.](image)

This pattern of midday measured $\beta$ with wind direction suggests that penetration of wind into the canopy enhances $\lambda E$ and reduces $H$, which is not surprising if we consider that the lower canopy was often wet long after termination of precipitation.
Estimates of the effect of changes in zero plane displacement with wind direction upon $D$ indicate a 30 to 90% reduction for smaller values of $d$ (flow across rows) and thus corresponding increases in $\beta$. The implication of this observation is that orienting the planting rows parallel to the mean wind flow can enhance water loss if the canopy is subject to frequent precipitation.

8.6.2 Energy budget closure

The energy budget provides the most readily available verification of eddy covariance measures of energy exchange for an ecosystem. In the energy budget equation the amount of available energy $A = R_n - G - S$ must equal the amount of energy lost through the fluxes of latent and sensible heat. Ideally, there should be a one to one relationship of available energy, $A$, to $H + \lambda E$.

Figure 8.54 shows that the correction of fluxes resulted in improvement of energy budget closure from 64% for real-time fluxes to 91% for corrected fluxes. This is an improvement on the typical 80% closure for FluxNet sites reported by Wilson et al. (2002). Wilson et al. also observed poorer closure at night (35%) and when turbulent mixing was weak (low $u^*$), but noted no effects related to sensor type (open/closed path H$_2$O) or site characteristics. In contrast to Wilson (2002), we found both
instrumental and surface cover effects on energy budget closure. These topics will be discussed below.

Examination of the diel patterns of energy budget closure regressions (figure 8.55), reveals a pattern of almost constant slope (0.65) and intercept (0 W m\(^{-2}\)) during nocturnal periods and a gradual increase in slope (to 1.1) and decrease in intercept (to \(-10\ W\ m^{-2}\)) during the diurnal period. Although not shown, this diel pattern of energy budget closure is persistent over all seasons. Inspection of the individual regression curves indicates that the improvement in diurnal energy budget closure is not caused by the decrease in \(H+\lambda E\) at low fluxes of available energy but by larger values of \(H+\lambda E\) at high values of available energy.

![Figure 8-55](image)

**Figure 8-55** Diel pattern of energy budget closure slopes (centre panel), intercepts (bottom panel). Values of regression estimate root mean square error (RMSE) and correlation coefficient (\(R^2\)) values are given in the top panel.

The annual pattern of midday energy budget closure slope (figure 8.56) increases gradually from February (0.8) through September (1.0) followed by more rapid decrease to a minimum in January (0.55). The pattern of energy budget closure intercept remains relatively constant throughout the year with minima observed in
January, October, and December. As with the diel cycles, the annual change in intercept is small and unlikely to be the cause of the improvement in closure observed between spring and autumn.

![Diagram showing annual pattern of energy budget closures slopes and intercepts](image)

**Figure 8-56** Annual pattern of energy budget closures slopes (top graph) and intercepts (bottom graph) using sonic (closed circles) and thermocouple (open circles) estimate of sensible heat flux. Values are obtained from regressions of half hourly values grouped by month.

When grouped by wind direction, figure 8.57, energy budget closure is best for wind directions of approximately 100 and 210 degrees. The improvement in closure for south-westerly directions agrees with the observations of reduced $H$ and enhanced $\lambda E$ for wind flow along rows while that for 100 degrees is slightly more southerly than the along-row flow. Energy budget closure intercepts also show variation with wind direction, with more negative values for winds from south-westerly directions and more positive values for winds from 130 and 300 degrees. It appears that the pattern of intercepts is not closely in-phase with the pattern of energy budget closure.

There are several possible explanations for lack of energy budget closure. Perhaps the simplest would result from an error in $R_n$ through improper calibration or levelling. Calibration error can be ruled out because it would produce a constant closure error that would not change in response to other variables. A levelling error would produce a change with time of day but would not show the observed wind direction
dependency. The possibility of errors associated with energy storage terms exists but cannot be addressed without further research into the spatial and temporal variability of energy storage components.

A second source of error may be the assumption that spectra correspond to an average sensor height ($z-d$). It has been shown that $z-d$ changes with wind direction from 9 m for flow across rows to 13 m for flow along rows. Because height was assumed to be 9 m for along-row flow but velocity will correspond to 13 metres, this assumption will introduce a shift in spectra to lower frequencies for along-row winds. In fluxes with significant low frequency attenuation this may result in corrections failing to account for low frequency attenuation, resulting in the observed underestimate for along-row flow. This does not, however, explain diurnal and annual patterns, as there is no expected preference for along-row flow earlier in each day or earlier in each year.

![Figure 8-57 Relation of the energy budget closure regression slope (top panel) and intercept (bottom panel) to wind direction using sonic (filled circles) and thermocouple (open circles) sensible heat flux.](image)

Specific to measurements of $\lambda E$, the change in relative humidity over the day may result in larger morning and early season signal attenuation by the processes described in section 5.3.2.7. However, the unimodal changes of $D$ with wind direction (see
figure 8.45) do not show the required bimodal behaviour required to explain the pattern of energy budget closure seen in figure 8.57.

Inadequate corrections for sonic angle of attack errors may change with wind direction and canopy structure but there is no reason to expect their behaviour to vary regularly over the day or over the year. Further investigation is needed to determine the dependency of this correction on environmental conditions.

![Diagram showing variation of run mean vertical velocity with wind direction. Error bars represent ± one standard error, n > 300.]

Figure 8-58 Variation of run mean vertical velocity with wind direction. Run mean vertical velocities represent the residual value after planar fit coordinate rotation. Error bars represent ± one standard error, n > 300.

The use of planar fit coordinate rotation for this site provides another possibility for introducing errors in measurements of both $H$ and $\lambda E$. An analysis of the residual mean vertical velocities (figure 8.58) indicates a negative residual mean vertical velocity for winds from 60 and 220 degrees and positive residuals for winds from 140 and 340 degrees. Unfortunately, these residuals would suggest a negative bias in fluxes associated with along-row flows and a positive bias for across-row flow. This would enhance the directional behaviour of energy budget closure observed in figure 8.58.

Friction velocity has also been observed as a controlling factor of energy budget closure (Turnipseed et al. 2002; Wilson et al. 2002) and may be associated with the assumption of negligible advective flux components in equation 5.1.
Figure 8-59 Response of energy budget closure regression parameters slope (left panels), intercept (center panels), and R² (right panels) for day (filled circles) and night (open circles). All five years data are given in the top panels and separated by 60-degree wind direction bins in lower panels. Regressions presented have $n > 100$. 

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In general energy budget closure has been observed to be worse at low friction velocities, particularly at night. The pattern of energy budget closure in response to $u^*$ shown in figure 8.59 indicates a diurnal linear increase from about 0.8 at low $u^*$ to near 1.0 at high $u^*$, while the nocturnal pattern has a maximum ($\sim 0.8$) at $u^* \sim 0.3$ m s$^{-1}$ and decreasing energy budget closure at higher and lower $u^*$. The decrease in energy budget closure at higher $u^*$ at night seems inconsistent with the daytime observations, although great reliability can not be placed on these values if we consider the very low $R^2$ values associated with these conditions, as shown in figure 8.59 (right panels).

These patterns are almost identical to those observed by Turnipseed et al. (2002) and Aubinet et al. (2000), although their peak in nocturnal energy budget closure occurred at a higher friction velocity ($\sim 0.8$ m s$^{-1}$). The effect of topography on energy budget closure has been mixed; Wilson et al. (2002) found no conclusive effect of topography, Turnipseed et al. (2002) found better closure with down slope flow and Humphreys et al. (2003) found better closure with upslope flow. For this study energy budget closure at low $u^*$ was similar regardless of wind direction. At higher $u^*$ we are in agreement with Turnipseed et al. (2002) and find better closure for down slope flows. However, as shown in figure 8.59 (centre panels), some of this effect appears to be associated with a negative bias in $H + \lambda E$ for down slope flow and positive bias for up slope flow. Initial impressions were that this directional response was inconsistent with the bimodal pattern that appeared to be associated with canopy architecture. However, in the following section it is shown that similar topographic influences exist for the fluxes of carbon dioxide.

In conclusion, it appears that there may not be a single cause for the lack of energy budget closure. Instead, instrumental errors, canopy architecture, site topography and other as yet unexplained causes all are contributing to our inability to obtain accurate measurements under all conditions. It is apparent that further in-depth analysis of the possible causes outlined above is needed.
8.7 Carbon dioxide flux

The net assimilation of carbon dioxide by an ecosystem, $F_N$, can notionally be separated into components describing the gross photosynthetic assimilation, $F_G$, and respiration, $R$. The respiration component can be further separated into heterotrophic ($R_h$) and autotrophic ($R_a$) components.

\[ F_c \approx F_N = F_G + R_a + R_h \]  

(8.28)

At Griffin, $F_N$ was estimated using the CO$_2$ flux, $F_c$, determined from eddy covariance measurements, as described in Chapter 5. The following sections describe the patterns of $F_c$ in relation to time, space and relevant controlling variables. Modelling of $F_c$ in response to environmental conditions is done for the purpose of gap-filling in order to obtain annual estimates of carbon uptake.

8.7.1 Correction effects

Uncorrected values of carbon dioxide flux obtained as output from the data collection program, Edisol, compared with final corrected values are shown in figure 8.60 and regression coefficients given in table 8.10. The correction methods are discussed in detail in Chapters 5, 6, and 7. The comparison indicates an approximate 30\% increase in flux resulting from corrections. Slight increases in correction over the five years may be a result of a spectral shift to lower frequencies as described in section 8.5.1. The observed increase indicates that the low frequency losses are larger than the high frequency gains for $F_c$; this is appropriate if high frequency losses are initially small (see Chapter 7). The RMS error of the regression of corrected with uncorrected $F_c$ was about 5\% of the entire range of observed values, intermediate to the corrections of $H$ and $\lambda E$. 

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Figure 8.60 The effect of correction on carbon dioxide flux for the year 1998. Uncorrected values are those obtained directly from the Edisol data collection software while corrected values are the results of the procedures described in Chapter 5. The solid line is the linear regression for the 1998 data and the dashed line is the 1:1 line.

Table 8.10 Linear regression of corrected against uncorrected carbon dioxide flux, n > 10,000 for all years.

<table>
<thead>
<tr>
<th>Year</th>
<th>Intercept</th>
<th>Slope</th>
<th>RMSE</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1997</td>
<td>0.41</td>
<td>1.23</td>
<td>1.71</td>
<td>0.94</td>
</tr>
<tr>
<td>1998</td>
<td>0.33</td>
<td>1.25</td>
<td>1.93</td>
<td>0.94</td>
</tr>
<tr>
<td>1999</td>
<td>0.35</td>
<td>1.33</td>
<td>2.03</td>
<td>0.95</td>
</tr>
<tr>
<td>2000</td>
<td>0.42</td>
<td>1.34</td>
<td>2.65</td>
<td>0.91</td>
</tr>
<tr>
<td>2001</td>
<td>0.31</td>
<td>1.35</td>
<td>2.09</td>
<td>0.94</td>
</tr>
</tbody>
</table>

8.7.2 Temporal patterns

The annual patterns of monthly average \( F_c \) shown in figure 8.61, range from a small November/December release of +1 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) to a maximum uptake of -5 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) during the month of May. In contrast, annual patterns of monthly average \( R \), have a slightly later, January/February minimum release of 0.5 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) and a maximum release of 7 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) in August. Although \( F_G \) was not measured
directly; it was estimated, $F_{cg}$, as the difference between $F_c$ and $R$. The resulting pattern of monthly average $F_{cg}$, shown in figure 8.61, reaches a maximum over the period of May to August. Because $F_c$ is a combination of the processes of photosynthesis and respiration, each of which is controlled by several variables, no single climate variable can explain the pattern of $F_c$.

The post mid-year peak in $R$ corresponds to the annual pattern of monthly average temperature, (see figure 4.40). There are no patterns of relevant variables that correspond well with the annual pattern of monthly average $F_c$. The pattern of $F_{cg}$ is similar to that observed for $\lambda E$, suggesting that the controls on $F_{cg}$ are similar to those controlling transpiration, (Tuzet et al. 2003). A more detailed examination of the environmental controls of $F_c$ and $R$ will be made in following sections.
The relationship of the annual pattern of carbon uptake to the growth of the canopy provided a further, interesting observation. The period of net uptake from February through September seen in figure 8.62 differs from the May through October period of canopy growth deduced from drag coefficient observations (see section 8.3.3.3). This difference suggests that springtime (February through April) carbon uptake may be supporting root growth while aboveground growth may be benefiting from carbon uptake later in the season. This is in agreement with observations of early season root growth (Law et al. 1999) followed by development of aboveground biomass starting in May and extending into October (McWilliam 1972).

![Figure 8-62 Annual pattern of monthly mean carbon dioxide net assimilation ($F_c$), respiration ($R$) and gross assimilation ($F_{eq}$). All five years of gap-filled data were used. Diurnal values of $R$ were obtained as described in section 8.7.2.3. Error bars indicate ±1 standard error with $n > 5000$.](image)

Seasonal analysis of the diel patterns of $F_c$ shown in figure 8.63 suggests that $F_c$ declines later in the afternoon during the period from January through April than it does during May through December. This delayed decline indicates either increased photosynthesis or decreased respiration in the afternoon during January through April as a result of lower temperature or higher radiation in the afternoon during winter.
Figure 8-63 Diel pattern of carbon dioxide flux for bi-monthly periods. Error bars represent ± one standard error, n > 400.

Figure 8-64 Normalized diurnal patterns of $F_c$, and associated variables for bi-monthly periods plotted against percent of diurnal period. Curves were normalized by subtracting the minimum value of the monthly averaged curve and dividing by its range.
The normalized diel patterns presented in figure 8.64 demonstrate the shift to higher afternoon of both $F_c$ and relevant variables during January-April. Because radiation is a primary driving variable, the observed shift in $Q_{pg}$ suggests that the pattern of radiation may be controlling the observed patterns in $F_c$ and other variables. The cause of the afternoon increase in winter $Q_{pg}$ is not known. A closer analysis of environmental factors is done below.

The date/time contour plot of $F_c$ shown in figure 8.65, however, does not have the large pre mid-year fluxes notable in the monthly averaged plots in figure 8.61. Instead, both diurnal and nocturnal $F_c$ are higher after mid-season with the highest nocturnal $F_c$ occurring prior to midnight. We know from the annual plot that nocturnal fluxes have larger values after midyear, resulting in the pattern of monthly averaged fluxes observed in figure 8.62.

![Figure 8-65 Date/Time contour plot of carbon dioxide flux.](image)

**8.7.3 Environmental relations**

Analysis of carbon dioxide fluxes by regression of the deviations of the annual monthly means against those for potential controlling variables provided a distinct breakdown of factors possibly affecting inter-annual variability (table 8.11). (A similar analysis using annual deviations from the 5-year mean gave different results but was considered less reliable because of the low number of data points used to form the regression.)
Table 8.11 Ranked correlation coefficients for regression of the deviation of monthly average $F_c$ from five-year mean monthly $F_c$ against various environmental variables. The columns labelled Radiation, Humidity, and Other are visual assistants to ease identification of variable groupings. The column ‘Correl’ specifies the sign of the correlation of the deviations.

<table>
<thead>
<tr>
<th>Variable</th>
<th>$R^2$</th>
<th>Correl</th>
<th>Radiation</th>
<th>Humidity</th>
<th>Other</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_g$</td>
<td>0.46</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pa}$</td>
<td>0.38</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pg}$</td>
<td>0.33</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_t$</td>
<td>0.29</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pt}$</td>
<td>0.23</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.20</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$p_g$</td>
<td>0.16</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$e$</td>
<td>0.15</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$h_f$</td>
<td>0.13</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$D$</td>
<td>0.11</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$g_c$</td>
<td>0.07</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T_s$</td>
<td>0.06</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$T_a$</td>
<td>0.03</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pt}/Q_{pg}$</td>
<td>0.02</td>
<td>-</td>
<td>X</td>
<td></td>
<td>X</td>
</tr>
<tr>
<td>Albedo</td>
<td>0.01</td>
<td>-</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{pf}$</td>
<td>0.00</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$U$</td>
<td>0.00</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$%R_p$</td>
<td>0.00</td>
<td>+</td>
<td>X</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

As expected, radiation is the main determinant of the inter-annual variability of $F_c$, with increases in radiation causing higher carbon uptake (i.e., a negative correlation). The next most strongly related variables, precipitation and other humidity measures suggest that higher humidity is associated with smaller flux. These positive correlations are likely a result of negative correlations of humidity and radiation. Temperature ($T_a$, and $T_s$) is also positively correlated with $F_c$, although this correlation cannot be explained by a negative correlation of temperature and radiation. It is more likely that the strong dependence of respiration on temperature is the cause of this correlation (see section 8.7.2.3). The weak correlations of interannual variability of other variables with that of $F_c$ does not necessarily suggest that they have no effect on $F_c$, but more likely that their effects occur over time scales smaller than the monthly averaging periods used for this analysis.
It is also interesting to note that inter-annual variability was not constant throughout
the year. Variability was smaller during the springtime period of increase in $F_c$ than
during the period of more gradual decline from May to December. To determine if
this were true, five-year monthly coefficients of variation were calculated for $F_c$ and
$Q_{pg}$ (figure 8.66). Ignoring the first and last two months of the year because of the
generally small values of the fluxes, we still see that the variation of $F_c$ increased
relative to that of $Q_{pg}$ over the course of the year.

![Figure 8-66 Comparison of mean monthly values of $F_c$ and $Q_{g}$ (left panel)
and normalized variability (right panel). Mean monthly values of $F_c$ were
modified to obtain a zero offset corresponding to that observed for $Q_{g}$.]

### 8.7.3.1 Relationship to stability

The relationship of turbulent intensity of carbon dioxide to Monin-Obukhov stability
shown in figure 8.67 is nearly identical to that observed for water vapour in section
8.5.3.1, showing only a slightly more rapid fall off as neutrality is approached from
unstable conditions. Notably different for $\sigma_c/C*$ is its magnitude at near neutral
conditions; similar to $\sigma_v/Q*$ the magnitude is lower than predicted. As neutral
conditions are approached $\sigma_c/C*$ tends towards a constant value of approximately 5, a
value similar to that observed for $\lambda E$ but half that observed for temperature.

As with water vapour, some of the observed discrepancy from predicted values may
result from using uncorrected values of $\sigma_c$. An examination of the negative bias in $F_c$
caused by katabatic flow was not analyzed, but should have caused underestimation
for unstable conditions and an improvement for stable conditions. However, it is not
clear what effect such a bias would have on values of $\sigma_c$. Once again, a definitive
explanation of the behaviour of turbulent intensities is not readily available.

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Figure 8-67 Average carbon dioxide turbulent intensities as a function of stability. Dashed lines represent model relations given in equation 8.13, and error bars represent one standard error. $C$ and the standard deviations of carbon dioxide were not corrected for frequency response or nocturnal flux loss in this calculation. Error bars represent ± one standard error, $n > 100$.

### 8.7.3.2 Environmental control of nocturnal flux

Because the variables controlling $R$ and $F_{cg}$ are different, determination of these variables is approached by separation of data into diurnal and nocturnal periods. Nocturnal flux analysis is covered first, followed by analysis of diurnal fluxes.

The primary controls of respiration, and hence nocturnal $F_c$, have been identified by previous research as temperature, soil moisture, soil substrate, canopy photosynthesis and limitations to transport (Fang and Moncrieff, 1999; Buchmann 2000; Hogberg et al. 2001; Hu et al. 2001; Law et al. 2001; Jassel et al. 2004). No experimental research was carried out during this experiment to quantify potential substrate limitations resulting from substrate quality or photosynthesis. Addition of a term for cumulative radiation for the preceding day, as a proxy for photosynthesis, did not improve the respiration models described below and therefore analysis focused on the effects of temperature, soil moisture and atmospheric transport.

Because distinct measurements of component respiration were not available, the strong effects of temperature upon both autotrophic and heterotrophic respiration components were assumed similar in response (Moncrieff and Fang, 1999; Boone et al. 1998; Lloyd & Taylor 1994a) and were exploited using the available temperature
measurements. Respiration was attributed roughly to soil and canopy respiration contributions— for which measurements of temperature were available. To assess the relationship of respiration to temperature, two common models of the response of respiration to temperature were employed. An exponential relationship,

$$ R = R_0 \cdot e^{bT} \quad (8.29), $$

contains an empirical exponent $b$ and the parameter $R_0$ corresponding to respiration at a temperature of $0 \, ^\circ C$. The so-called $Q_{10}$ parameter is defined as the increase in respiration rate in response to a $10 \, ^\circ C$ temperature increase and is given by $Q_{10} = e^{(10b)}$. An Arrhenius type model proposed by Lloyd and Taylor (Lloyd & Taylor 1994b) was also implemented. This model accounts also for the decrease in the temperature sensitivity of respiration as temperature increases (i.e., the decline in $Q_{10}$).

$$ R = R_{10} e^{\frac{E_0}{283.15-T} - \frac{1}{T-T_0}} \quad (8.30) $$

In this model, the coefficient $R_{10}$ represents the respiration rate at $10 \, ^\circ C$, $E_0 (= 308.56 \, K)$ and is similar to an activation energy, and $T_0 (= 227.13 \, K)$ is a reference temperature. In this model, temperature, $T$, is in degrees Kelvin. An alternative version of this model was tested in which the right hand side of equation 8.30 was repeated with one term representing soil respiration based on soil temperature, $T_s$, and the other term representing canopy and soil surface respiration and based on air temperature, $T_a$ (Swanson & Flanagan 2001).

$$ R = R_{10s} e^{\frac{E_0}{283.15-T_s} - \frac{1}{T_s-T_0}} + R_{10a} e^{\frac{E_0}{283.15-T_a} - \frac{1}{T_a-T_0}} \quad (8.31) $$

In this formulation the two $R_{10}$ coefficients implicitly include the proportions of respiring biomass of the two sources.

To account for effects of soil moisture on respiration, a Gaussian curve was incorporated into each of the above three formulations. This formulation,
was applied as a gain function and was intended to account for both the reduction in soil gas transport and oxygen limitation at high soil moisture and for limitation at low soil moisture content. In this model, the coefficient $c$ is an empirical factor describing the broadness of the response curve while the coefficient $e_{v0}$ describes the volumetric soil moisture at which its effects are minimal. For the two-source Lloyd and Taylor model (equation 8.30) the soil moisture function was applied only to the soil temperature term of the equation.

\[
e^{-0.5 \left( \frac{v - e_{v0}}{c} \right)^2}
\]

(8.32)

---

The data fitted to the models given above were taken from all five years of $F_c$ data, and were limited to periods when solar elevation angle was larger than 0 degrees and $u_*$ was higher than 0.6 m s$^{-1}$. For models based on only one source, an averaged temperature value was employed ($T = T_x / 3 + 2T_y / 3$) derived from previous examination of the data. Because soil moisture was only available for the last year of
the experiment, the preceding four years were approximated using one-week average $O$, obtained from the final years data.

The response to soil moisture of the model coefficients of equations 8.29, 8.30 and 8.31 are shown in figure 8.68. The temperature range associated with each soil moisture class used to determine these coefficients was near or higher than 10 °C. The covariance of annual patterns of temperature and soil moisture suggests that effects of soil moisture (and other annually varying factors) could be concealed within temperature response.

From figure 8.68 it is apparent that maximum soil respiration occurs at a soil moisture fraction of about 0.35 for the one-source models, and at a soil moisture fraction of about 0.4 for the two-source Lloyd and Taylor model. The decrease in $R_{10}$ of biomass with increasing soil moisture may be a result of strong effects of soil surface respiration because of the temperature similarities. The behaviour of the exponential model coefficients suggests that the response of the $R_0$ and $b$ coefficients to soil moisture is not independent. As soil moisture is more likely to affect the magnitude of respiration than its sensitivity to temperature, the soil moisture response function was applied only to the $R_0$ coefficient. The pattern $R_0$ with soil moisture also suggests a peak at a soil moisture fraction of ~0.38.

Unfortunately prevailing high soil moisture fractions makes the extrapolation to drier conditions in figure 8.68 inappropriate. Respiration values estimated from the exponential model coefficient fits of soil moisture did not give reasonable values at low soil moisture, while estimates of respiration from the fit of the Lloyd and Taylor $R_{10}$ coefficients were more reasonable.

Better model fitting was obtained using non-linear regression methods (SAS, NLIN); the resulting coefficients are presented in table 8.12. Although both $R_{10}$ coefficients of the two-source model responded to soil moisture, difficulties in obtaining convergence with the non-linear fitting process meant that a soil moisture function could only be applied to the $R_{10}$ term of equation 8.31.
Both models, fitted using non-linear regression techniques, explained about 60% of the variation in respiration when using only temperature, and this increased to about 65% with the inclusion of soil moisture. The exponential model produced an annual $Q_{10}$ value of 4.7 that decreased to 3.0 with the addition of a soil moisture component. This latter value of $Q_{10}$ is comparable to the range of values found by chamber measurements of respiration at sites of similar climate (Berbigier et al. 2001; Buchmann 2000; Moren & Lindroth 2000; Widen 2002). The fit of the two-source, soil moisture version of the Lloyd and Taylor model did not improve the fit in comparison with the single-source, soil moisture model; however, the proportion of $R_{10s}$ to $R_{10b}$ of the two source model did support the proportions used in averaging soil and air temperatures in the single source model. Examination of the $R^2$ values of the model fits suggest that the exponential model with a soil moisture component was slightly better than the other model versions.

### Table 8.12 Coefficients of respiration model.

<table>
<thead>
<tr>
<th></th>
<th>Exponential</th>
<th>Exponential w/ soil moisture</th>
<th>Lloyd &amp; Taylor</th>
<th>Lloyd &amp; Taylor w/ soil moisture</th>
<th>Lloyd &amp; Taylor two source w/ soil moisture</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_{10}$</td>
<td></td>
<td>5.77±0.029</td>
<td>7.08±0.065</td>
<td>4.203±0.108 $T_s$</td>
<td>2.79±0.096 $T_a$</td>
</tr>
<tr>
<td>$A$</td>
<td>1.36±0.018</td>
<td>2.336±0.055</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$B$</td>
<td>0.154±0.001</td>
<td>0.110±0.002</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$Q_{10}$</td>
<td>4.7</td>
<td></td>
<td>3.00</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\theta_0$</td>
<td></td>
<td>0.343±0.007</td>
<td>0.334±0.008</td>
<td>0.363±0.004</td>
<td></td>
</tr>
<tr>
<td>$C$</td>
<td></td>
<td>0.186±0.007</td>
<td>0.194±0.008</td>
<td>0.106±0.005</td>
<td></td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.59</td>
<td>0.66</td>
<td>0.60</td>
<td>0.65</td>
<td>0.64</td>
</tr>
</tbody>
</table>

The addition of soil moisture provided the largest improvement in model fit for both the exponential (7%) and Lloyd and Taylor models (5%). The model fits indicate a 50% reduction in respiration for soil moisture outside the range of $\theta_s=0.1$ to 0.55 for the single-source models and $\theta_s=0.25$ to 0.5 for the two-source model. The upper end of this range is similar to observations by several researchers (Elberling 2003; Moncrieff & Fang 1999; Subke et al. 2003), while the lower end is higher than other observed values, $\theta_s \sim 0.05$ (Elberling 2003; Moncrieff et al. 1999).
Figure 8.69 shows the soil moisture gain functions associated with equation 8.29, 8.30 and 8.31 and the frequency distributions of soil moisture both for averaged values and for ridge and unploughed surfaces. When applied to average soil moisture, the soil moisture gain function of the single-source soil respiration model indicates that less than 10% of measurements fell under soil moisture conditions ($\theta_s < 0.10, \theta_s > 0.55$) that would have restricted respiration by more than 50%. However, as shown in figure 8.69, when soil moisture is presented as two soil moisture regimes with non-overlapping distributions, the broader range of moistures has a correspondingly higher probability of restriction by either high or low soil moistures. Examination of such effects requires spatially sampling of soil respiration measurements to define the underlying processes, and information for scaling up these non-linear processes. As figure 8.69 suggests, improvement in estimates of $R$ could be obtained by including spatial variability of soil moisture.

An example of the Lloyd and Taylor, single-source model fitted to nocturnal flux data is presented in figure 8.70. There remains considerable variability unaccounted for by the model; a more complex model, which addressed this variability, was beyond the scope of this experiment. As such, the single-source Lloyd and Taylor model with soil moisture term was employed in the final analyses.
An additional adjustment to this model to account for flux behaviour at low friction velocities, the “nocturnal $u_*$” problem, is described in the following section.

Figure 8-70 Relationship of nocturnal measured (open circles) and modelled (closed circles) ecosystem carbon dioxide flux to soil temperature. Included data were limited to conditions of $u_*$ greater than 0.5 m s$^{-1}$. Model data were obtained using the Lloyd and Taylor single temperature model with soil moisture.

8.7.3.3 Nocturnal flux wind speed dependence

A commonly observed phenomenon associated with eddy covariance measurements of nocturnal carbon dioxide flux is the so-called ‘$u_*$ problem’. It has been observed (Aubinet et al. 2000; Carrara et al. 2003; Goulden et al. 1996; Knohl et al. 2003) that, at night, under conditions of low friction velocity, values of $F_c$ are less than expected, based on scaled-up surface chamber measurements and $F_c$-temperature relationships developed under higher wind speed conditions. A similar analysis has been done using the Griffin data. However, instead of removing and replacing nocturnal data occurring under some threshold $u_*$ condition, a model was created for correcting $F_c$ values as a function of friction velocity. This approach was taken for two reasons.
First, the nature of the empirical soil respiration model used precludes it from capturing all of the natural variability of the true soil respiration. A second reason is that $u^*$ related flux loss may not be limited to values below a specific $u^*$ threshold. The correction method employed was intended to correct fluxes over wider range of friction velocities as well as retaining any variability of nocturnal $F_c$ not accounted for by the respiration model.

![Diagram](image)

**Figure 8.71** Nocturnal relationship of log of carbon dioxide flux and air temperature to friction velocity. Five years of data averaged by 0.033 m s$^{-1}$ $u^*$ bins.

To investigate the possibility of flux loss, it has been assumed that the relationship described by equation 8.30, and developed using data when $u^* > 0.5$ m s$^{-1}$, is appropriate at low values of $u^*$. With this assumption, plots of log($F_c$) and $T_a$ against friction velocity should have nearly identical patterns. Figure 8.71 shows the discrepancy in the patterns of $T$ and log($F_c$) for $u^*$ values below 0.6 m s$^{-1}$. A comparison between air temperature, soil temperature and averaged air and soil temperature, and a comparison between storage corrected and uncorrected $F_c$ values is also shown to demonstrate the discrepancy that could arise from errors associated
with temperature input into the modelled \( R \) or corrections to \( F_c \) values. It is clear that these errors are small compared to the discrepancy related to \( u^* \). For many sites (Aubinet \textit{et al.} 2000), large apparent underestimates of nocturnal \( F_c \) have been reported for friction velocities below a \( u^* \) threshold ranging from 0.3 to 0.6 m s\(^{-1}\).

To quantify this underestimate, the difference between the curves of nocturnal \( F_c \) estimated using equation 8.30 and measured values were averaged for 0.033 m s\(^{-1}\) friction velocity bins (figure 8.72). The resulting underestimations of nocturnal \( F_c \) follow a skewed distribution with small effects at friction velocities of \( u^* \sim 0 \) m s\(^{-1}\) and \( u^* \sim 0.7 \) m s\(^{-1}\). At high \( u^* \) the trend in discrepancies continues to negative values, suggesting that the effect may continue to affect fluxes at even higher velocities. That the underestimate achieves a zero value at \( u^* \sim 0.7 \) is probably an artefact of the selection of conditions for which the modelled \( R \) was developed. Based on the pattern shown in figure 8.72, it is suggested that flux loss is occurring under all \( u^* \) conditions but is maximal at \( u^* \sim 0.1 \) m s\(^{-1}\) and minimal at very high and very low friction velocities.

![Figure 8-72 Underestimate of nocturnal \( F_c \) as a function of \( u^* \). The magnitude of the underestimate was determined as the difference between measured and modelled (equation 8.30) nocturnal \( F_c \). Error bars represent \( \pm \) one standard error, \( n > 20 \).](image)

Correction for nocturnal \( F_c \) loss is not justified unless the cause of the loss is known and understood. Recalling the evidence given in section 4.12.2, we know that katabatic flow exists at the Griffin site. It is likely that katabatic drainage is only
partially responsible for nocturnal flux loss, as it is not suspected to play a significant role over the wide range of friction velocities indicated in figure 8.72 (Doran 1991; Mahrt et al. 2001).

An alternative cause of nocturnal $F_c$ loss may be advective fluxes (see section 5.2) induced by the complex topography of the Griffin site (Aubinet et al. 2003; Staebler & Fitzjarrald 2004; Turnipseed et al. 2003). If present, such advective flows are likely to affect both fluxes and their associated CO$_2$ concentration profiles. The vertical profiles of CO$_2$ concentration in a forest, on a horizontally homogeneous site, should be similar under similar environmental conditions and should not be function of wind direction. To test this assumption for Griffin, the differences in the shape of the CO$_2$ concentration profiles were determined for up-slope and down-slope flows. Profiles (relative to above canopy concentration) were averaged for ranges of environmental conditions and the averaged profiles for upslope and downslope flow were differenced to determine the effect of topography on the profile.

![Figure 8-73 Change in average difference between above and within canopy CO$_2$ concentration for down slope and upslope flow directions.](image)

The differences, averaged for a range of radiation and friction velocity regimes is shown in figure 8.73. The resulting profile differences indicate that down-slope flows have higher within canopy concentrations (1 to 4 ppm) than up-slope flows. Such observations are consistent with previous observations of canopy temperature profiles in section 4.12.2, and raise the possibility that topographically induced advection may be affecting the fluxes. Further investigation and modelling are needed to determine
if such advection could be responsible for the observed within canopy profile
behaviour and distribution of flux losses as a function of friction velocity.

The possibility of topographic effects on nocturnal flux losses suggests the potential
for wind direction dependence. To analyse for this effect, we recall that both
respiration and wind direction (see figure 4.64) vary seasonally. Separation of flux
underestimates by both season and wind direction, as shown in figure 8.74, provided
verification that wind direction effects were occurring throughout the year. Because
the wind direction pattern is difficult to observe from these $F_c$ vs $u^*$ curves, the peak
underestimates at low $u^*$ were selected for both seasonal (figure 8.74) and annual (not
shown) curves and plotted against wind direction in figure 8.75. In this figure a
directional dependence is observed, with a minimum flux loss for westerly directions
and a maximum for easterly directions. The curve for annual peak underestimation by
wind direction shown in figure 8.75 follows a similar pattern.

An adjustment was also made to the annual $F_c$ vs $u^*$ curves to account for the
variability in $F_c$ at $u^* > 0.5$ m s$^{-1}$, like that seen in figure 8.74. The annual curves for
each wind direction were shifted by an amount that minimized the offset of $F_c$ underestimate at $u^* > 0.5 \text{ m s}^{-1}$. The peak underestimates of $F_c$ discrepancy at low $u^*$ were then determined and are also shown in figure 8.75. These adjusted values indicate a slight exaggeration of the existing annual pattern of peak flux underestimation as a function of wind direction.

The sine wave fitted to the unadjusted annual average peak flux loss by wind direction shown in figure 8.75 indicates that the minimum flux loss occurs for winds from 275 degrees and maximum flux loss for winds from 95 degrees. These directions correspond to the primary slope of the valley at Griffin (see figure 4.62). It is seen from figures 8.73 and 8.75 that down-slope flows correspond to larger flux underestimates and higher within canopy concentrations while, for up-slope flow, flux underestimates are halved and within canopy concentrations are lower.

![Figure 8-75 Relation of peak nocturnal flux underestimate to wind direction for seasonal (top panel) and annual (bottom panel) averages. The annual patterns of underestimation peaks are shown for both unadjusted underestimation curves and for the same curves adjusted for flux loss at high friction velocities. A sine wave has been fitted to the unadjusted annual peak values.](image)
The patterns of nocturnal flux loss, in particular the wide range of $u^*$ at which it may be occurring, suggest the correction should be applied to all nocturnal $F_c$ data. In the unlikely event that $F_c$ underestimation is purely the result of katabatic flow than it should be restricted to nocturnal and early morning periods, whereas if it is associated with anabatic flow or topography-induced advection it may need to be applied to all periods, a topic which will be discussed in the following section. Additionally, if flux loss is caused by katabatic flow and/or advection, it may be necessary to correct all fluxes and not just $F_c$, as has been suggested by other research findings (Turnipseed et al. 2002; Wilson et al. 2002).

An empirical correction for flux underestimation, $\Delta R$, was developed using the product of three functions. One function describes the peak flux underestimation at low friction velocities, $f(R)$, a second defines the variation of flux underestimation with friction velocity, $g(u^*)$, and the third gives its variation with wind direction, $h(\theta)$, i.e.,

$$\Delta R = f(R) g(u^*) h(\theta)$$

Figure 8.76 Annual pattern of monthly average measured (open circles) and modelled (closed circles) nocturnal $F_c$. The peak in the relationship of flux loss to $u^*$ for each month is also shown (black triangles). Curves were obtained using all five years nocturnal data.

The peak losses, $f(R)$, follow a seasonal trend with maximum rates in summer and minimum rates in winter, in correspondence with the seasonal pattern of respiration, as shown in figure 8.76. The magnitude of the peak flux underestimation was
therefore assumed to be a constant proportion of the true flux. As a surrogate, it was defined as a proportion (48%) of the respiration flux modelled using equation 8.30,

\[ f(R) = R_{\text{peak}} = 0.48 \cdot R_{\text{model}} \]  

(8.35)

The response of flux underestimation to friction velocity was modelled using a Gaussian distribution in which the parameter \( u_{*0} \) is the friction velocity at which peak flux underestimation occurs and \( c \) is a parameter defining the broadness of the underestimation curve,

\[ g(u_*) = \left[ 1 + \left( \frac{u_* - u_{*0}}{c} \right)^2 \right]^{-1} \] 

(8.36)

The wind direction dependence function was defined as a sine wave fitted to the data presented in figure 8.75. The coefficient \( b \) is the amplitude of the wind direction adjustment and the coefficient \( \theta_0 \) determines the directional dependence of the adjustment.

\[ h(\theta) = b_r \cdot \sin(\theta + \theta_0) \] 

(8.37)

These three functions defined by equations 8.35, 8.36, and 8.37 are combined into a single function (equation 8.38 and table 8.13), which is added to the measured respiration to account (empirically) for the observed effects of flux underestimation.

\[ \Delta R = R_{\text{peak}} \left[ 1 + b_r \cdot \sin(\theta + \theta_0) \right] \left[ 1 + \left( \frac{u_* - u_{*0}}{c} \right)^2 \right]^{-1} \] 

(8.38)

<table>
<thead>
<tr>
<th>( b_r )</th>
<th>( c_r )</th>
<th>( \theta_0 )</th>
<th>( u_{*0} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.344 ± 0.054</td>
<td>0.086 ± 0.014</td>
<td>-6.4 ± 16.9</td>
<td>0.095 ± 0.008</td>
</tr>
</tbody>
</table>
While equation 8.38 empirically addresses the "nocturnal u*" problem, because of the data used in its development, it is limited to application at night. It remains a question as to whether it, or a similar correction needs to be applied during the day. As a test of this hypothesis, the peak in flux underestimation, and measured values of flux loss were determined as a function of nocturnal period to assess their values near sunrise and sunset, as shown in figure 8.77. It is apparent that neither the peak of the flux loss u* relationship, or the measured flux loss tend to zero at the nocturnal boundaries, suggesting that the flux loss is likely to be occurring during at least a portion of the day near sunrise and sunset. This possibility will be explored further in the following section.

The temporal pattern of flux loss over the nocturnal period shown in figure 8.77 (right panel) indicates that flux loss increases from a minimum at sunset to a maximum at sunrise. The modelled flux loss also captures this pattern but does not capture the magnitude of this pattern, probably because of covariance of driving variables (R_{model}, u*, and \theta) over time. It is possible that the decrease in u* over the night could account for this increase. However, the increase in peak flux loss with respect to u* over the night (figure 8.77, left panel) suggests that R_{model} is also increasing over the night. This is counter-intuitive because R_{model} is determined by temperature, which is known to decrease over the night. The only factor of equation 8.38 that might counteract the
expected decline in peak flux loss is wind direction. From figure 4.63 it is recalled that wind direction has a tendency to change to down slope flow at night under low wind speeds (<2 m s\(^{-1}\)). However, this shift is present throughout the night and does not occur gradually over the nocturnal period as would be required to explain the behaviour shown in figure 8.77.

As mentioned, the discrepancy between modelled and measured flux loss (figure 8.77 right panel) indicates an inadequacy of the model of nocturnal flux loss. The lack of a proper explanation of such a pattern may also suggest that friction velocity may not be the appropriate controlling variable for explaining the behaviour of nocturnal flux loss. A more in depth investigation of the effects of katabatic flow and topographically induced flow patterns are needed to identify more appropriate scaling variables. One possible variable may be the Froude number (John Finnigan, pers. comm.), but further investigation into its relevance is beyond the scope of this thesis.

8.7.3.4 Environmental control of diurnal flux

The net diurnal uptake of carbon by an ecosystem is a balance between losses incurred through respiration and gains obtained by photosynthesis. Although soil respiration may be related to photosynthesis rates (Hogberg et al. 2001; Tang et al. 2003), it was assumed that diel variability of the underlying processes of and controls on respiration were well described in the preceding section, and that other diel variability was insignificant (Buchmann 2000; Rayment 1998).

Although photosynthesis is driven by radiation, it may be limited by either the flux density of radiation (electron transport limited) or by the availability of photosynthetic substrate (Rubisco activity limited). At low to moderate radiation flux density, the assimilation of carbon dioxide during photosynthesis is approximately linearly related to absorbed PPFD (Hall & Rao 1999). This linear relation tends to zero at a small positive radiation flux density because of the offsetting effect of maintenance respiration. This offset is referred to as the light compensation point with respect to CO\(_2\), analogous offset may also be expressed as a CO\(_2\) compensation concentration, \(\Gamma^*\), the concentration at which respiration is balanced by photosynthesis, (Jones 1992). Although PPFD is the main control of assimilation at low radiation flux density, both sub-stomatal CO\(_2\) concentration and temperature may affect the photosynthetic rate
under these conditions. The effect of sub-stomatal CO₂ concentration appears directly in the equation describing electron transport-limited photosynthesis while the effect of temperature appears in the equation through $\Gamma$. At high PPFD, a plant's ability to maintain levels of photosynthetic substrate (RuBP) limits photosynthesis. Under these conditions sub-stomatal CO₂ concentration and temperature are the primary limiting variables with sub-stomatal CO₂ concentration again appearing directly in the equation describing Rubisco-limited photosynthesis, while temperature affects $\Gamma$ as well as the maximum rate of Rubisco formation and the Michaelis-Menten rate constants. The availability of other photosynthetic substrates (e.g., nitrogen, phosphate) may also have an effect on Rubisco-limited photosynthesis, but were not considered in this study.

**Conductance model**

Although photosynthesis is a chemical process that plants cannot actively control, plants can affect the rate of photosynthesis by modifying sub-stomatal CO₂ concentration and leaf temperature through control of stomatal aperture. Variation in stomatal aperture determines the stomatal conductance, which describes the rate at which CO₂ can be transported from the ambient air to the sub-stomatal cavity and the rate at which water vapour may be transpired and transported away from the sub-stomatal cavity. The stomatal conductance to CO₂ has been defined by Ball *et al.* (1987) and modified by Leuning (1995) as

$$g_{sc} = g_o + \frac{a \cdot F_{cg}}{(C_c - \Gamma)(1 + \frac{D}{D_o})}$$

and is used here to represent the relationship of canopy stomatal conductance, $g_{sc}$, to canopy photosynthesis, $F_{cg}$. The term $g_o$ is a constant defining the canopy stomatal conductance as PPFD and $F_{cg}$ approach zero, $C_c$ is the leaf surface CO₂ concentration, $D$ is the vapour pressure deficit, and $a$ and $D_o$ are empirical constants. The $\Gamma$ parameter is again the CO₂ concentration that has been estimated using a temperature dependent polynomial following Brooks and Farquhar (1985), with leaf temperature approximated by air temperature:
\( \Gamma = 44.7 + 1.88(T_a - 25) + 0.036(T_a - 25)^2 \) \hspace{1cm} (8.40)

Soil moisture deficit or excess may also cause reductions in stomatal conductance (Turner 1991) by affecting plant water status (Beadle et al. 1979; Tuzet et al. 2003). However, neither severe drought or water logging of the soils at Griffin occurred during the period of the experiment, and soil moisture is thought, in general, not to limit growth of Sitka spruce in Great Britain (Waring 2000).

Practical problems limit the application of equation 8.39 for determination of canopy gross CO\(_2\) uptake. Most importantly, this equation actually defines the behaviour of a leaf under uniform conditions whereas canopies have complex radiation, CO\(_2\), and humidity environments. Some models attempt to replicate the complexity of this environment (Amthor et al. 2001; Wang & Jarvis 1990) while others assume the canopy behaves as a single leaf with a uniform environment (Amthor 1994; Lloyd et al. 1995).

While equation 8.39 defines the relationship between assimilation, canopy conductance, CO\(_2\) concentration and humidity deficit, the definition of the constant parameters require independent measures of assimilation and canopy conductance. For Griffin these constants were obtained using estimated \( F_{cg} \) and canopy conductance (see section 8.5.3.2) converted to molar conductance of CO\(_2\).

| Table 8.14 Parameter constants for the Ball-Berry-Leuning stomatal conductance model, equation 8.39. Error values are the standard error of the estimate, \( n = 21325 \). |
|---|---|---|---|---|
|  | \( g_0 \) \hspace{1cm} ( \mu\text{mol m}^{-2} \text{s}^{-1} ) | \( a \) \hspace{1cm} ( dimensionless ) | \( D_0 \) \hspace{1cm} ( kPa ) | \( R^2 \) |
| This study | 0.277 ± 0.001 | 16.7 ± 0.2 | 0.195 ± 0.005 | 0.65 |
| Leuning (1995) | 0.01 to 0.11 | 1.5 to 8.8 | 0.7 to 1.5 | |
| Wingate (2003) | 0.038 | 3.5 | 2.9 | |
| Rayment (1998) | 0.0065 | 1.54 | 6.8 | |

The parameter constants were obtained using non-linear fitting methods (SAS NLIN), with data limited to conditions of \( D > 0.05 \text{ kPa} \) and \( \lambda E > 10 \text{ W m}^{-2} \); the resulting
constants are presented in Table 8.14. The values of \( a \) and \( g_0 \) are higher than those observed by Leuning (1995) while the value of \( D_0 \) is lower. The larger values of \( a \) and \( g_0 \) may reflect the inclusion of evaporation as well as transpiration in the values of \( g_{sc} \) obtained via inversion of the Penman-Montieth equation.

**Radiation response model**

Missing values of \( F_{cg} \) were estimated using knowledge of the behaviour of the light-limited and substrate-limited behaviour of photosynthesis. From descriptions of photosynthesis we know that canopy assimilation consists of two near-linear processes. A plot of \( F_{cg} \) against \( Q_{pg} \) should appear as a linear response of \( F_{cg} \) to \( Q_{pg} \) at low values of \( Q_{pg} \) while at high \( Q_{pg} \) the value of \( F_{cg} \) is constant. However, the complex distribution of radiation, humidity, plant water status, \( \text{CO}_2 \), and nitrogen throughout the canopy result in a smoother transition between light-limited and substrate-limited regimes. The bimonthly response of \( F_{cg} \) to \( Q_{pg} \) at Griffin shown in Figure 8.78 suggests that the canopy is predominantly PPFD limited at \( Q_{pg} \) of less than about 250 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) and \( \text{CO}_2 \) limited above 1000 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) while at intermediate levels of \( Q_{pg} \) a combination of PPFD and \( \text{CO}_2 \) limitation is occurring.

![Figure 8-78 Response of \( F_{cg} \) to \( Q_{pg} \) for bi-monthly periods. Curves include all five years of Griffin data. Error bars represent ± one standard error, \( n > 20 \).](image)

The resulting response of \( F_{cg} \) to \( Q_{pg} \) has commonly been approximated with a hyperbolic curve (Thomley 1976). Two forms of the hyperbolic curve may be employed, both of which include the constants of maximum assimilation rate, \( A_{\text{max}} \).
and the quantum yield, \( \varepsilon \) (\( \Delta F_{\text{eg}} / \Delta Q_{\text{pg}} \) at a \( Q_{\text{pg}} \) just above zero). A non-rectangular hyperbola, equation 8.41, includes an additional parameter, \( \vartheta \), that defines the smoothness of the transition between light-limited and substrate-limited regimes, (Thornley 1976).

\[
F_{\text{eg}} = \frac{\varepsilon \cdot Q_{\text{pg}} + A_{\text{max}}}{\sqrt{\left(\varepsilon \cdot Q_{\text{pg}} + A_{\text{max}}\right)^2 - 4 \cdot \vartheta \cdot \varepsilon \cdot Q_{\text{pg}} \cdot A_{\text{max}}}} - 2 \cdot \vartheta
\]  

(8.41)

A rectangular hyperbola, equation 8.42, may be obtained from equation 8.41 by assuming \( \vartheta = 0 \).

\[
F_{\text{eg}} = \frac{A_{\text{max}} \cdot \varepsilon \cdot Q_{\text{pg}}}{A_{\text{max}} + \varepsilon \cdot Q_{\text{pg}}}
\]  

(8.42)

The rectangular hyperbola is commonly used to define the relationship of \( F_{\text{eg}} \) to \( Q_{\text{pg}} \); however, if modifications of the equation constants are required to account for environmental effects, the rectangular hyperbola suffer from an interdependence of \( A_{\text{max}} \) and \( \varepsilon \) that is not present in the non-rectangular hyperbola. A comparison of these two PPFD response equations fitted to the data is presented in table 8.15, and figure 8.79.

<table>
<thead>
<tr>
<th>Method</th>
<th>( \varepsilon )</th>
<th>( A_{\text{max}} )</th>
<th>( \vartheta )</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Non-rectangular hyperbola</td>
<td>0.058 ± 0.001</td>
<td>29.9 ± 0.3</td>
<td>0.62 ± 0.02</td>
<td>0.81</td>
</tr>
<tr>
<td>Rectangular hyperbola</td>
<td>0.070 ± 0.001</td>
<td>36.1 ± 0.3</td>
<td></td>
<td>0.80</td>
</tr>
</tbody>
</table>

Table 8.15 Coefficients for rectangular and non-rectangular curves fitted to all data \( n > 50000 \). Coefficient variability values are RMSE of coefficient estimates.

The good comparison of the two methods in autumn and winter is likely the result of a lack of high radiation during those periods - resulting in the inability of both methods to obtain an accurate value of \( A_{\text{max}} \). In contrast, the data in table 8.15 and spring and summer data from figure 8.79 indicate that the rectangular hyperbola gives 20% higher values of \( A_{\text{max}} \) and \( \varepsilon \) than those from the non-rectangular hyperbola. This difference suggests that the rectangular hyperbola overestimates fluxes at high PPFD and underestimates them at lower levels. To check this possibility the residuals of the
fitted models were averaged by $Q_{pg}$ bins and compared in figure 8.80. The residuals show that the rectangular hyperbola overestimates values of $F_{cg}$ for $Q_{pg}$ between 50 and 300 μmol m$^{-2}$ s$^{-1}$ and underestimates them for $Q_{pg}$ between 300 to 700 μmol m$^{-2}$ s$^{-1}$. This will cause errors in $F_{cg}$ of up to ±1 μmol m$^{-2}$ s$^{-1}$ but will have a small net effect unless the frequency distribution of $Q_{pg}$ for the missing values of $F_{cg}$ is different from the frequency distribution of $Q_{pg}$ for the data used to derive the parameters in the rectangular hyperbola equation.

![Graph showing transition parameter, quantum efficiency, and maximum assimilation over months]

**Figure 8.79** Coefficients for rectangular (filled circles) and non-rectangular (open circles) hyperbola fitted to all data separated by month. Error bars represent ±1 standard error (n > 2700).

Figure 8.79, for the non-rectangular hyperbola in spring and summer, shows that $\mathcal{G}$ remains roughly the same throughout the year, at a value of ~0.75, while $A_{\text{max}}$ appears to rise gradually whilst $\varepsilon$ increases after midyear. The pattern of $\varepsilon$ may be the result of development of new leaves and the resulting higher mesophyll conductance after expansion of new shoots in June (Ludlow & Jarvis 1971). The cause of the rise in $A_{\text{max}}$ in relation to relevant environmental variables will be assessed in the following section through analysis of the non-rectangular $F_{cg} - Q_{pg}$ response curve.
Using the concepts of Montieth (1997) the slope of weekly averaged values of $F_c$ vs $Q_{pg}$ is $-0.0155 \, \mu$mol/\mu mol that, assuming a 48% carbon content of biomass, would give an radiation conversion efficiency of 3.2%. However, examining the relationship of monthly average $F_c$ and $Q_{pg}$ shown in figure 8.81 we can see a seasonal hysteresis in the relationship. The hysteresis of $F_c$ is the result of a stronger seasonal hysteresis in $R$ than that of $F_{eg}$. The seasonal hysteresis of $R$ is the result of the phase shift of the
annual pattern of temperature. The seasonal hysteresis of $F_{ci}$ is likely a result of the seasonal acclimation of quantum efficiency seen in figure 8.79.

**Model parameters: Environmental controls in radiation response**

Initial inspection of $A_{\text{max}}$, $\varepsilon$, and $\theta$ in relation to $T_a$, $D$, $g_c$, and percent $R_p$ showed smaller effects associated with $g_c$ and $D$ and larger effects from $T_a$ and percent $R_p$. Given existing evidence of temperature effects on $A_{\text{max}}$ (Johnson & Thornley 1985; Medlyn et al. 2002), a temperature response function to was fitted to $A_{\text{max}}$ before examining the effects of other environmental variables. A beta function, given by equations 8.43 and 8.44, was fitted to the values of $A_{\text{max}}$ derived by fitting a non-rectangular hyperbola to data separated by two-degree temperature classes. The beta function is given as:

\[
A_{\text{max}} = A'_{\text{max}} \frac{(T_a - T_L)(T_H - T_a)^\theta}{(T_o - T_L)(T_H - T_o)^\theta} \tag{8.43}
\]

and

\[
\theta = \frac{T_H - T_o}{T_o - T_L} \tag{8.44}
\]

where $A'_{\text{max}}$ is the maximum expected value of $A_{\text{max}}$, $T_L$ and $T_H$ are the low and high temperatures at which $A_{\text{max}}$ goes to zero and $T_o$ is the temperature at which $A_{\text{max}}$ reaches its optimum value. The measured values of $A_{\text{max}}$ and resulting model are shown in figure 8.82. Parameter values of $T_L$, $T_o$ and $T_H$ obtained by least squares optimisation of equation 8.43 are $-3.7$, $12.2$, and $25.7 ^\circ C$, respectively. An attempt to obtain values of all parameters ($A_{\text{max}}$, $\varepsilon$, $\theta$, $T_L$, $T_o$, and $T_H$) coincidentally using non-linear methods gave similar results although unreasonably high values of $T_H$ were obtained and the value of $\theta$ was close to zero. By setting the value of $T_H$ to a predetermined value of $45 ^\circ C$, values of $T_L$ and $T_o$ of $-4.1$, $14.7$ were obtained as well as reasonable values of $\varepsilon$, $A_{\text{max}}$, and $\theta$ (i.e., $0.069$, $29.9 \mu $mol m$^{-2}$ s$^{-1}$, and $0.69$ respectively). The corresponding values for $T_L$, $T_o$ and $T_H$ of $-4.1$, $14.7$ and $45 ^\circ C$, were employed.

The value of $T_L$ is similar to the $-6 ^\circ C$ observed by Neilson et al. (1972) for Sitka, while the value of $T_o$ falls near their value of $18 ^\circ C$ (range of $10$ to $22 ^\circ C$). The value
of \( T_H \) was selected to correspond to their value of 45 °C. It was also noted by Neilson et al. (1972) that mesophyll resistance was the primary cause of the \( A_{\text{max}} \) response of Sitka over the range of 5 to 25 °C and that stomatal conductance was relatively constant over this temperature range. This is consistent with the initial observations of insensitivity to \( g_c \) of the non-rectangular hyperbola parameters.

![Graph](image)

**Figure 8-82** Variation of \( A_{\text{max}} \) parameter of the non-rectangular hyperbola with air temperature (filled circles). The solid line represents the beta function fitted to the response.

Inclusion of the temperature response of \( A_{\text{max}} \) described by equation 8.43 in the non-linear fitting of equation 8.41 resulted in an increase of the \( \theta \) parameter while other parameters remained similar, as given in table 8.16.

To examine the potential for effects of two other possible controlling factors, examples of the \( F_{cg}/Q_{pg} \) response curves were fitted to all data, fitted to data with \( D > 0.9 \) kPa and fitted to data with percent \( R_p \) between 70% and 90% as shown in figure 8.83. Relative to the curve fitted to all data, the curve obtained using data with \( D > 0.9 \) kPa shows a reduction in \( F_{cg} \) at all \( Q_{pg} \), and is likely an effect of stomatal closure as indicated by equation 8.39. In contrast, higher percent \( R_p \) causes a variable effect, with a reduction in \( F_{cg} \) at low to medium levels of \( Q_{pg} \) and enhanced \( F_{cg} \) at high \( Q_{pg} \). Similar flattening of the \( F_{cg}/Q_{pg} \) curve has been observed by Law et al. (2002) and may indicate the effect of more unilateral illumination of the canopy (Leverenz & Jarvis 2002). However, Law et al. (2002) did not observe a similar enhancement of \( F_{cg} \) at high \( Q_{pg} \).
Table 8.16 Non-rectangular hyperbola and beta function coefficients. Value of \( T_H \) was assumed. Values of \( T_0 \) and \( T_L \) were obtained by fitting \( A_{\text{max}} \) of non-rectangular hyperbola to 2 °C binned data. The coefficients were obtained by fitting a non-rectangular hyperbola to all data using predetermined beta function coefficients.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \varepsilon )</td>
<td>( 0.059 \pm 0.001 )</td>
</tr>
<tr>
<td>( A_{\text{max}} )</td>
<td>( 28.6 \pm 0.2 \ \mu \text{mol m}^{-2} \text{s}^{-1} )</td>
</tr>
<tr>
<td>( \vartheta )</td>
<td>( 0.80 \pm 0.01 )</td>
</tr>
<tr>
<td>( T_L )</td>
<td>(-4.1 \ ^\circ \text{C})</td>
</tr>
<tr>
<td>( T_0 )</td>
<td>(14.7 \ ^\circ \text{C})</td>
</tr>
<tr>
<td>( T_H )</td>
<td>(45 \ ^\circ \text{C})</td>
</tr>
<tr>
<td>( R^2 )</td>
<td>0.78</td>
</tr>
</tbody>
</table>

Figure 8-83 Comparison of non-rectangular hyperbola fitted to all data (filled circle) with those fitted to data with \( D > 0.9 \ \text{kPa} \) (open circle) and to data with percent \( R_p \) between 70% and 90% (filled triangle).

At similarly high \( Q_{pg} \) values, \( F_{cg} \) under cloudier conditions is enhanced relative to \( F_{cg} \) under clear sky conditions because of a relative reduction in uptake of shade leaves (Roderick et al. 2001). While a high percent \( R_p \) does not necessarily imply clear sky conditions, figure 8.84 shows that a large proportion of runs with high percent \( R_p \) also have large standard deviations of \( R_g \), suggesting that cloud-reflected radiation under partly cloudy conditions may be creating an environment of combined...
direct and strong diffuse radiation which may be responsible for the enhanced $F_{cg}$ at high percent $R_p$ and high $Q_{pg}$.

![Graph showing comparison of variability of $R_g$ with percent $R_p$.](image)

**Figure 8-84** Comparison of variability of $R_g$ with percent $R_p$.

**Diurnal flux loss**

Based on the observations of a nocturnal flux loss relationship to $u^*$ and $u^*$-related lack of energy budget closure (see section 8.6.2), it was considered prudent to examine the possibility of similar losses during the day. This possibility was checked by plotting the difference between modelled and measured values of $F_e$ grouped by values of $u^*$ in figure 8.85. Three models were used in this analysis: the non-rectangular hyperbola described by equation 8.41, the SPA model (Williams *et al.* 2001) and for the Maestra model (Medlyn *et al.* 1999; Wang *et al.* 1990). The rectangular hyperbola was parameterised using all available data while Maestra was parameterized using only Griffin data from 1998. The SPA model was parameterized using alternate data sets. The somewhat larger differences observed between the measured data and SPA model are likely caused by the use of this parameterization.

The differences between modelled and measured diurnal $F_e$ were averaged by $u^*$-groupings. The shape of the resulting curves shown in figure 8.85 are similar to those observed for nocturnal fluxes in figure 8.72. The negative values at low $u^*$ indicate that measured fluxes are smaller than modelled fluxes, suggesting that flux magnitudes are reduced at low $u^*$ values during diurnal as well as nocturnal periods.
To see if diurnal $F_c$ underestimation had a directional dependence, the response of midday modelled $F_c$ residuals (using the non-rectangular hyperbola) were separated by 60 degree wind direction bins, and minimized assuming zero residuals at $u^- > 0.4$ m s$^{-1}$. The resulting curves are shown in figure 8.86 and values of the curve peak underestimates are shown in figure 8.87 along with the nocturnal peak underestimates by wind direction taken from figure 8.75. The curves in figure 8.87 indicate that the effect on flux underestimation of wind direction related processes is different from the $u^*$ related processes. While the $u^*$ related processes are acting as gain functions to reduce the magnitude of $F_c$, the wind direction effects appear to be causing an additive adjustment of $F_c$. The directional (i.e., topographic) effect acts to make $F_c$ more positive for down slope flows ($\sim 120$ deg) and more negative for upslope flows ($\sim 300$ deg).

![Figure 8-85 Difference between measured and modelled diurnal $F_c$ as a function of $u^-$. Curves are presented for three models, a non-rectangular hyperbola fitted to the data (filled circle), the SPA model (open circle) and the Maestra model (no symbol). Error bars represent $\pm$ one standard error, $n > 250$.](image)

To better understand the transition from nocturnal to diurnal $F_c$ underestimations, the difference between non-rectangular hyperbola modelled and measured $F_c$ was examined by percent of day $\Pi_d$. The change in modelled $F_c$ residuals as a function of $u^*$, and grouped by $\Pi_d$, is shown in figure 8.88. The curves show a positive residual at low $u^*$ during the first and last 20% of the day and a negative residual during...
midday periods, as shown in figure 8.87. From figure 8.63 we can see that the pattern of underestimation roughly corresponds to the pattern of diurnal $F_c$. Such a response during the day is also consistent with equation 8.38 describing nocturnal underestimation as being proportional to flux magnitude, suggesting that the underlying process is acting as a gain function modification of the flux.

Figure 8-86 Difference between measured and modelled diurnal $F_c$ as a function of $u_*$ and wind direction. Error bars represent ± one standard error, n > 20.

Figure 8-87 Difference between measured and non-rectangular hyperbola modelled $F_c$. Average residuals are plotted against $\Pi_d$ in the left panel. In the right panel peak values of the error-$u_*$ curves are plotted in relation to wind direction. Both plots assume that $u_*$ related errors are minimal at $u_* > 0.4$ m s$^{-1}$. 

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It is hypothesised that a gain error, perhaps a sampling error loss associated with low frequencies, is causing flux loss under conditions represented by low $u_*$ values. The additional error associated with wind direction may be caused by topography-induced horizontal advection and results from the assumption of no horizontal advection made in obtaining the reduced flux equation 5.11. Further research is needed to identify if these separate sources of error are justifiable, easily identifiable, and quantifiable.

Figure 8-88 Difference between measured and modelled diurnal $F_c$ as a function of $u_*$ and fraction of diurnal period, $\Pi_d$. Error bars represent ± one standard error, $n > 20$.

To account for the pattern of observed flux loss, equation 8.38 was replaced with a modified version to take into account the bias character of the wind direction error. By assuming that the causes of the errors were similar during both day and night the other functions describing the correction and their associated parameters ($b, c, \theta_0, u_{*0}$) were not changed.

$$\Delta F_c^* = \left[ F_{c\text{-peak}} + F_{c\text{-peak}} \cdot b \cdot \sin(\theta + \theta_0) \right] \cdot \frac{l + \left( \frac{u_* - u_{*0}}{c} \right)^2}{1}$$  \hspace{1cm} (8.45)

In equation 8.45 the peak in $F_c$ loss was assumed to be similarly proportional to the peak in nocturnal flux loss (i.e. equation 8.35), and employed the modelled value of gross photosynthesis and respiration.

$$F_{c\text{-peak}} = 0.48 \cdot \left( F_{ig} + R_{model} \right)$$  \hspace{1cm} (8.46)
The flux adjustment obtained from equation 8.45 was applied to both diurnal and nocturnal fluxes. Similar to the nocturnal flux loss analysis, the diurnal loss analysis was applied as a correction to retain the natural variability of the measurements that may not have been explained by a model and also because the analysis indicates that there may not be a threshold $u^*$ above which flux values are "good". The effect of this application on estimates of ecosystem carbon uptake will be described below.

### 8.7.3.5 Spatial relations

The average annual cumulative flux footprints shown in figure 8.89 indicate that the region to the northwest of the tower acts as the primary sink for CO$_2$ while the region to the southeast is a weaker sink and a source at larger distances from the tower. This pattern is more similar to that observed for $\lambda E$ than that of $H$. In general the footprint lies within the region of interest, although the main footprint to the NW is dissected by a 10 m wide, grass-covered forest ride. No separate information concerning this ride was available and it was assumed that it did not significantly bias the site-averaged fluxes of carbon dioxide.

![Figure 8-89 Cumulative flux footprint of carbon dioxide flux by year for the Griffin experiment site.](image-url)
8.7.4 Carbon budget

The effects of variations in gap filling and \( u^* \) correction methods were examined by comparing carbon exchanges, \( A_N \). Four variants of annual cumulative \( A_N \) are compared; all four variants are based on measured fluxes with nocturnal gaps filled using a soil moisture adjusted Lloyd and Taylor model (Lloyd et al. 1994b). For compatibility with many other research estimates, missing diurnal fluxes in the first variant were gap filled with a rectangular hyperbola and corrected for nocturnal \( u^* \) errors (equation 8.38). The second variant also applied \( u^* \) correction only at night but used the non-rectangular hyperbola for gap filling of diurnal fluxes. The third and fourth variants were gap filled using the non-rectangular hyperbola and did not apply \( u^* \) corrections for the third variant but applied a diel \( u^* \) correction using equation 8.45 for the fourth variant. The gap filled values of \( F_c \) were multiplied by the appropriate molar density of air to obtain values of \( A_N \). A comparison of the cumulative carbon uptake for Griffin for each year of the experiment using the four estimates is given in table 8.17.

Table 8.17 Annual and site average values of \( A_N \). Values of \( A_N \) were gap filled using either a non-rectangular hyperbola model (equation 8.41) or rectangular hyperbola (equation 8.42) and corrected for \( u^* \) related errors using either equation 8.38 or 8.45 or not corrected, as specified. All values are in units of g m\(^{-2}\) per annum. Error values of the averages are one standard deviation of the mean (n=5).

<table>
<thead>
<tr>
<th>Year</th>
<th>Rectangular hyperbola (equ. 8.40)</th>
<th>Non-rectangular hyperbola (equ. 8.39)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Nocturnal ( u^* ) correction (equ. 8.34)</td>
<td>Nocturnal ( u^* ) correction (equ. 8.34)</td>
</tr>
<tr>
<td>1997</td>
<td>-626</td>
<td>-623</td>
</tr>
<tr>
<td>1998</td>
<td>-610</td>
<td>-595</td>
</tr>
<tr>
<td>1999</td>
<td>-768</td>
<td>-751</td>
</tr>
<tr>
<td>2000</td>
<td>-684</td>
<td>-680</td>
</tr>
<tr>
<td>2001</td>
<td>-714</td>
<td>-684</td>
</tr>
</tbody>
</table>

The differences between rectangular and non-rectangular hyperbola annual estimates of \( A_N \) vary by between 3 and 30 \( \mu \)mol m\(^{-2}\) a\(^{-1}\), with an average of 13 \( \mu \)mol m\(^{-2}\) a\(^{-1}\).
The values from the rectangular hyperbola are always larger than those from the non-rectangular hyperbola, a difference that arises from the slightly larger number of missing values of $F_c$ at low values of $Q_{pg}$ and smaller number of missing $F_c$ at higher $Q_{pg}$, as shown in figure 8.90, combined with the overestimation of rectangular hyperbola modelled $F_c$ at low $Q_{pg}$ and underestimation at higher $Q_{pg}$ as shown in figure 8.90. This two percent difference is, however, small in comparison to the differences arising from variations in $u^*$ related corrections.

![Figure 8-90 Frequency distribution of available and missing values of $F_c$ in relation to $Q_{pg}$.

The differences between annual cumulative $A_N$ corrected for nocturnal $u^*$ errors with annual estimates corrected using either no $u^*$ correction or using $u^*$ corrections over the entire diel period vary between 40 and 100 $\mu$mol m$^{-2}$ a$^{-1}$, with the nocturnal $u^*$ corrected $A_N$ values always being smaller. If $u^*$ related errors are indeed occurring during the day, this result suggests that application of only the nocturnal aspect of this correction will result in an 8% underestimation of the annual value of $A_N$. Although applying no $u^*$-correction is better than applying only a nocturnal correction, for this site, such a tactic would result in a 4% overestimation of $A_N$. The results of this thesis lead to the conclusion that $u^*$ related corrections should be applied to all data and not just to nocturnal data; however, for the purpose of comparison, results for both diel and nocturnal $u^*$ corrected data will be carried through the remainder of this thesis. Clearly, the significant effect of $u^*$ related corrections justifies further research.

Examination of the interannual variability of gap filled $A_N$, gross carbon exchange, $A_G$, and carbon respiration, $A_R$, are compared in Table 8.18. It is seen that $u^*$ correction methods had negligible effect on interannual variability expressed as
coefficient of variation, CV. More interannual variability exists in $A_G$ (3.7%) than in $R$ (2.4%). Although the variability of $A_N$ is proportionally larger (9.1%), the CV value of $A_N$ is not a reliable estimate because $A_N$ is not a zero based value (i.e. it can have either positive or negative values).

It should be noted that these values are subject to other variation depending upon subtle differences in the methods of data processing and analysis. Although it is not possible to present all the possible permutations of such calculation procedures within this thesis, it is the experience of the author that variations on the order of $\pm 100$ g m$^{-2}$, in addition to those caused by $u^*$ corrections or gap filling methods, can be expected from such an exercise. It is suggested that such an analysis should be carried out and should include variations in assumptions, methods, and modelling at all stages. However, for the purposes of this thesis the values obtained by the methods employed will be considered correct.

Table 8.18 Annual and site average values of $R$, $A_N$, and $A_G$. Values of $A_N$ were gap filled using a non-rectangular hyperbola model (equation 8.41) and nocturnal or diel $u^*$ flux loss (equation 8.38 and 8.45 respectively). Values of $A_G$ were determined as the difference between $A_N$ and $R$. All values are in units of g m$^{-2}$ per annum. Error values of the averages are one standard deviation of the mean (n=5).

<table>
<thead>
<tr>
<th>Year</th>
<th>$R$</th>
<th>Nocturnal $u^*$ correction (equ. 8.34)</th>
<th>Diel $u^*$ correction (equ. 8.43)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$A_N$</td>
<td>$A_G$</td>
<td>$A_N$</td>
</tr>
<tr>
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<td>-2174</td>
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<td>1513</td>
<td>-680</td>
<td>-2183</td>
</tr>
<tr>
<td>2001</td>
<td>1488</td>
<td>-684</td>
<td>-2164</td>
</tr>
<tr>
<td>Avg.</td>
<td>1532±37</td>
<td>-667±61</td>
<td>-2198±81</td>
</tr>
<tr>
<td>C.V.</td>
<td>0.024</td>
<td>0.091</td>
<td>0.037</td>
</tr>
</tbody>
</table>

The average annual $A_N$ at Griffin was either $-667$ g m$^{-2}$ when corrected for nocturnal $u^*$ flux loss, or $-721$ g m$^{-2}$ if corrected for diel $u^*$ flux loss. Although validation of the annual $A_N$ estimate by independent measures was not possible, checks for
reasonableness were done using closure residuals of the carbon budget. The values of $A_N$ obtained by eddy covariance and gap filled using empirical modelling were compared with the biomass and soil carbon accumulation values described in Chapter 2.

Although no extensive effort was made to determine the components necessary to close the carbon budget, enough measurements were available to obtain a rough estimate of the partitioning of carbon. Given the average $A_N$ obtained from gap-filled eddy covariance measurements (this Chapter), average aboveground biomass accumulation obtained by biometric measurements over the period of the experiment (Chapter 2), and average annual soil carbon accumulation since planting (Chapter 2) it was possible to estimate the amount of annual root biomass accumulation, Table 8.19.

<table>
<thead>
<tr>
<th>Component</th>
<th>Carbon accumulation using nocturnal $u^*$ correction g m$^{-2}$ a$^{-1}$</th>
<th>Carbon accumulation using diel $u^*$ correction g m$^{-2}$ a$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_N$</td>
<td>-667</td>
<td>-721</td>
</tr>
<tr>
<td>Aboveground biomass increase</td>
<td>421</td>
<td>421</td>
</tr>
<tr>
<td>Soil carbon increase</td>
<td>111</td>
<td>111</td>
</tr>
<tr>
<td>Root biomass increase by difference</td>
<td>135</td>
<td>189</td>
</tr>
<tr>
<td>Root : shoot ratio</td>
<td>32%</td>
<td>45%</td>
</tr>
<tr>
<td>Proportion of C below ground</td>
<td>24%</td>
<td>31%</td>
</tr>
</tbody>
</table>

The cumulative annual $A_N$ for Griffin compared with values for other sites in table 8.20 shows the relatively high values and low variability of $A_N$ at Griffin. The higher values and small variability of $A_N$ at the Griffin site is to be expected because of the moderating influence of the maritime climatology of the British Isles. Such influences may also be present at other sites (eg., Brasschaat, Le Bray, Loobos, Vielsam), although the actual climatological effects are likely to be complicated combinations of different factors beyond the analytical scope of this thesis. The values of the ratio of annual average carbon accumulation of root to shoot obtained from the two values in table 8.19 (32% and 45%) are larger than root to shoot estimates (27% to 37%) observed for a range of environments and species.
(Cairns et al. 1997; Chang et al. 1996; Mund et al. 2002), but within the range of root to shoot ratios values (41 ± 16%) obtained for Sitka spruce in the United Kingdom by Levy et al. (2004). Unfortunately, because of the nature of this experiment it was not possible to obtain true root to shoot ratios to compare with these other experiments.

Following Carrara et al. (2003), variability in $A_N$ at Griffin was compared to that of other sites as the ratio of the maximum observed annual deviation from the mean, normalized by the site average $A_N$ and expressed as a percentage. From Table 8.20, it is seen that the average inter-annual variability, based on this small sample of years, is approximately 13% of $A_N$. This value was lowest for the Griffin site and was generally lower for coniferous species, and sites with higher average $A_N$. However, this measure of variability is probably inappropriate for the same reason that the CV of $A_N$ is inappropriate, as described for Table 8.18. A more appropriate measure of site variability would be the coefficient of variation of the processes of respiration and assimilation considered separately.

Table 8.20 Comparison of site $A_N$ and variability (Maximum difference/average) for several experiment sites. (Taken from Carrara et al. (2003)). Forest type is either coniferous (Con), deciduous (Dec) or mixed (Mix).

<table>
<thead>
<tr>
<th>Site</th>
<th>Forest type</th>
<th>Planting date</th>
<th>Variability max diff/average %</th>
<th>$A_N$ average g m⁻² a⁻¹</th>
<th>$A_N$ maximum difference g m⁻² a⁻¹</th>
<th>Measurement period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vielsam</td>
<td>Con</td>
<td>1925</td>
<td>26</td>
<td>-720</td>
<td>190</td>
<td>1997-2001</td>
</tr>
<tr>
<td>Hyytiala</td>
<td>Con</td>
<td>1962</td>
<td>40</td>
<td>-194</td>
<td>77</td>
<td>1997-2001</td>
</tr>
<tr>
<td>Vielsam</td>
<td>Dec</td>
<td>1925</td>
<td>41</td>
<td>-460</td>
<td>190</td>
<td>1997-2001</td>
</tr>
<tr>
<td>Loobos</td>
<td>Con</td>
<td>1920</td>
<td>55</td>
<td>-319</td>
<td>175</td>
<td>1996-2001</td>
</tr>
<tr>
<td>Harvard</td>
<td>Dec</td>
<td></td>
<td>64</td>
<td>-220</td>
<td>140</td>
<td>1991-1995</td>
</tr>
<tr>
<td>Soro</td>
<td>Dec</td>
<td></td>
<td>93</td>
<td>-177</td>
<td>165</td>
<td>1996-2001</td>
</tr>
<tr>
<td>Hesse</td>
<td>Dec</td>
<td>1965</td>
<td>160</td>
<td>-322</td>
<td>514</td>
<td>196-2001</td>
</tr>
<tr>
<td>Norunda</td>
<td>Con</td>
<td>1900</td>
<td>205</td>
<td>220</td>
<td>450</td>
<td>1995-2001</td>
</tr>
<tr>
<td>Brasschaat</td>
<td>Mix</td>
<td>1929</td>
<td>238</td>
<td>111</td>
<td>264</td>
<td>1997-2001</td>
</tr>
</tbody>
</table>
A comparison of the $A_N$ value for Griffin with values from a more extensive list of FLUXNET sites (Falge et al. 2002) is shown in a flux-climagram plot in figure 8.91. The values of $A_N$ in this figure are proportional to the size of the circles and are plotted with respect to the relevant 0.5 degree latitude/longitude grid cell temperature and precipitation (New et al. 1999). It is apparent that warmer and wetter conditions lead to larger values of $A_N$ and that the moderate temperature and high precipitation of the Griffin site place it in a climatological region conducive to large values of $A_N$. Although this figure provides a compact graphic representation of the effect of climate on carbon exchange, a greater range of flux sites, located in more marginal regions are needed to obtain a more complete picture of the climate response of $A_N$. A flux-climagram as shown in figure 8.91 provides a useful tool for identifying climatic regions in which further research could be focused.

![Figure 8-91 Climagram of $A_N$ values of selected FLUXNET sites (taken from Falge et al. (2002)). The area of the circles are proportional to the size of $A_N$ with larger circles representing a larger carbon exchange. The black circles represent carbon losses, the grey circles carbon gain and the white circle represents carbon uptake at the Griffin site. Temperature and precipitation data are taken from New et al. (1999).](image-url)
8.8 Summary

This chapter describes the fluxes and storage terms obtained over the five years of this experiment. Additional derived variables as well as energy, water, and carbon budgets are described.

The large magnitude of the sensible heat storage term indicates that effort expended on its measurement is worthwhile, with the greatest focus on biomass and soil heat storage. In contrast, the low magnitude of the latent heat storage term and its similarity to that of the air column sensible heat storage, suggests that modelling of this component may be more beneficial than expenditure of effort on continued measurement. The magnitude of the CO\textsubscript{2} storage component, however, was large enough to merit continued monitoring. For all storage components, better spatial averaging of within canopy values are needed to allow more realistic models of storage components to be developed.

Duplication of fast-response sensible heat flux measurements proved beneficial in determination of flux errors. It is recommended that duplication of other scalars also be considered as a method of quality assessment and control. Because duplication of sensors is not always possible it is also recommended that further investigation into instrument errors, such as those found in this thesis, be undertaken. Similarity assumptions may be improved by placement of sensors further from the canopy and away from the roughness sublayer. Spatial placement of sensors in the landscape to minimize advective effects requires further investigation into the dynamics of hill flow and drainage and how they may affect fluxes. Measurements of profiles of scalars may be important in demonstrating advective effects. Corrections to momentum and sensible heat fluxes were small, corrections to $F_{c}$ were an acceptable 30\% but those for $\lambda E$ were a substantial 100\%. Despite this quite sizable correction, which was justified in Chapter 5, the resulting values of $\lambda E$ lead to reasonable results. Sensible heat fluxes were determined from both sonic and thermocouple temperature measurements. The sonic-derived $H$ required correction for significant errors caused by sensor head deformation.
Vertical velocity turbulent intensity followed existing similarity models but stability adjustment parameter values did not adhere to an existing model. Both showed evidence of a topographic dependency. Scalar turbulent intensity values followed similarity models.

Diel patterns of momentum appeared to be related to boundary layer development while annual patterns were likely related to global circulation patterns. The momentum derived parameters $C_D$ and $z_0$ showed strong directional dependencies related to canopy structure, but no evident topographic dependence. Although estimates of zero-plane displacement based on momentum were less successful than those derived from wind speed profiles, the value of $z_0$ were observed to varied from 0.74 to 1.3 m over the period of the experiment.

Annual and diel patterns of $H$ and $\lambda E$ were strongly controlled by radiation with secondary effects of humidity. As a result, both the annual and diel patterns of $\beta$ indicated a reduction in the available energy partitioned to $H$ over the course of both cycles. The resulting change in $\beta$ could be attributed to direct effects of increased $D$ on $\lambda E$ and there appeared to be little restriction caused by reductions in canopy conductance, the values of which were derived using the Penman-Monteith equation. A directional dependence of energy partitioning was also observed with values of $H$ reduced and $\lambda E$ enhanced for along row flow. To estimate missing values of $\lambda E$, a simple empirical model of canopy conductance was combined with a model of canopy wetness. Gap filled values of $\lambda E$ were combined with precipitation and stream flow measurements to obtain reasonable closure ~92% of the site water budget. Energy budget closure was approximately 91% and showed distinct increases over both annual and diel cycles. A directional dependence of energy budget closure with a bimodal pattern suggested a relation to canopy structure.

The diel pattern of $F_c$ was controlled primarily by radiation so that larger post-midday $F_c$ could be attributed to higher afternoon irradiance, despite reductions in afternoon canopy conductance. This effect was also observed to be larger prior to midyear. Despite the strong control of radiation, the overall variability of $F_c$ was similar to that of radiation prior to midyear but larger than that of radiation after midyear. The
interannual variability of $F_c$ was dominated by radiation, as was the annual pattern of $F_{eg}$. However, the annual pattern of $F_c$ peaked in May because of the lagged peak in the annual pattern of $R$, which was dominated by temperature. Both exponential and Lloyd and Taylor models of respiration responded similarly and were improved most by the addition of a soil moisture term. Further improvement could be obtained by including effects of the spatial variability of soil moisture.

Evidence of nocturnal flux loss was estimated by comparison of data with modelled respiration. $F_c$ flux loss was found to be proportional to flux magnitude and a function of $u^*$, occurring at all $u^*$ and not only at low $u^*$. Further analysis of the temporal dependence of nocturnal flux loss and analysis of diurnal model residuals indicated that flux loss was not restricted to nocturnal periods. An additive, wind direction-dependent, flux loss was also observed and suggests the existence of advection as a partial cause of flux loss. Analysis of energy budget closure in relation to $u^*$ points to the possibility of diurnal flux loss in all measured fluxes, not only in the carbon dioxide fluxes.

The diel $u^*$ corrected $F_c$ flux estimate of annual carbon uptake of 721 g m$^{-1}$ a$^{-1}$ were combined with the above-ground biomass increment estimates, and estimates of soil carbon increases to estimate carbon budget components. Missing estimates of root biomass increments made it impossible to close the carbon budget. The resulting estimates of root shoot ratio of 45% and proportion of below-ground carbon accumulation of 31% were consistent with observations by other researchers.
The primary purpose of the experiment reported in this thesis was to determine the carbon exchange of a forest representative of Scotland and to determine the atmospheric, biospheric, and edaphic factors that are affecting that carbon exchange. Sitka spruce was selected as being representative because of its extensive plantings in Scotland. This thesis presents five years of data obtained from this experiment and details the methods used to obtain justifiable values of site carbon exchange and related processes. In this chapter conclusions are presented based on the results of this experiment and recommendations are given for the conduct of future research at this or similar sites.

**Energy and water budget**

An average energy budget closure of 91% was obtained using all five years of data, with Bowen ratio values typically between 1 and 1.5 during midday. A closer, temporal examination of energy budgets showed evidence of improvements in closure from 80% to near 100% over both diurnal and summer periods, while Bowen ratio values tended to decrease over the same periods. These temporal patterns suggest a relationship of higher proportion of $H$ with poorer energy budget closure. Directional variation of energy budget closure additionally suggests that both site topography and canopy architecture are related to errors in flux measurement. The current corrections to the fluxes of sensible and latent heat and inclusion of all energy budget components allowed for the improvement in energy budget closure from a value of 64% determined from real-time fluxes to the final value of 91%. This improvement in closure resulted primarily from corrections to $H$ and especially $\lambda E$ but also from inclusion of heat storage. Storage of $H$ was a substantial proportion of $H$ (median of 17%), composed primarily of energy stored during photosynthesis and the changes in heat content of soil and canopy moisture; the proportion of storage of $\lambda E$ was small (3%). Energy budget closure is often used to justify the quality of flux measurements; however, the flux specific corrections to $H$ and particularly those to $\lambda E$ along with the observed potential for differing low frequency flux losses for different fluxes, suggests that energy budget closure may be an inadequate measure of the quality of non-energy budget fluxes. Estimates of water budget closure were used.
as an alternate verification method. A similar closure was obtained in the site water budget, though over a shorter period.

**Carbon exchange**

The carbon uptake was estimated using currently accepted methods, with only slight modifications to frequency response and $u^*$ correction methods. A resultant mean annual uptake of carbon of $671 \pm 61 \text{ g m}^{-2} \text{ a}^{-1}$ was obtained, in which the error term represents the standard deviation of interannual variations. Further modifications to corrections to account for diurnal flux loss gave a revised site carbon uptake of $749 \pm 59 \text{ g m}^{-2} \text{ a}^{-1}$, implying that the current approach of applying $u^*$ corrections only at night may be leading to substantial underestimations of carbon uptake. Radiation was the primary driver of the variability of net CO$_2$ flux, in spite of the strong control of respiration by temperature. This is substantiated by examination of the variability of the flux components of carbon uptake, $A_G$, and respiration, $R$ (five-year mean values of $-2253 \pm 83 \text{ g m}^{-2} \text{ a}^{-1}$ and $1532 \pm 37 \text{ g m}^{-2} \text{ a}^{-1}$, respectively). Expressed as a coefficient of variation, the interannual variability of $A_G$ (3.7%), which is primarily driven by radiation, was larger than that of $R$ (2.4%), which is controlled by temperature. Implicit error estimation was not performed on the flux values, though a range of calculations used in the analysis of carbon fluxes suggests that errors of $\pm 100 \text{ g m}^{-2} \text{ a}^{-1}$ could be expected as a result of small changes in flux calculation and gap filling techniques. Measurements of soil carbon accumulation gave estimates of soil carbon sequestration of $111 \text{ g m}^{-2} \text{ a}^{-1}$, while allometric estimates of biomass increases indicated an average above ground annual carbon allocation of $421 \text{ g m}^{-2} \text{ a}^{-1}$. Using the site carbon uptake, $A_G$, it was possible to calculate the apportioning of carbon uptake with 68.0% lost as respiration, 18.7% allocated to above ground biomass, 8.4% stored in below ground biomass and 4.9% sequestered to the soil.

A comparison of the carbon uptake at Griffin with values from other FLUXNET sites shows that the uptake at Griffin is among the largest observed. This large uptake may appear anomalous considering the site's high latitude but is consistent with the site's climate regime. If we consider the large plantings of coniferous species, of primarily Sitka spruce, in Scotland over the past 20 years ($\sim 10^9 \text{ m}^2$) and the even larger plantings in the preceding 40 years ($\sim 10^{10} \text{ m}^2$), it is apparent that this forest cover type
has the potential of retaining a substantial amount of carbon over the next half century.

**Experiment infrastructure**

The distance to the experiment site limited access to, and thus maintenance of, the experiment. This limitation and lack of mains power at the site affected both the quality and quantity of information that could be obtained during the experiment. It is recommended that, if possible, future experiment location be selected to optimise either accessibility or power availability.

The lack of mains power required a combination of power sources (wind, solar, propane) to supply the experiment. Power consumption exceeded initial experiment specifications causing excess reliance upon the propane generator power supply. To reduce the potential for contamination by burning fossil fuels, it is recommended that power consumption be reduced by eliminating high power demand profile CO$_2$/H$_2$O measurements and replacing them with modelled values based on campaign measurements supported by temperature profile measurements. It is also recommended that power supplies be met by increasing the amount of solar supplied power. Because solar power will likely be inadequate during the winter, it will be necessary to structure the eddy covariance data collection system to enable automated initiation of data collection after power outages during the winter months. Data loggers should be given independent power sources that will enable them to maintain data collection year round.

The two-tower instrument deployment scheme used at Griffin met the needs of unobstructed instrument deployment and may have provided more representative canopy profile measurements and a wider variety of surface measurements. However, the large distance between towers probably resulted in inefficient use of power supplies and negated the possibility of synchronization and computer backup of data loggers. It is recommended that if a second tower is required that it be located closer (20 to 40 m) to the primary, eddy covariance tower. The location of the eddy covariance tower included a forest ride in the center of its flux footprint. It is recommended that the tower be moved to reduce the amount of non-forest area in the flux footprint and to maximize the distance to forest edges.
Site characterization

The 10 m by 10 m forest sample plots employed provided samples of sufficient size to be easily measured and still have enough trees in each plot to obtain usable statistics. The gridded layout of the plots provided a useful spatial representation of canopy biomass and provided distinct locations for ancillary measurements. However, the plot layout grid was impractical for repetition of measurements because of the difficulties in relocating many of the plots after several years of non-use. It is recommended that well marked transects be used in place of grids. Such transects would ease relocation and re-measurement of specific trees, and would simplify identification for remote sensing purposes.

Measurement of forest DBH at annual intervals was probably excessive and it is recommended that extensive initial measurements be followed by re-measurement at three to five year intervals. Measurement of canopy height and structure were insufficient and greater effort should be expended on obtaining spatial variation of these values, especially in the flux footprint.

Monthly collection of litter traps was appropriate despite the irregular nature of Sitka litter fall rates. Less frequent collection of litter would have led to increased loss of litter mass through litter decomposition. Because of the spatial variability of litter fall, it is recommended that larger litter traps be distributed to sample regions of known canopy variability. Placement of litter traps in areas of impending canopy closure would be of particular interest. Separation of litter fall into components and analysis for mass and C/N ratios should be done for representative canopy areas over the course of a year to see if the relationships found in this thesis vary over longer time periods.

It is recommended that more information be gathered on the spatial variability of soil properties (organic layer depth, soil texture). This information should cover the entire experiment site and should be correlated with remotely sensed canopy cover information. Greater information on spatial variability of soil moisture, temperature and topography should also be gathered in the region of the flux tower and should be co-located with soil respiration measurements to allow improved modelling of soil
efflux. This expanded sampling should be stratified further by topography, and by radiation environment if the canopy is thinned.

**Data management**

Signal sampling rates for eddy covariance and logger data acquisition systems were sufficient. However it is recommended that data logger systems be used as backup for eddy covariance methods (by use of variance methods), which will require increased sampling rates and retention of standard deviations. Establishment of computer/data logger connectivity is recommended to allow less frequent collection of data from both eddy covariance and data loggers systems. Retention and back up of raw eddy covariance data is highly recommended.

It is recommended that a common (i.e. networked) data storage system be devised for storage of unprocessed data from all sources. In addition, software tools for analysis of unprocessed data should be made available from the same source. It is also highly recommended that extensive manual and automated quality control be applied to all data sets. The quality control should produce incremental flags for each signal and should not remove data from the unprocessed data set. The flags should be used later in the processing stream to remove data at a stage that would minimize propagation of errors.

**Microclimatology**

The consistency of global and reflected radiation sensors was such that duplicate sensors should not be required (so long as sensors are properly deployed). However, it is recommended to have backup sensors available and ready in case of sensor failure. Based on the subtle differences observed between short wave radiation and PAR, it is recommended to measure both incoming and reflected short wave radiation and PAR using sensors of similar construction. Multiple net radiation sensors should be deployed both to ensure data coverage in case of failure and as a check of instrument stability. If employing multiple radiation sensors (or any duplicated microclimate sensors) it is recommended that they be deployed on separate loggers with independent backup-power supplies since this will ensure data coverage in case of logger failure.
Sensors for measuring transmitted PAR should be placed to optimise sensing of canopy variability. In a thinned canopy this may best be accomplished using line transects, while in unthinned canopies placement may require spatial assessment of canopy structure to determine appropriate placement of sensors.

Measurement of diffuse radiation is also recommended. However, the use of shade band determination of the diffuse component is discouraged because the possibility of infrequent site visits and predominantly cloudy conditions makes consistent adjustment of such sensors difficult.

Profile measurements of air temperature may be made using thermocouples referenced to platinum RTDs. It is recommended to determine the effects that sensor shading, ventilation, rain/mist and ice have upon the sensor probes and the resulting temperature values. Soil temperature measurements should be sufficient to capture the variability of soil surface topography (for the purpose of soil respiration modelling) as well as any variability resulting from canopy cover variability. Multi-level soil temperatures should also be obtained for the purpose of determining soil thermal properties and as an alternate method of estimating soil heat flux. As with soil temperature, canopy bole temperatures should capture any variability of the sub-canopy radiation environment resulting from thinning. It is recommended that a minimum of two depths and three levels be employed in each tree measured. If possible infrared measurements of canopy temperature should be made.

Resistance or IRGA humidity sensors should be used in preference to psychrometers because of infrequent site visits and because of poor measurement ability of psychrometers during winter. Profiles of humidity are not crucial and it is recommended to attempt modelling of this component based on correlation within canopy CO\textsubscript{2} or temperature. It is recommended to assess the effects of liquid water or ice on the sensor housing and ventilation upon measurements.

Although CO\textsubscript{2} concentration measurements would be available from eddy covariance measurements, it is recommended to also obtain profile measurements of CO\textsubscript{2} for estimation of CO\textsubscript{2} flux storage. To minimize power requirements these measurements should be done on a campaign basis. These profile measurements should be more
highly concentrated near the surface and intakes should also be distributed horizontally to determine effects of canopy thinning on the spatial distribution of storage. If possible, methods of improving instrument stability should be implemented to reduce the burden of post-processing correction of CO₂ concentrations, which is both time consuming and prone to inaccuracy.

Both cup and sonic anemometer measurements should be retained for duplicate above canopy wind speed measurements. Because estimates of wind direction are less crucial, duplication of wind direction measurements is not required, but accurate positioning of the sonic anemometer should be undertaken. The use of cup anemometer profiles provided the only reliable measure of canopy zero plane displacement. However, these profiles should be removed during winter months for refurbishment and calibration and to minimize the amount of erroneous data resulting from snow and ice deposition on the anemometer cups. It is also recommended, if possible, to deploy sonic anemometers below canopy to obtain higher accuracy profiles of within-canopy wind speed and direction. Such data may be useful for determination of pressure gradient induced horizontal advection and katabatic flows.

Precipitation and stream flow measurements proved useful in assessing the water budget of the site and should be continued. Precipitation may be measured in open areas but losses resulting from interception by adjacent forest canopy should be considered. Installation of the gauge above the canopy and testing for reliability of such a measurement should be done. The precipitation gauge should be installed for easy maintenance access to allow for frequent cleaning to minimize data gaps. If possible, duplicate, manual or automated gauges should be deployed. It is also recommended to deploy a low power precipitation incidence sensor to identify periods of precipitation, which are immeasurable by the precipitation gauges. Such information may help in estimation of missing latent heat flux values. Precipitation throughfall measurements were useful in determining canopy storage capacity and the modelling of latent heat flux. It is recommended to continue these measurements, preferably using automated precipitation gauges. Stem flow values were generally small and can likely be estimated using information from the literature.
Soil moisture should receive better temporal coverage with a minimum of daily measurements throughout the experimental period. Spatial coverage of soil moisture measurements should reflect the soil microtopography and other factors resulting from canopy thinning that would affect soil moisture. Soil moisture probes should be co-located with spot measures of soil respiration measurements to allow more appropriate modelling of soil respiration.

Soil heat flux was a small energy flux component at the Griffin site. For future measurements in similar unthinned, closed canopies a minimum of effort should be expended on soil heat flux measurements. Soil heat flux measurements in a thinned canopy should focus on identifying soil heat flux variations caused by changes in surface radiation as a result of thinning.

**Flux measurements and techniques**

The eddy covariance system deployed at Griffin was sufficient for monitoring site fluxes. However, it is recommended that a duplicate system be installed on a campaign basis to address several possible questions. First, the system should be installed at a spatially separate location or locations to test for spatial variability of fluxes and to ensure that the primary location is not subject to flux bias resulting from its selected location. Second, the duplicate system should be installed in the same vertical profile as the first system to identify if flux divergence exists and, if possible, to identify statistical variations across the roughness to surface layer transition.

Measurement of momentum and turbulence values showed only small correction effects, however, it is recommended to further examine sensor attack angle errors and their effects on turbulence and derived parameters. Estimations of canopy roughness length and drag coefficient were reasonable but showed distinct effects of canopy architecture. It is recommended to continue these estimations and to employ a duplicate sonic anemometer to probe the spatial variability of the thinned canopy and the effect of this variability upon the estimation of canopy parameters from eddy covariance derived parameters. Further, refined, attempts at determination of canopy zero plane displacement should be attempted.
Estimates of site energy budget closure indicated small underestimates in measurement of energy budget components under some conditions. Duplication of sensible heat flux and net radiation allowed for identification of possible errors and it is recommended that duplication of all energy budget components be measured in future experiments. Improvement in energy budget closure benefited from improved corrections to sensible and latent heat flux components. Corrections for sonic head distortion were significant and should be applied if similar probe designs are used in future experiments. Wind tunnel tests of probes should be carried out to parameterise correction equations. Frequency response corrections to latent heat flux for enhanced tube attenuation should be applied if closed path sensors are used. It is recommended that an analytical solution to this attenuation transfer function be attempted and should be generalized to include parameters such as gas species and tube length. Additional corrections as described in this thesis should be applied as is considered appropriate. It is recommended to further examine data from Griffin for causes of lack of energy budget closure. Particularly, physical reasons for the relation of lack of closure to diel and annual patterns and to friction velocity should be investigated. Estimation of site water budget should be improved if previous recommendations of sensor deployment are followed.

Site annual carbon exchange measurements were obtained with a possible uncertainty of at least 15%. Verification of errors associated with low $u_*$ conditions would help to reduce this uncertainty. Recommendations for auxiliary flux experiments mentioned previously should include measurements of carbon dioxide flux to assist in verification of the errors observed in this thesis. In addition, it is recommended that measures of belowground allocation of biomass be attempted in order to obtain an independent measure of site net carbon exchange. To this same end, further measurement of soil carbon stocks should be conducted to improve accuracy of soil carbon accumulation and above canopy biomass should continue to be estimated from DBH measurements. For both energy and carbon fluxes it is also recommended to take a more direct role in modelling of exchanges. Such efforts are being undertaken by cooperating researchers but closer contact and interaction should be developed with these modellers.
General comments

The objective of this experiment was to determine the carbon uptake of a Scottish plantation forest and to improve the understanding of the environmental controls of such exchanges as well as the methods of measuring these exchanges. This is a common goal, and many researchers are undertaking this task in response to governments' need to understand the carbon and energy cycling in regions for which they are responsible. This goal is becoming increasingly common, with corresponding pressure to provide increasingly refined values for ecosystem exchange and an increasingly sophisticated understanding of underlying processes. This situation has resulted in an increase in the amount of information available but has also revealed some of the limitations of current methods of measurement. The additional pressure to produce high quality information on short time scales has led, at least in the case of this experiment, to the distribution of inferior, preliminary data. Such data should be flagged to alert users to its status and users of such data should consult with data producers and take extraordinary care in the conclusions drawn from such data. Considering the complexity of meta-data associated with final flux data, similar consultation should also be extended to the final distributed data. Ideally, distribution of preliminary data should be severely restricted because of the relative ease with which it can obtain permanence within the body of scientific knowledge if its quality is misinterpreted, or assumed by secondary users.

It is believed that this thesis reports justifiable and accurate estimates of both mass and energy exchanges of an ecosystem characteristic of Scotland and provides an improved understanding of both the processes which control these exchanges as well as an improved understanding of the methods employed in obtaining these values. It is hoped that these results have a positive influence on the ongoing research into the effects of ecosystems upon the earth's climate.
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A Climatological measurement methods

This appendix describes the methods of climatological variable measurement at the Griffin experimental site. Each section contains a description of measurement methods and a comparison of sensors employed. The measurement methods section describes the sensors used and their installation and maintenance. When necessary, theoretical and/or empirical relations are development to improve the quality of the variable obtained from an instrument's signal.

A.1 Radiation

The radiation spectrum has been divided into named regions. Those of interest to this experiment include short-wave radiation at wavelengths between 0.3 and 3.0 μm (1991), long-wave radiation which ranges over 3 to 50 μm, and photosynthetically active radiation, PAR, from 0.4 to 0.7 μm (Brock & Richardson 2001). The measure of PAR quanta is referred to as photosynthetic photon flux density, PPFD.

The short-wave spectrum is important as it represents the primary energy input to the ecosystem. As a subset of short-wave radiation, PAR is of special interest because of its use in the photosynthetic processes of plants. The wavelengths of long-wave radiation correspond to those emitted by a body with a temperature equivalent to that of the earth's surface. In contrast to short-wave radiation, the up-welling component of long-wave radiation is greater than the down-welling component so that long-wave radiation represents a loss of energy from the earth's surface.

Down-welling radiation is composed of direct beam and diffuse components as well as small amounts of radiation emitted by the overlying atmosphere or reflected up-welling radiation. By definition, direct beam radiation will have followed a direct path from the sun to the observer, while the path of diffuse radiation will have been altered by scattering or reflectance prior to reaching the observer. In comparison, up-welling radiation consists of radiation reflected by or emitted from the underlying surface and atmosphere. The net exchange of up-welling and down-
welling radiation defines the net surface irradiance, which approximates the net energy gained or lost by radiative transfer

A.1.1  Global irradiance

Global irradiance is defined as the total energy received as radiation per unit surface area tangential to the earth's surface. As mentioned above, its measurement includes both direct and diffuse components of radiation.

A.1.1.1  Short-wave

A sensor used to measure the energy flux density of short-wave radiation is referred to as a pyranometer. Pyranometers are typically constructed from thermopiles, which measure the heating of a black body surface and respond to the amount of short-wave energy received. Thermopile pyranometers are sensitive to all radiation and depend upon glass domes to shield them both from long-wave and ultraviolet radiation and from the convective cooling effects of wind.

Photodiodes, with appropriate band-pass filters to exclude radiation outside the desired spectral region may also be used to measure the quanta flux density as a photoelectric response to short-wave radiation. Because the amount of energy carried by light of different wavelengths varies, the two measures may only be compared directly if the irradiance distribution at all relevant wavelengths is identical for calibration and measurement situations (Ross & Sulev 2000). Nevertheless, if calibrated and properly exposed under natural radiation conditions, the accuracy of a photodiode pyranometer can be ±3 to 5% because of the relatively constant proportions of solar radiation at different wavelengths.

Both types of pyranometers have a non-linear cosine and azimuth response to solar elevation and azimuth angles, respectively. For thermopile pyranometers the cosine response may result in underestimates of up to 10% at low solar elevation angles (Michalsky, Harrison, & Berkheiser 1995). The design of most photoelectric pyranometers adjust for cosine response with resulting errors of ±10 to ±15% of radiation at a solar elevation angle of 10 degrees (Michalsky et al. 1995).

Additional errors for both types of sensors arise from dirt, water, or snow on the
pyranometers dome or diffuser, or condensation inside the sensor. These effects are rectified through proper sensor maintenance or devices to actively remove water and snow from the sensor. Sensor response will also be affected by interference of the sensing area by structures, instrumentation or vegetation. This problem should be avoided during sensor installation.

Appendix A

Measurement methods

Global short-wave radiation, $R_g$, measurements were made with Kipp and Zonen pyranometers, models CM2 or CM3, which measure wavelengths between 0.305 and 2.8 $\mu$m. The most recent available calibration coefficients were used to convert the sensor’s millivolt signal to units of W m$^{-2}$. Unfortunately, the most recent calibrations available were from the mid 1980s. A calibration check of some of the pyranometers was carried out in 1997. However, as no reference sensor was available, an updated calibration of the sensors was not possible. A comparison with recently calibrated PPFD sensors exhibited a short-wave to PPFD ratio ranging from 0.49 to 0.6 for cloudy to clear conditions, respectively. These ratios are in line with results from other research, see section 4.3, suggesting that no profound pyranometer problems existed.

Initially one pyranometer was installed a short distance above the mean canopy height, on the eddy covariance tower. To increase the surface area represented by its measurements, the sensor was later mounted on an arm extending approximately 1 m to the southeast of the tower at a position 6 m below the top of the tower. This position was selected to minimise potential flow distortion effects on the wind vane and anemometers mounted higher up the tower. Because the position of the pyranometer was below the top of the tower, a maximum of 1.5% of the sensor’s view angle was obscured by the tower (Sparrow & Cess 1970); further obscuration by instrumentation may have slightly increased this percentage. A second pyranometer was installed at the top of the profile tower in March 1997. This sensor’s position afforded it an unobstructed view of the sky.

Pyranometer domes were cleaned using distilled water and a clean tissue. The frequency of cleaning varied from weekly to monthly depending upon workload
and sensor accessibility. The internal desiccant of the pyranometers was changed annually. However, moisture ingress in the pyranometer mounted on the eddy covariance tower occurred occasionally until the end of 1999, requiring frequent replacement of the internal desiccant. The problem resulted from a failure of the seal between the dome and the dome's mounting ring. Upon discovery, the dome was replaced.

The infrequent nature of site visits and the frequent occurrence of precipitation prevented timely removal of liquid water, snow, or ice from the sensor’s glass domes. When observed, snow and ice were removed. It was believed that the reduction of radiation because of rain on the sensors was negligible. The effect of snow on the short-wave irradiance may have been significant.

The data were converted from millivolts to W m⁻² and half-hour averages were recorded using data loggers. The half-hour averaged data were collected and quality controlled as described in chapter 3, and used in subsequent analysis. Zero drift of global and reflected pyranometers was determined using the quality-controlled data set. In this process, data were removed for solar elevation angles greater than -5°, after which outliers were re-assessed and removed. The data were grouped by nocturnal periods and examined for consistent patterns within each period, long-term trends, and relations to environmental conditions (wind speed, air temperature, wet bulb temperature, wet bulb depression, and net radiation). No consistent relationship to environmental conditions and no long-term trends were observed in the data. Additionally, there was sufficient evidence to suggest that radiation sensor offsets remained relatively consistent over a nocturnal period so that a single offset would apply over a nocturnal period. The resolution of the pyranometer on the profile tower was approximately 3 W m⁻², so that the zero offset applied to this sensor was less accurate than that applied to the pyranometer on the eddy covariance tower, for which the resolution was 0.7 W m⁻². Radiation sensor zeros from each nocturnal period were linearly interpolated over diurnal periods.

Calibration changes and adjustment for sensor zero drift were applied to the radiation sensors during post-collection processing and quality control flags were
used to remove data from the processing stream. A final value of $R_g$ was obtained by averaging values from both pyranometers using equal weighting.

**Sensor comparison**

Comparison of the two pyranometers deployed during this experiment shows that, on average, the two sensors have nearly identical response, with a regression slope of 0.99, $R^2 = 0.99$, figure A.1. However, visual inspection of the relationship indicate a larger than expected amount of scatter.

![Graph](image)

*Figure A.1 Comparison of measured $R_g$ values for sensors on the eddy covariance and profile towers.*

Further analysis was done in an attempt to discover the cause of this scatter. Linear regression coefficients were determined for the data grouped into 10-degree bins for both solar azimuth and elevation angle, figure A.2. The near zero intercepts for sunrise and sunset periods (elevation angles between 0 and 10 degrees and azimuth angle magnitudes larger than 45 degrees) indicate that there was no inherent offset difference between the sensors. For elevation angles greater than 10 degrees, the increasing intercept and decreasing slope for lower solar elevation angles suggests a non-linearity in the relationship between the two sensors. This effect may be the result of one of the two sensors being tilted more towards the sun.
To investigate this problem further sensor tilt angle, $\beta$, and sensor tilt aspect, $\alpha_s$, were estimated by fitting measured values of $R_g$ to an equation describing the effect of surface slope on radiation received at the correctly oriented sensor, $R_{go}$ as a function of $\beta$, $\alpha_s$, solar elevation angle, $\beta$, and solar azimuth, $\alpha$, (Linacre 1992):

$$R_{go} = R_g \left( \cos(\beta) + \frac{\sin(\beta_s) \cdot \cos(\alpha - \alpha_s)}{\tan(\beta)} \right)$$

(A.1)

![Graph](image)

Figure A.2 Comparison of linear regression slope and intercept coefficients with solar azimuth angle for different values of solar elevation angle.

A linear least squares fit indicated a (relative) sensor slope of approximately 0.8 degrees to the north. When this effect was applied, only slight improvement was observed in the relationship. Because of the difficulty in separating direct and diffuse beam radiation, the likelihood that this tilt had changed over the course of the experiment, and the possibility of different cosine responses of the two sensors it was decided not to implement this adjustment. As a result, a small systematic
error that was a function of solar position and date, may have existed, though no further attempts were made to explain this error.

A.1.1.2 PAR

While photosynthetic photon flux density, or PPFD (units of μmol m$^{-2}$ s$^{-1}$), is the quanta equivalent measure of PAR (units of W m$^{-2}$), their units are not directly convertible because of the difference in energy transferred by photons of different wavelengths. An accurate conversion would require knowledge of the respective flux densities for all wavelengths within the 0.4 to 0.7 μm band. Several studies where found which provide factors for the conversion between units (Gonzalez & Calbo 2004; Jacovides et al. 1999; McCree 1972) suggest that the conversion is similar under many conditions. However, to avoid the potential for conversion errors, comparisons of PPFD with short-wave radiation will be in units of μmol m$^{-2}$ s$^{-1}$ and W m$^{-2}$, respectively.

Measurement methods

Photodiodes are used to measure global PPFD, $Q_{pg}$. These sensors have characteristics similar to those of photodiodes short-wave detectors described in section A.1.1. The PPFD sensors were calibrated in May of 1997 by comparison with a reference sensor, which had recently been calibrated by the manufacturer. The resulting calibrations were applied to the sensor millivolt output to convert it to the PPFD units of μmol m$^{-2}$ s$^{-1}$. The data were further managed as described for $R_g$ measurements in section A.1.1. The data presented are the resulting half hour averaged values.

A single $Q_{pg}$ sensor was installed on the top of the profile tower. Care was taken to ensure that the level of the diffuser of the $Q_{pg}$ sensor was at the same level as the sensing element of the $R_g$ sensor. The dome of the $R_g$ sensor and the shade band of the diffuse radiation sensor inflicted much less than 1% sky obstruction (Sparrow et al. 1970) and neither obstructed the direct solar beam.

Additional measurements of $Q_{pg}$ were collected using the high frequency data collection system in order to inspect the higher frequency characteristics of $Q_{pg}$. The use of the 0.048 second sampling interval corresponded well with the fast time
response of photodiode sensors (< 1ms) (Licor 1992), and was necessary because only run-averaged data were retained from the low frequency loggers.

Sensor comparison

The run mean values of the two $Q_{pg}$ sensors employed are compared in figure A.3, though the $Q_{pg}$ sensor on the eddy covariance tower was employed primarily for obtaining high frequency radiation information. Because of the amplifier required to use the $Q_{pg}$ sensor on the eddy covariance tower with the fast response logging system, the calibration of this sensor was cross referenced with the output of the sensor on the profile tower. Therefore, figure A.3 is intended to show the amount of scatter in the relationship between the two sensors. The scatter in this comparison is similar to that observed in the comparison between the two $R_g$ sensors; however, it is known that much of the scatter in the $Q_{pg}$ relationship is the result of signal output drift associated with the amplifier used with the fast response logging system. In figure A.3, nocturnal offsets of $\pm 50 \mu$mol m$^{-2}$ s$^{-1}$ in the eddy covariance tower $Q_{pg}$ signal are evident at when the profile tower values are near zero.

Figure A.3 Comparison of $Q_{pg}$ values for sensors on the profile and eddy covariance towers. The figure is intended to show the scatter of the comparison and is not meant for comparison of absolute values, see text.
A.1.2 Diffuse radiation

A.1.2.1 Measurement methods

Diffuse short-wave radiation was measured using a pyranometer and shadow band combination. The shadow band was constructed from was a design provided by Licor Inc. which was based on the work of Horowitz (1969) and Turner (1983).

The shadow band and sensor were installed at the top of the profile tower on 29 July 1997. The shadow band was adjusted on a weekly basis for changes in the solar path. This adjustment was only possible when direct sunlight was available as the appropriate markings were not available on the shadow band to make the adjustment on a theoretical basis; this resulted in infrequent shadow band adjustment. During the periods surrounding solstices, the diffuse radiation measurements may have had a persistence of several days to a week or more; however, during equinox periods the measurements were only valid for a maximum of 3 to 4 days. The resulting data set was therefore restricted to the periods of five days following a recorded realignment of the shadow band, resulting in a sparse data set.

As with other radiation sensors, data were converted from millivolts to W m$^{-2}$ and handled as described for $R_g$ values in section A.1.1. However, quality control of this data set was more restrictive than for other radiation data sets because of the potential for error resulting from mal-adjustment of the shade band.

The values of $Q_{pr}$ required additional attention because of their sensitivity to the existence of surface snow. Winter days with abnormally high values of $Q_{pr}$ were removed, as were periods when either $Q_{pr} < 100$ μmol m$^{-2}$ s$^{-1}$ or $R_g < 50$ W m$^{-2}$ before values of $R_r/Q_{pr}$ were determined.
A.1.3 Reflected radiation

A.1.3.1 Measurement of reflected short-wave radiation and PPFD

All reflected short-wave radiation, $R_r$, measurements were made with Kipp and Zonen pyranometers, models CM2 or CM3. These sensors have the same characteristics and were handled the same as the sensors described in section A.1.1. Initially one pyranometer was installed a short distance above the mean canopy height, on the eddy covariance tower. The sensor was later moved above the canopy (3 m) to improve the sensor's viewing area. This sensor could not be moved to a higher position for the reasons stated in section 3.3. The reflected short-wave component measured on this tower was only operated until 11 December 1997, at which time water ingress caused sensor failure and this sensor was removed. A second reflected short-wave sensor was installed at the top of the profile tower on 7 March 1997. This sensor was located 8 m above the canopy; a position that provided improved spatial integration for the reflected measurements (Schmid 1997). The data were converted from millivolts to W m$^{-2}$ and handled as described for values of $R_g$ in section A.1.1.

Values of reflected PPFD, $Q_{pr}$, were measured using photodiode sensors. These sensors were calibrated, operated and maintained in the manner described for the global PAR sensors in section A.3.1.2. A single $Q_{pr}$ sensor was installed on the profile tower. A $Q_{pr}$ sensor was installed on the eddy covariance tower but proved susceptible to zero drift so that a second $Q_{pr}$ was later installed on that tower. Leaching of iron oxides into its diffuser later contaminated this sensor.

Sensor comparison

Comparison of the two available $R_r$ sensors, figure A.4, indicates that the values obtained from the profile tower were approximately 8% larger than those from the eddy covariance tower. The cause of this discrepancy may have stemmed from either sensor error or variation in the sites being measured.
Given that the sensor on the eddy covariance tower was much closer to the surface than the sensor on the profile tower, it is possible that this sensor was viewing a "dark spot" in canopy. If this were true, similar reductions may have been observed in $Q_{pr}$ and $R_r$. Such differences may have existed in $Q_{pr}$, see below, though they were not observed in $R_r$, see section A.1.6. Several possibilities of sensor error provide alternative explanations. As mentioned previously, the body of the $R_r$ sensor on the eddy covariance tower was more exposed to direct solar radiation, which could have lead to errors in $R_r$ (Brock et al. 2001). In addition, this sensor was installed prior to the calibration comparison of radiation sensors so that its calibration was not checked. It may be that this sensor’s calibration was in error; without more information, proper determination was not possible.

Similar to $R_r$, discrepancies were observed in the measurements of $Q_{pr}$. Table A.1 provides the regression slopes for the linear regressions of $Q_{pr}$ values from the two sensors located on the eddy covariance tower against those from the sensor on the profile tower. The data in table A.1 indicate a decrease over time in the values of $Q_{pr}$ from the eddy covariance tower relative to those from the profile tower. The
decline appears to be more rapid and more severe for the second sensor installed, which may be the result of the previously mentioned iron oxide contamination of the sensor's diffuser.

Table A.1 Linear regression coefficients (intercept = 0, $R^2 > 0.92$) for the $Q_{pr}$ values from sensors on the eddy covariance tower against values from the $Q_{pr}$ sensor on the profile tower.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>$Q_{pr}$ #1</td>
<td>0.882</td>
<td>0.787</td>
<td>0.693</td>
<td>0.674</td>
<td>0.620</td>
</tr>
<tr>
<td>$Q_{pr}$ #2</td>
<td>0.967</td>
<td>0.723</td>
<td>0.627</td>
<td>0.557</td>
<td></td>
</tr>
</tbody>
</table>

To determine if these discrepancies were caused by the sensor on the profile tower or because of degradation of both sensors on the eddy covariance tower, annual average values of $R_r/Q_{pr}$ were calculated for individual sensors, figure A.5. In this figure, data were limited to periods when $R_r > 10$ W m$^{-2}$ and $Q_{pr} > 15$ μmol m$^{-2}$ s$^{-1}$ and excluded the months of November through February.

The annual average values of $Q_{pr}$ and $R_r/Q_{pr}$ in figure A.5 suggest that the sensitivity of both $Q_{pr}$ sensors located on the eddy covariance tower decreased over time. Based on this information, values of $Q_{pr}$ from the profile tower were used in preference to those on the eddy covariance tower. When necessary, values of $Q_{pr}$ from the sensors on the eddy covariance tower, adjusted for approximate sensor drift, were used to represent $Q_{pr}$ in data analysis.

Figure A.5 Annual averaged $Q_{pr}$, left, and $R_r Q_{pr}^{-1}$, right, for three $Q_{pr}$ sensors deployed. Error bars represent ±1 standard error.
A.1.4 Albedo and $Q_{pg}/Q_{pr}$ ratio

Albedo values were found to be unacceptable at low radiation levels. In an attempt to exclude these and other questionable data, mid-day (10:45-13:15) mean values were calculated after removing values with corresponding solar elevation angles less than 15 deg., component $R_g$ values of less than 20 W m$^{-2}$ or with component $Q_{pg}$ values of less than 40 μmol m$^{-2}$ s$^{-1}$. Run mean values were then removed if they deviated from the mid-day mean by more than 0.015 for albedo and 0.01 for $Q_{pg}/Q_{pr}$. After performing this quality control, remaining values were used to calculate daily mean values. Further quality control was carried out by removing days with mean albedos less than 0.06 or greater than 0.14 and mean $Q_{pg}/Q_{pr}$ less than 0.01 or greater than 0.045. This approach will have excluded variations caused by either small values of global radiation or high surface reflectivity associated with intermittent snow cover. This approach differs from that used by Betts and Hall (1992) because of differences in data availability.

The large LAI of the canopy should have reduced the effects of the understory and surface on the forest albedo. However, because of the positioning of the towers in plots with canopies that were not closed, it is possible that surface reflectance will have affected the measurements of albedo (Miller et al. 1997). Another possible error may have arisen from the assumption that the point measurements were representative of the canopy cover (Knyazikhin et al. 1997).

A.1.5 APAR, FIPAR

A.1.5.1 Measurement methods

Values of $Q_{pt}$ were measured using the PPFD sensors described in section A.1.1.2. These sensors were calibrated, operated, maintained in the manner described for global PAR sensors. Sensors for measuring $Q_{pt}$ were placed in three plots and along one transect. Two of the plot installations, and the transect, were located close to the profile tower. The canopy was predominantly open in one plot and was closed in the other while the canopy along the transect varied. The third plot installation was adjacent to the eddy covariance tower, where the canopy was also
open. The sensors in the profile tower plots were co-located with soil heat flux plates and were deliberately placed to exemplify the variability in the plots. The sensors located on the transect were placed at an even spacing along, and at a random distance perpendicular to, the plough line which the transect followed. The sensors in the plot adjacent to the eddy covariance tower were placed randomly. In calculating $Q_{pf}$, data were excluded if $Q_{pg}$ was less than 40 $\mu$mol m$^{-2}$ s$^{-1}$ or the resultant value of $Q_{pf}$ was less than zero or greater than 100.

A.1.6 Net radiation

Net radiation is measured with a net pyradiometer, commonly referred to as a net radiometer. Net radiometers are commonly constructed as a pair of pyradiometers facing in opposing directions. The sensor thus consists of a differential thermopile that is used to measure the temperature difference between the sensing elements of the two pyradiometers. The temperature difference is related to the amount of radiant energy received by each sensor. Net radiometers often have shielding domes for the same purposes of those employed by pyranometers. These domes must minimize blockage of either short or long-wave radiation. Similar to the pyranometer, the net radiometer is sensitive to contamination of the domes by dirt, water, or snow. The net radiometer dome material is known to degrade under prolonged exposure and should be changed at annual intervals.

A.1.6.1 Measurement methods

Net radiation was measured using three net radiometers (REBS Q*6 and Q*7). The most recently available calibration coefficients were used to convert the sensor’s millivolt signal to units of W m$^{-2}$. It is known that the Q*7 has an additional wind speed correction required to account for advective cooling, however, this correction was not employed.

An initial net radiometer (Q*6) installation was located on the eddy covariance tower, in close proximity to the global and reflected short-wave pyranometers. This installation was a short distance above the canopy and was moved to a higher position in January of 1997. Two net radiation sensors (Q*6, Q*7, REBS, Seattle Washington) were installed at the top of the profile tower, (see chapter 3). These
two sensors were located at a greater height above the canopy (5 - 9 m) and at a
distance further from the tower (2 m), thus providing a greater spatial integration
for the reflected measurements and unobstructed view angles for incoming radiation
components. Sensor domes were replaced on an annual to bi-annual basis and the
domes were cleaned on a weekly to monthly basis.

The data were converted from millivolts to W m\(^{-2}\) and half hour averages were
recorded using data loggers and data were collected and quality controlled as
described in sections A.1.1. Post processing of the values was carried out to apply
any calibration adjustments.

No adjustment was made to the values of \(R_n\) for the difference in received short-
wave radiation due to the difference in slope of the surface being represented and
the horizontally installed radiation sensor. The associated error would have been
less than 1% of the measured value.

A final value of \(R_n\) was obtained after corrections by averaging the results of the
three available sensors.

**A.1.6.2 Sensor comparison**

A comparison of the net radiometers, figure A.6 revealed a good comparison
between the two Q*6 radiometers located on the two separate towers. However, \(R_n\)
values from the Q*7 net radiometer, located on the profile tower, were
approximately 18% smaller than those from both Q*6 sensors. Another obvious
problem is observed in the relationship between the two \(R_n\) sensors on the profile
tower, in which two distinct linear relations are observed. A comparison of the
regressions of each \(R_n\) sensor against values of \(R_g\), figure A.7, shows that the
sensitivity of the Q*6 sensor on the profile tower increased while that of the Q*7
sensor decreased between the years 1998 and 1999. Closer analysis not shown here
suggests that the change occurred in late December of 1998 or early January of
1999. Records to not indicate any profound event related to this change.
The resulting relationships afford the opportunity to determine error magnitudes associated with these sensors. The average RMSE value for the annual regression between the two sensors on the profile tower was 7.6 W m\(^{-2}\) while that between the
two Q*6 sensors on the two separate towers was 15.4 W m$^{-2}$. This suggests that approximately half of the observed error of 15.4 W m$^{-2}$ was due to random error associated with spatial separation of the sensors. From comparison of the sensor regressions done on an annual basis, there appears to be bias errors on the order of 2 to 6% associated with long term drift of the sensor calibration, in addition to the large bias errors (18%) between different sensor types. Such bias errors would account for at least 50% of the observed error between the two Q*6 sensors on the separate towers.

Only the strong bias associated with the Q*7 sensor was removed by adjusting for its long term regression against the averaged values of the two Q*6 sensors. Biases on shorter time scales were not removed but were instead minimized by averaging the results of all three net radiometers to obtain a single final value of $R_n$.

A.2 Temperature

In this experiment, four methods of temperature measurement were employed: thermocouples, thermistors, resistance temperature devices, and speed of sound. Thermocouples take advantage of the Seebeck effect in which a small current is created between two different metals with two junctions at different temperatures (Brock et al. 2001). This effect can be used to create a small non-linear voltage response to temperature ($\sim 40 \mu V C^{-1}$) but requires that the temperature of one of the junctions be known. Thermistors and resistance temperature devices (RTD) depend upon the temperature dependent change in electrical resistance of semiconductors and metals, respectively. Temperature can also be derived from the dependence of the speed of sound on the density of air when the pressure and composition of air are known. Measurement methods are described with respect to the specific variable to be obtained.

A.2.1 Measurement

Measurements of temperature were made in association with psychrometric, eddy covariance, canopy and soil temperature profile and biomass heat storage
measurements. Instrument deployment and associated measurement methods are described in the following sub-sections.

A.2.1.1 Psychrometer air temperature

Three dry/wet bulb psychrometers (Model VP1/TM1 Delta-T Devices, Cambridge UK) were installed during this experiment. One psychrometer was mounted on the eddy flux tower at a height of 10.7 m and two psychrometers were mounted on the profile tower at elevations of 14.6 m and 7.9 m. The upper two psychrometers provided above canopy temperatures throughout the experiment. Both the psychrometers on the profile tower were co-located with profile thermocouples to allow cross checking of the profile temperatures.

The only maintenance required by these sensors was the replacement of ventilation fans and the removal of insect residue from the fan housings. Occasionally fans stalled or failed because of this build up, causing sensor ventilation to fail, possibly resulting in temperature measurement errors. The psychrometer dry bulb temperature was checked for sensor bias with respect to its corresponding wet bulb temperature by removing the wick from the wet bulb probe and recording any temperature differences between the two sensors. The observed differences ($0.02 \pm 0.04$, $n = 16$) were small enough to be considered insignificant.

Thermistor voltages obtained from three-wire half bridge measurements were converted to temperature values using the manufacturer’s calibration coefficients. The resulting temperature values were recorded and handled as described in section A.1.1.

Although the psychrometer thermistors are shielded from direct radiation by enclosure in a ventilated double walled enclosure, they may still have been affected by reflected radiation impinging from the downward facing opening of the enclosure. The equation (Brock et al. 2001) describing the temperature increase, $\Delta T$, of the probe by radiation heating.
\[ \Delta T = \frac{\pi \cdot \alpha \cdot R_r}{c} \cdot \sqrt{\frac{d}{V_a}} \]  

(A.2)

suggests that the maximum probable temperature increase will be 0.3°C, assuming approximate values of reflected radiation \( R_r \) (50 W m\(^{-2}\)), effective sensor diameter, \( d \) (0.01 m), sensor absorptivity, \( \alpha \) (0.3), flow velocity \( V_a \) (4 m s\(^{-1}\)), and an empirical coefficient \( c \) (8.011 W m\(^{-2}\) K\(^{-1}\) s\(^{-0.5}\)).

An opposing effect may have resulted from the wetting of the sensor by precipitation droplets resulting from ventilation of the psychrometers. The size of droplets from large clouds or drizzle is in the range of 50 to 200 \( \mu m \) (Yangang et al. 1995). These droplets have a terminal velocity \( \dot{w} \) described by the empirical equation:

\[ \dot{w} = -\frac{70}{P} \left( w_0 - e^{\frac{R_0 - R}{R_1}} \right) \]  

(A.3)

where \( P \) is the air pressure, kPa, \( R \) is the droplet radius (\( \mu m \)), and the empirical coefficients \( w_0, R_0, R_1 \) have the values 12 m s\(^{-1}\), 2500 \( \mu m \), and 1000 \( \mu m \), respectively (Stull 2000). Although no quantitative measure of precipitation type was made, observation suggests that drizzle accounted for a significant proportion of the occurrences of precipitation. A typical droplet velocity for drizzle of 1.5 m s\(^{-1}\) is obtained from the 200 mm radius and typical air pressure for the site. This velocity is slower than the ventilation velocity of the psychrometer (4 m s\(^{-1}\)), suggesting that the dry bulb probe may have become wetted by the advected droplets. Therefore, under conditions of fog and drizzle, the psychrometer dry bulb temperature may have underestimated the true air temperature.

**A.2.1.2 Sonic anemometer air temperature**

Speed of sound derived air temperature measurements were obtained from the sonic anemometer used for eddy covariance measurements. The sonic anemometer measures the transit time of a sound pulse over a known distance. This transit time may be converted to air temperature \( T_{su} \) as:
Where $L$ is the sonic path length in meters, and $t_{up}$ and $t_{down}$ are the transit times, in seconds, of sound pulses travelling in opposing directions along this path. Errors caused by cross path wind velocity (Kaimal & Gaynor 1991) can be accounted for by the equation:

$$T_s = T_{so} - \frac{v_n^2}{403}$$  \hspace{1cm} (A.5)

Addition errors caused by the actual path length $L_m$ being different from $L$, from effective path shorting by liquid water of depth $L_w$ on the sonic transducer faces, and changes in flow rate along the sonic path between samples $\Delta V_d$ have also been derived in Appendix B. Incorporation of these additional corrections produces the modified equation:

$$T_s = \left( \frac{L_m - L_n \cdot 335}{1482} \right)^2 \frac{L_m}{L} T_{so} + \frac{V_n^2}{403} + \frac{\Delta V_d \cdot T_{so} \cdot \sqrt{2}}{20.07}$$  \hspace{1cm} (A.6)

This equation may be further modified, (Kaimal et al. 1991; Schotanus, Nieuwstadt, & De Bruin 1983) to obtain the true air temperature, $T$, by incorporating the density effects of water vapour on the corrected sonic temperature:

$$T = \left( \frac{L_m - L_n \cdot 335}{1482} \right)^2 \frac{L_m}{L} T_{so} + \frac{V_n^2}{403} + \frac{\Delta V_d \cdot T_{so} \cdot \sqrt{2}}{20.07} - 0.00032 \cdot \bar{T} \cdot q$$  \hspace{1cm} (A.7)
The mean value of \( V_n \) will typically be no more than about 3.5 m s\(^{-1} \) so that the second term of equation A.7 will be on the order of 0.03 C. Values of \( \Delta V_d \) can be assessed from typical sample to sample velocity changes, figure A.8, and from the sub-sampling rate for the anemometer (48 sub-samples/output for the Gill Solent R2). This would mean that for a worst case sample to sample velocity change of ±0.15 m s\(^{-1} \) the value of \( \Delta V_d \) would be approximately 0.003 m s\(^{-1} \). The third term in equation A.7 would have an instantaneous value of approximately 0.0025 C. The fourth term in equation A.7 is more significant and may be on the order of -1 C for typical values of \( T \) and \( q \).

While the last three terms of equation A.7 are quite small, they become more important when considering the use of sonic temperature for eddy covariance purposes. Under some conditions, the most important term in determining the true air temperature will be the ratio of the liquid water adjusted actual path length to the assumed path length, \( \bar{L} \). Liquid water layers of 2 mm are easily obtained in the laboratory and the accuracy of the path length may be on the order of 0.5 mm. Combined, these errors could result in an overestimation in the measured sonic temperature of 2%, which would result in a temperature error of up to 6 C. It is
unlikely, however, that a liquid water layer of 2 mm thickness would remain on the transducer faces for extended periods so that the actual error in $T_s$ would probably be less than this worst-case scenario.

Additional problems of temperature response non-linearity caused by sonic transducer behaviour have also been observed by Mortensen et al. (Mortensen & Hostrup 1995). As a test, the correction applied to the University of Edinburgh BOREAS data set (Massheder 1999), was applied to the Griffin data set.

$$T = -3.7263 + 1.4429 \cdot T_s + 0.0111 \cdot T_s^2 - 0.0022 \cdot T_s^3 + 0.000005 \cdot T_s^4 + 0.000002 \cdot T_s^5$$

(A.8)

In addition, upon comparing the entire data set, it became apparent that consistent non-linear relationships between $T_s$ and $T_{psy}$ only existed when the data was separated by sonic anemometer probe. Three different anemometer probes were employed during the experiment, one of which required a transducer replacement during the experiment. Equations were developed for each probe to adjust the sonic temperature using $T_{psy}$ as a reference. Two different linear relations between $T_s$ and $T_{psy}$ were found above and below a critical temperature $T_\alpha$ of approximately 2 C. However, only one of the sensors exhibited strong differences in this relation at lower temperatures table A.2.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Condition</th>
<th>offset</th>
<th>slope</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>20</td>
<td>$&lt; 2$ C</td>
<td>-0.52</td>
<td>0.79</td>
<td>0.88</td>
</tr>
<tr>
<td>20</td>
<td>$&gt; 2$ C</td>
<td>-1.29</td>
<td>1.087</td>
<td>0.99</td>
</tr>
<tr>
<td>83</td>
<td>All</td>
<td>-1.16</td>
<td>1.047</td>
<td>0.98</td>
</tr>
<tr>
<td>58</td>
<td>Before 21/4/2000 8:00</td>
<td>-1.74</td>
<td>1.204</td>
<td>0.97</td>
</tr>
<tr>
<td>58</td>
<td>After 21/4/2000 8:00</td>
<td>-1.07</td>
<td>1.158</td>
<td>0.97</td>
</tr>
</tbody>
</table>

The sonic temperature data under non-precipitation conditions, uncorrected and corrected, using the BOREAS and $T_{psy}$ corrections, are presented in figure A.9 and the linear regression coefficients are given in table A.3. It is obvious from the results that the BOREAS correction obtained from equation A.9 is inappropriate for this data set. As expected, the psychrometer correction produces a better
relationship; the slope of 0.96 may have been further improved by manually excluding positive outliers caused by precipitation below the level measurable by the rain gauges.

![Graph showing comparison of psychrometer with sonic derived air temperatures under non-precipitation conditions.](image)

**Figure A.9** Comparison of psychrometer with sonic derived air temperatures under non-precipitation conditions. Uncorrected, BOREAS correction, and manual psychrometer adjusted sonic temperatures are presented. The line shown is the 1:1 slope.

**Table A.3** Linear regression coefficients of $T_{\text{psy}}$ against $T_{\text{sonic}}$ for the data presented in figure A.9.

<table>
<thead>
<tr>
<th></th>
<th>Offset</th>
<th>slope</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Uncorrected</td>
<td>1.648</td>
<td>0.848</td>
<td>0.96</td>
</tr>
<tr>
<td>BOREAS correction</td>
<td>-2.269</td>
<td>1.177</td>
<td>0.96</td>
</tr>
<tr>
<td>$T_{\text{psy}}$ correction</td>
<td>0.496</td>
<td>0.960</td>
<td>0.97</td>
</tr>
</tbody>
</table>

As shown in equation A.8, the effect of the presence of precipitation on transducer faces can have a large effect on the mean temperature measured by the sonic anemometer. This effect can be seen in figure A.10, which shows the distribution of difference between sonic and psychrometer temperatures for different rainfall rates.
Figure A.10 Distributions of the difference between sonic and psychrometer dry bulb temperatures ($T_s - T_{psy}$) for different precipitation rates.

Only a small temperature difference between sensors is observed under non-precipitating conditions. Under precipitation conditions, a positive offset is observed in the sonic anemometer data of approximately 1.5°C. This offset corresponds to an average liquid water depth of about 1.6 mm on the transducer faces. It is notable that this offset remains relatively constant for different rainfall rates.

Figure A.11 Distributions of the difference between sonic and psychrometer dry bulb temperatures ($T_s - T_{psy}$) for different sonic anemometer probes under precipitation and non-precipitation rates.

When the same data set was grouped by anemometer probe, figure A.11, it appeared that different probes had different positive offsets under precipitation.
conditions. It is suspected that the cause of this difference is a result of slight variations in anemometer construction or installation as the physical cause of the temperature difference will be the same for all three probes. While the effect of precipitation on the mean temperatures measured with the sonic anemometer is observable, its correction remains difficult. While it appears likely that a correction could be applied to the mean temperature if periods of transducer wetting are identified, identifying these periods would be difficult.

The sonic anemometer/thermometer was installed on the top of the eddy covariance tower at an height of 15.2 m. Data were gathered and recorded as described in chapter 3. The anemometers speed of sound output was converted to temperature using equation A.8, assuming that \( (L_m - L_w) = L \) because of the difficulty in determining the effect of sensor distortion and thickness of liquid water on the transducer faces. Further corrections for probe transducer response, as previously detailed, were then applied to obtain the air temperature measurements.

A.2.1.3 Fast response air temperature

In association with the eddy covariance system, fast response measurements of air temperature were collected using fine wire thermocouples and dedicated electronics. The thermocouples consisted of a soldered fine-wire copper/constantan junction attached to type T thermocouple wire and amplifying electronics are described by (van Asselt et al. 1991). Two of these thermocouple probes were mounted in close proximity (0.3 m, 0.45 m) to eddy covariance sonic anemometer. Though no rigorous calibration was performed, signals were checked for reasonable values in the laboratory before field deployment. These probes were not shielded from solar radiation and small errors may have been incurred under conditions of high radiation and low wind speed (Jacobs & McNaughton 1994). Fast response thermocouple signals were converted from voltage to degree C and recorded as described in Chapter 3.
A.2.1.4 Temperature profile

Air temperature profile

A vertical profile of air temperature measurements was made at 10 levels within and above the canopy using thermocouples (type T) referenced to a thermistor (Betacurve, Betatherm, Shrewbury, Ma, USA). The thermistor and thermocouple reference junctions were embedded in an aluminum heat sink and encased in an insulated housing to minimise potential temperature gradients between the thermocouple reference junctions and the thermistor, figure A.12. Radiation shields for the thermocouple junctions were constructed of two 12 cm, reflective, plastic disks mounted horizontally with a separation of 1 cm between the disks. The disks were fixed to the profile tower at the desired heights and a thermocouple was mounted 1 cm below the northern edge of the disk assembly. This position reduced direct solar irradiation of the thermocouple junction for all times except during early morning and late evening during summer periods, figure A.13. Assuming the thermocouple was shaded from direct solar radiation, calculations using equation A.2 assuming a thermocouple diameter of 1 mm and wind speed of 0.5 m s⁻¹ would result in temperature overestimations similar to that of the psychrometer (i.e. 0.3°C).
Only small offset errors in thermocouple measurements were observed and were attributed to drift in the reference thermistor. Infrequent field checks of the reference thermistor were performed by inserting a thermocouple junction into hot and cold water-baths and referenced to a mercury in glass thermometer. Errors based on these checks were less than 0.3°C. As a second check, the thermocouple readings were compared with those of the psychrometer.

The thermocouple and thermistor signals were converted from millivolts to deg C and half hour averages were recorded as described in chapter 3.

Similar to the psychrometer, both the air temperature profile and fast response thermocouple types may have been affected by wetting of the sample junction by fog, light drizzle, or condensation. Such occurrences will have reduced the sensed temperature with respect to the true air temperature. No quantitative assessment of this effect is possible without knowing the conditions (amount of water on sensor and vapour pressure deficit) for any particular wetting incident.

**Soil temperature profiles**

Soil temperatures were measured at 5, 6, 8, 12, 20, and 30 cm below the surface. The same type of thermistor used for the air temperature profile was used to
measure the 30 cm soil temperature. This thermistor also acted as the reference temperature for the reference junctions of the thermocouple temperature measurements made at the other levels. The thermistor and thermocouple sensors were embedded in the wall of a 25 cm diameter PVC pipe. The exposed sensors were covered in epoxy to prevent water intrusion and the centre of the pipe was filled with foam insulation to prevent air circulation. The soil temperature profile probes were calibrated against a mercury in glass thermometer (0.1 C resolution) in a water bath, over the temperature range of 4 to 20 C.

Four soil temperature profile probes were installed in close proximity to the profile tower, figure 3.3, two in the open canopy and two in the closed canopy plot locations. In each plot, one probe was installed in unploughed soil and one is a plough trough. The thermocouple and thermistor signals were converted from millivolts to deg C and half hour averages were recorded using a data logger (CR21X, Campbell Scientific, Logan Utah, USA).

Because of instability of the thermistor sensor readings in some of the profile probes, it was necessary to adjust those temperature readings for zero drift. In these circumstances, die averages of the top-level thermocouple readings were compared with the soil surface platinum RTD sensor averages to obtain the zero offset. Post processing of the values was carried out any calibration adjustments to the data and to obtain a final value. A final value of T was obtained by an equally weighted average of quality-controlled results.

A.2.1.5 Near surface soil temperature

The near surface (0 - 5 cm) soil temperature was measured using a platinum resistance temperature device, RTD (RS components, Corby, UK). Platinum RTDs have a known resistance of 100 Ohm at a temperature of 0 degrees C. These sensors require calibration but are quite stable. The RTDs employed came in the form of a 4 cm by 1 cm adhesive patch, which was applied, to a 5 x 2.5 x 0.5 cm aluminum plate. The aluminum plate acted to integrate, over a 5 cm depth, the temperature in contact with the PRT. The side of the aluminum plate to which the PRT was
affixed was potted with approximately 0.5 cm of epoxy adhesive to reduce the
direct effects of soil temperature on the PRT.

The PRTs were calibrated, following procedures outlined in the Campbell Scientific
manual, in the laboratory using a water bath and referencing the sensors to mercury
in glass thermometer (0.1 C resolution). The sensors remained quite stable in the
field and zero adjustment was not necessary. One sensor did produce erroneous
results for a period as the result of its half bridge circuitry being immersed in
rainwater. Upon discovery and drying the sensor recovered its normal operation.

Only two surface layer soil temperature sensors were deployed. These sensors were
installed in the open canopy and closed canopy plot locations in close proximity to
the profile tower, figure 3.3. Both sensors were installed in a region of unploughed
soil, and in close proximity to one of the soil temperature profile sensors. The
probes were inserted vertically so that their length spanned a 5 cm soil depth.

The resistance of the PRTs was measured using a high precision reference resistor
in a four wire half bridge. The resistance was converted to temperature using the
laboratory calibration results and the conversion procedures available in the data
logger. Data were handled as described in chapter 3. Signals were checked by
comparison with other soil temperature signals and by limitation to reasonable
values (-5 To 30 C) before being used for further data analysis.

A.2.1.6 Bole temperature

Bole temperature was measured on four trees near the profile tower. The
thermocouples measurements were referenced to a thermistor in the same manner as
those for profile air temperature. Thermocouple junctions were waterproofed and
inserted at two levels (0.75 and 1.7 m) corresponding to two air temperature profile
levels. At each level two probes were inserted; one to a depth corresponding to the
centre of the tree and the second approximately 1 cm below the tree’s bark. To
insert the probes, a hole the size of the thermocouple wire was drilled horizontally
into the tree. A probe was inserted to the full depth of the hole and the hole
backfilled with silicone sealant. The second probe was then inserted to the
specified depth. The thermocouple wire leading to the probes was not insulated; the
canopy was closed at these locations so that conduction of radiation heating along the lead wires was thought to be small. The calibration and signal handling of these measurements was identical to those of the profile temperature sensors. Similar to the soil temperature profile measurements, it was necessary to make zero drift adjustments to some of the bole temperature measurements. Zero offsets were determined from comparison of the diel averaged time series of bole and canopy air temperatures.

A.2.1.7 Common techniques

Recorded air temperature values were retrieved weekly and maintained in a database of results. Post processing of the values was carried out to apply any calibration adjustments to the data and to obtain a final value. No significant zero drift was observed in the air temperature sensors and no corrections for zero drift were applied. The half hour values of air temperature were quality checked by comparison and limited to reasonable values of between -20 C and 30 C before being used for further processing and analysis. A final value of \( T \) was obtained by an equally weighted average of quality-controlled results for sensors at equivalent heights.

A.2.2 Sensor comparison

Despite their spatial separation, the best sensor agreement is observed between the above canopy psychrometers on the eddy covariance and profile towers (figure A.14 top left panel, \( T_{\text{psy, eddy}} = -0.159 + 1.007 \, T_{\text{psy, up}}, \, R^2 = 0.99 \)). The comparison, between the vertically separated psychrometers on the same tower, top right panel, exhibits greater variability (\( T_{\text{psy, low}} = -0.442 + 1.033 \, T_{\text{psy, up}}, \, R^2 = 0.98 \)), as can be expected from the temporal variability of the temperature profile variability, see section 4.9.
The second best comparison was between the co-located psychrometer and thermocouple measurements, figure A.14 bottom right panel. There existed a slight non-linearity in relation between these two sensors but still a reasonable correlation ($T_{TC} = 0.283 + 1.001 T_{psy\_up}, R^2 = 0.99$).

Comparison of the aspirated psychrometer dry bulb and non-aspirated, shielded thermocouple, figure A.15, reveal a relative increase in the thermocouples' temperatures relative to those of the psychrometers for increasing levels of impinging radiation. Similar patterns were observed when the data were viewed at different temperature ranges, and wind speeds. It is unlikely that this difference is the result of calibration difference, as the observed 0.5 degree difference over the 3 degree grouping used in figure A.15 would correspond to a 16% greater temperature response for the TC measurements, which is inconsistent with the general relationship between the two sensors. The close agreement between the two sensors at low radiation supports the hypothesis that radiation is the cause.

Although the calculations from equation A.2 suggested that radiation effects were similar for both psychrometer and thermocouple measurements, this finding points to errors in the assumptions or application of equation A.2. This implies possible
positive temperature biases of approximately 0.5 degrees under high radiation conditions. Furthermore, these biases would be greater for measurements above the canopy or in the upper canopy where reflected radiation was the greatest.

Figure A.15 Temperature difference between aspirated psychrometers and non-aspirated thermocouples in response to reflected radiation at two heights. Data are separated by precipitation (triangles) and non-precipitation (circles) conditions. Non-precipitation data are for the air temperature range 12 to 15 degrees while precipitation data are for the range 9 to 12 degrees. Error bars represent +/- 1 standard deviation.

Despite the numerous corrections applied to the sonic temperature, the poorest comparison is between the psychrometer and the sonic anemometer derived air temperature, (figure A.14, bottom left panel, $T_{\text{sonic}} = 0.836 + 0.951 T_{\text{psv_up}}$, $R^2 = 0.95$), which exhibits considerable scatter and a remaining 5% underestimation. The cause of this poor comparison is probably related to the inability to properly account for the effects of precipitation on the sonic derived temperature values.

A.3 Humidity

There are two categories of humidity classification, measures of the amount of water vapour in the air (vapour pressure, $e$, absolute humidity, dew point temperature, mixing ratio, specific humidity, $\rho_v$, concentration, $Q_v$) and measures of the proportion of water vapour relative to the maximum possible (relative humidity, $RH$, wet bulb temperature, $T_w$). In this experiment, two methods were employed to measure atmospheric humidity: psychrometric, which provides wet bulb
temperatures and thus a proportionate measure of humidity, and spectroscopic methods, which provide absolute measures of humidity as a concentration.

A.3.1 Measurement methods

A.3.1.1 Psychrometric

Psychrometric measurements of humidity were obtained using a ventilated dry bulb/wet bulb psychrometer (VP1/TM1, DeltaT Instruments Ltd., Burwell, UK). Psychrometer deployment and measurement specifics associated with the psychrometer dry bulb temperature sensors are described in section A.2.1.1.

Psychrometer maintenance was conducted weekly or biweekly depending upon time of year and previous state of maintenance. Ventilation fans were cleaned or replaced as necessary and water reservoirs were topped up with distilled water. Cotton wicking (Russell Scientific, Dereham, UK) was replaced if signs of contamination or drying existed. Water reservoirs occasionally developed cracks because of age and were replaced as needed.

Part way through the experiment it was determined that construction of the psychrometer housings was hindering proper maintenance. The psychrometers were re-designed to allow easier replacement of wicks and fans. This entailed making the air inflow duct removable to allow fast access to the wicks, attaching the radiation shield to the body of the psychrometer instead of the vent fan. The vent fan was then detached from the body of the psychrometer and attached to a removable plate to allow replacement of the fan without requiring removal of the entire sensor from its installation position.

Several sources of bias error can exist in the psychrometric determination of atmospheric humidity, these include; sensor calibration bias, addition of energy through radiation or conduction, and improper ventilation. As described in section A.2.1.1, dry and wet bulb temperatures were occasionally checked for sensor bias and no significant wet bulb temperature bias was observed. Sensor ventilation was maintained though no tests as to its effect were conducted. It is assumed that ventilation errors were small. The effects of radiation and heat conduction along the
wet bulb temperature probe supports were investigated in a laboratory based experiment. In this experiment (figure A.16), relative humidity values determined from a psychrometer were compared with values obtained from a new capacitive sensor (HMP35C, Campbell Scientific, Logan Utah, USA). It was assumed that the capacitive sensor was stable over the period of the experiment though not that its absolute value was accurate to more than ±3%.

![Figure A.16 Relative humidity comparison of psychrometer (black) with humicap sensor (white) done in laboratory.](image)

The two sensors were compared under three experimental conditions. The first condition was the state of the sensor as deployed in the field, (point A to C in figure A.16). This period was characterised by the length of wick covering the wet bulb (~1.5 cm) while a longer wick (~4 cm) was used from point C to the end of the test. The longer wick length was employed to determine $T_w$ biases caused by conductance of heat along the wet bulb sensor probe. For the longer wick, only the period following point D was assumed to have reached equilibrium. To determine if significant radiation effects existed, a 300 W lamp was placed 2 m below and 0.3 m to the side of the psychrometer and ventilated with a fan to minimize convective heating of the probes (points B to C in figure A.16). Tests with the lamp showed that at this distance the psychrometer temperature probes would have been exposed to between 40 and 50 W m$^{-2}$, which is equivalent to a sunny day at the site. The
capacitive sensor was shielded from this radiation. A comparison of the radiation effect (rank sum test, A to B vs B to C) suggested a small (0.16 °C) but significant (P < 0.001) difference between the two sensors because of radiation; this effect was about half that predicted by equation A.2. The effect of conductance along the wet bulb sensor probe was more substantial with a longer wick causing a significant decrease (3.6%) in relative humidity; equivalent to a 0.6 °C decrease in wet bulb temperature.

It was assumed that this error was prevalent throughout the experiment, as a similar wick length was used throughout. Comparison of the difference between water vapour concentrations, obtained from psychrometer and IRGA measurements, at different levels of D suggested that the bias errors were greater than those observed in the laboratory. The error caused by conduction was addressed by altering the psychrometric constant, γ, in the equation for calculating vapour pressure (equation 3.11) from a value of γ = 0.00066 to a value of γ = 0.00095.

\[ e = e_s - \gamma \cdot P \cdot \left(1 + 0.00115 \cdot T_a\right) \cdot \left(T_a - T_w\right) \]  

(A.9)

Recalculation of the data using the altered value of γ provided better agreement between psychrometer and IRGA under conditions of high D, figure A.17.

Further improvement in agreement could have been obtained with increased values of γ, but such an increase was not justified based on the laboratory experiment. The observed divergence of values at value of D near zero may be an indication of the cooling effect of advected precipitation on dry bulb temperature.
A.3.1.2 Infra Red Gas Analyser - closed path

In addition to the psychrometric readings, humidity concentration measurements were obtained by spectroscopic methods using a differential infrared gas analyser (IRGA) (model LI6262, Licor Instrument Co., Lincoln Nebraska, USA). An IRGA measures the amount of radiation absorbed at a specific wavelength as light of that wavelength passes through a column of sample air, see figure A.18. The amount of radiation absorbed in the sample column is compared with the amount of radiation absorbed in a reference column of air that contains no water vapour. The difference in radiation passing through the air columns is proportional to the amount of absorbing gas in the sample cell.

The IRGA model used in this experiment has a sub-sampling rate of 500 Hz, which is averaged to a minimum sample output rate of 0.1 Hz. The fastest sampling rate, which was used with the eddy covariance IRGA, provided a peak-to-peak noise level of 0.006 kPa at a mean vapour pressure of 0.2 kPa. Longer averaging times provide improved noise characteristics. The profile IRGA used a 1 second averaging time with a corresponding noise level of 0.002 kPa.

The IRGA has a non-linear response to changes in sample gas water vapour concentration. The non-linear output of the sensor requires the manufacturer’s fifth
order polynomial calibration to be applied as well as adjustments for cell pressure, cell temperature, and corrections for pressure broadening and dilution. These corrections are described in the manufacturers handbook (Licor 2002).

Figure A.18 Schematics of IRGA deployments for eddy covariance (top) and profile (bottom) measurements of CO₂ and H₂O. Thick solid lines represent sample and reference air flow paths. Thin solid lines represent other air flow paths. Thick dotted lines represent light paths. Thin dotted lines represent other signals. Circled characters represent: filter (F), temperature sensor (T), solenoid (L), pressure sensor (P), manual flow meter and flow valve (V), mass flow controller and meter (M), pump (U), radiation source (S), radiation detector (D), signal electronics (d). Circled characters with greyed backgrounds indicates signals for which run mean values were recorded. The greyed rectangle represents a weather resistant enclosure and the white rectangle represents the IRGA enclosure. Compressed gases are shown with their gas specific labels.

Of the two IRGAs used in this experiment, one sensor was deployed in association with the eddy covariance measurements while the other sensor was used to measure within and above canopy profiles of water vapour and CO₂. The eddy covariance
IRGA was installed as part of the Edisol flux measurement system (Moncrieff et al. 1997) at the base of the eddy covariance tower. Approximately 18 m of polyethylene coated aluminum tubing lined with ethylene copolymer (dekoron/decabon, Eaton Corporation, Aurora Ohio, USA) was used to transport a gas stream from an inlet near the sonic anemometer (0.2 below and 0.05 m north). A filter was placed at both ends of this tube; the filter at the inlet consisted of a 48 mm polypropylene filter housing (Cole Parmer, London, UK) containing a replaceable 1 μm PTFE filter disk (Whatman, Maidstone, Kent, UK). The gas flow rate through this tube (100 cm$^3$ s$^{-1}$) resulted in a signal delay of approximately 6 seconds under conditions of low relative humidity. This IRGA system was installed for the entire five years of the experiment.

The profile system IRGA was placed at the base of the profile tower. A data logger controlled, solenoid manifold was used to select between the eight incoming gas sample lines. Sample lines were of the same material used for the eddy covariance sample tube. An, inverted, screen-covered funnel was placed at each sample line inlet and a single filter housing, similar to that at the inlet of the eddy covariance IRGA, was placed prior to entry into the IRGA. This sensor was only deployed for three and a half years and was removed in the final year as a means to reduce power consumption.

Maintenance to both systems included filter replacement, sample pump replacement, annual replacement of internal chemical scrubbers, reference gas maintenance, and calibration adjustments.

Because of the higher flow rate (100 cm$^3$ s$^{-1}$) of the eddy covariance system, its sample inlet filter was changed approximately every ten days. The sample line filter on the profile system was changed every one to two months. The use of replaceable filter disks instead of encased filters provided a cost savings of approximately £1000. The downstream sample line filter of the eddy covariance system and the filters on the reference gas lines of both systems (Gelman Arcovent, Pall Corp, Portsmouth, UK) were replaced every six to twelve months.
A mains powered pump was used for the profile system. It was only necessary to replace this pump once over the duration of the experiment. In contrast, the 12V DC pumps in the eddy covariance system exhibited a mean time to failure of about 7 months. Generally, the DC pumps were disassembled, cleaned, and reinstalled after which they would last for approximately half of their initial lifetime. While recycling the DC pumps proved useful, switching to AC powered pumps would have saved approximately £300 on the purchase of pumps. However, the greater consumption of power would have taxed our power generation system, reducing these cost savings.

For the first five months of the project, molecular sieve was used to purge the reference cell of water vapour (and CO$_2$). Molecular sieve was selected because it was easy to use and provided the necessary sample absorption characteristics. It was found that the 10-day maintenance interval was insufficient for purge chemical upkeep and that small leaks in the reference cell gas re-circulation system caused significant data loss. As an alternative, compressed nitrogen gas was used as the purge for the reference cell during the remainder of the experiment. Initial cylinders were compared against chemicals and differences in zero offsets were less than the resolution of the sensor. Compressed nitrogen gas cylinders lasted from between one and 28 weeks with the average cylinder lasting 12 weeks; variations in cylinder longevity were related to connection seals, and purge flow rates.

Zero drift of the water vapour signal was checked every visit and adjusted as necessary for both IRGAs. Automated water vapour zero checks of the eddy covariance IRGA were carried out at midnight every two days using either the nitrogen or CO$_2$ span gas calibration gas flow. Insufficient equipment was available for automated calibration of the profile Licor. Compressed nitrogen and air were used as the dry air source for the profile IRGA offset checks. Humidity values obtained from psychrometers were employed for signal offset cross checking and determination in the data post processing procedures.

Because of the difficulty in obtaining an accurate gain calibration, water vapour span calibrations and adjustments in the field were carried out infrequently. In situ span calibrations were carried out primarily in the summer when air temperature
was high enough to allow a range of vapour pressures sufficient to determine sensor span. Otherwise, water vapour span was usually checked and/or adjusted when the sensor was returned for internal purge chemical replacement or other maintenance. A dew point generator (LI610, Licor, Lincoln NE, USA) was used as the reference for span calibrations.

Manual calibration results were recorded in the field logbook and later transferred to a spreadsheet used for calculating calibration factors; automated calibrations were initiated and recorded using data loggers. Recorded calibration values were retrieved from the data loggers and entered in the same spreadsheet used with the manual calibrations. Automated calibration data were retained only if the data logger, IRGA, and calibration system were known to be operating properly.

Several sources of error within the calibration process may have resulted in biased calibration coefficients. These errors include inaccurate reference or span gas concentrations, errors in IRGA cell pressure and temperature determination, and inaccurate determination of IRGA values by the attached data logger. It was not possible to identify the range of bias or random errors associated with these error sources.

Although care was taken both performing and evaluating the Licor sensor calibrations, data sets for both sensors contain some extreme offset values. These values were left in place until it could be determined if they were correct. This determination was made by correcting mean concentration values from both sensors and comparing those values with available or reasonable corresponding atmospheric values. The resulting data set retained calibration offset values that fluctuated by \( \pm 0.3 \text{ mmol mol}^{-1} \). As no continuous gain calibration was available, it was assumed that sensor gain was correct for IRGA water vapour signals. Based on experience with CO\(_2\) gain calibration it is expected that this assumption may have been in error by less than 1%.

Each calibration offset value was classified as either a beginning, middle, or ending calibration. These labels indicated whether the sensor offset was not adjusted (middle), or were adjusted immediately after (ending), or before (beginning) the
calibration of the sensor. Because calibrations occurred periodically, it was necessary to interpolate calibration values to periods corresponding to run values. Values were linearly interpolated between calibrations but were not interpolated to times preceding a beginning calibration or following an ending calibration. The resulting calibration values were applied to the raw recorded data of the eddy covariance system, and were used to correct the linear calibration that had been applied to the run averaged values of the profile IRGA.

A.3.1.3 Infrared gas analyser - open path

Two open path IRGA H2O/CO2 sensors were used in association with eddy covariance flux determination. Both sensors were deployed for only a short period as they were installed for testing and evaluation purposes. For both sensors, the manufacturer's calibration was used in the calculation of humidity. Note that the results of the OP2 sensor are not presented because of the shorter period over which that sensor's signal was recorded.

A.3.2 Sensor comparison

A comparison of water vapour sensor, figure A.19, shows good agreement between the psychrometric values of water vapour concentration, $Q_v$, ($Q_{v_10.7m} = 0.057 + 1.014Q_{v_14.6m}$, $R^2 0.97$, $Q_{v_10.7m} = 0.059 + 0.995Q_{v_7.9m}$, $R^2 0.99$). The reasonable agreement between the IRGA and psychrometer values is expected because of the dependence of the IRGA values thereupon. The IRGA measurements however, exhibit a much greater scatter in its relationship with psychrometric water vapour, with that from the profile IRGA ($Q_{v_10.7m} = 1.529 + 0.950Q_{v\_profile}$ $R^2 0.35$) being greater than that from the eddy covariance IRGA ($Q_{v_10.7m} = 0.096 + 0.985Q_{v\_eddy}$ $R^2 0.94$). Much of this uncertainty can be traced to inaccuracies in the determination of calibration coefficients. The better performance of the eddy covariance IRGA is the result of more frequent calibration checking and greater efforts at calibration quality control expended upon this sensor's data. Offset errors in the profile IRGA were acceptable as its water vapour values were used primarily for determination of profiles referenced to above canopy water vapour values.
Comparison of the water vapour spectra for open and closed path sensors, figure A.20, shows the severe damping of the water vapour fluctuations measured by the closed path sensor (LI-6262) for water vapour concentration oscillations shorter than approximately 10 seconds (0.1 Hz). The open path water vapour sensor and sonic temperature do not exhibit this strong damping but both show noise at high frequencies; with the noise of the open-path water vapour sensor being less than that of the sonic temperature. It appears as though the noise on both IRGAs is
apparent at frequencies an order of magnitude higher than is observed in the sonic temperature signal. This effect does not have direct implications for the measurement of mean quantities with the closed path sensor but does limit that sensors ability to measure higher order statistics and paired statistics.

A.4 Carbon dioxide

Currently, IRGAs provide the simplest method for real-time measurement of atmospheric carbon dioxide content. Other methods of measurement; gas chromatographs, TDL, FID, FTIR, PTR-MS etc. are available but are uneconomical or impractical to implement for continuous use in the field. Therefore, carbon dioxide concentration was measured using the same IRGAs employed for water vapour measurement. A closed path IRGA was used for both eddy covariance and profile measurements.

A.4.1 Measurement

The measurement principle and practise for measuring carbon dioxide with IRGAs is the same as that described for water vapour in section A.5. As with water vapour, both closed path IRGAs accounted for the majority of measurements while and the open path sensors were employed only for a brief period of testing.

A.4.1.1 Closed path IRGA

The primary differences between the measurement of CO₂ and H₂O using a closed path IRGA are associated with the noise characteristics and slight variations in the conversion from detector voltage to signal (Licor 2002). For the 0.1 and 1.0 second signal output averaging as used in the eddy covariance and profile systems, respectively, the peak to peak noise levels are specified as 1.0 and 0.3 ppm at a mean concentration of 350 ppm. Other potential errors associated with the measurement of CO₂ are identical to those previously described for water vapour.

The deployment, maintenance of and measurement with the closed path IRGAs is described in section A.5. Compressed nitrogen cylinders were used as reference and calibration zero gas as described for water vapour measurements in section A.5. The calibration span gas was purchased in uncalibrated compressed air cylinders.
The uncalibrated span cylinders were calibrated against a known gas concentration using an IRGA. Two independent CO₂ reference sources were employed; a calibrated cylinder of compressed air (Rivora, Italy), and a gas mixing pump (Wosthoff, Bochum Germany). The compressed air cylinder (512 ppm) was obtained as part of an effort by Euroflux to standardize calibration procedures, while the gas-mixing pump allowed user specified CO₂ concentrations to within 1 ppm.

A.4.1.2 Open path IRGAs

Two open path IRGA carbon dioxide measurements were used in association with eddy covariance flux determination. Both sensors were deployed for only a short period as they were installed for testing and evaluation purposes only. For both sensors, the manufacturer's calibration was used in the calculation of CO₂ concentrations.

A.4.2 Sensor comparison

Because of the frequent drift in the mean CO₂ signals for the eddy covariance and profile IRGAs, a comparison of the mean signals is not shown. They were similar or worse than the corresponding comparison shown for water vapour concentration in figure A.19. As the mean value of CO₂ concentration was not critical for either the measurement of CO₂ exchange or within canopy profiles.

Figure A.21 Comparison of open (LI-6262) and closed sensor path (LI-7500, OP2) carbon dioxide concentration spectra for three stability regimes. Sonic temperature spectra are plotted for comparison.

A comparison of the CO₂ spectra for open and closed path sensors, figure A.21, does not show the severe damping of the closed path sensor signal at medium to high frequencies that was observed for water vapour. Additionally, unlike the
observations on water vapour, high frequency noise is apparent in only one of the open path sensors and is not significant in the closed path sensor.

A.5 Wind speed

A.5.1 Measurement methods

A.5.1.1 Sonic anemometer

Sonic anemometers measure wind velocity as the difference in the speed of sound pulses travelling in opposite directions along the path between the two transducers that emit the pulses (1990; Coppin & Taylor 1983; Kaimal & Businger 1963).

\[
U = \frac{L}{2}\left(\frac{1}{t_1} + \frac{1}{t_2}\right)
\]  

(A.10)

Sonic anemometers are generally more accurate than cup anemometers but their high cost prevents the use of numerous sensors. The sonic anemometer used in this experiment was deployed primarily for the purpose of eddy covariance measurements, but also provided high-resolution measurements of wind speed and direction. The sensor was mounted, on a retractable arm, at a height of 15.5 m (0.5 m above the top of the tower) and approximately 1.5 m to the east-southeast of the eddy covariance tower. The anemometer was initially levelled to gravitational horizontal using a bubble level, but on 10/10/1998 was levelled to more closely correspond with the terrain slope.

No regular maintenance was required of the sonic anemometer. However, twice during the experiment it was necessary to exchange the sonic probes because of failure of a sonic transducer. The removed probe was replaced with a backup sensor and the faulty sensor repaired and recalibrated by the manufacturer.

The sonic anemometer wind components were recorded at 20.833 Hz, and mean half hour wind speeds were calculated using EdiSol software. The manufacturers calibration values were used in converting the recorded signals to velocities. This calibration was a simple conversion from cm s\(^{-1}\) to units of m s\(^{-1}\). However, implicit
Sonic anemometer wind speeds were recalculated during post processing to account for the most appropriate sensor coordinate rotation. Reprocessed half-hour values of wind speed were quality checked for reasonable values (0 to 30 m s$^{-1}$) and by comparison with cup anemometer values.

### A.5.1.2 Cup anemometer

Prior to installation of cup anemometers in the field, cup anemometers (pulses and analogue) were calibrated, using a pitot tube anemometer reference sensor, in a wind tunnel located at the University of Edinburgh. The pitot tube differential pressure was measured using a pressure sensor (FCO 40, Furness Controls Ltd., Bexhill-on-Sea, UK). Some of the calibrations exhibited linear calibrations to the lowest level measured but did not have a zero intercept. Upon reanalysis, the calibration slope offset was well correlated with the date of calibration. This suggested an installation dependent offset in the calibration standard. Large variability in the regression offset of sensors calibrated more than once supported this assumption. The offset was assumed to be related to the standard sensor so that the slope of the non-zero intercept regression was used while the offset was ignored. This provided more reasonable field results that did not contain either positive or negative wind speeds at very low wind conditions.

This linear relation, however, often failed at very low wind speeds because of the frictional forces of the anemometer resisting the forces imposed by very light winds. For the cup anemometers installed above the canopy the problem of anemometer cup stalling was less important as the mean wind speed was typically above the cup's stall speed. For average wind speeds between 0 and 0.3 m s$^{-1}$, there was no significant difference between sonic and cup anemometer wind speed readings. However, from sensor comparisons, it is noted that this problem was more
prevalent for cup anemometers mounted within the canopy because of lower wind speeds.

![Graph 1](image)

**Figure A.22** Top Panel: Cup anemometer response curves to wind elevation angle for ideal and typical cup anemometers (Brock et al. 2001) and for the Vector anemometer used in this experiment. The Vector anemometer gain function is also given (filled triangles). Top Panel: Frequency distribution of wind elevation angles for three conditions of wind direction variability.

![Graph 2](image)

**Figure A.23** Cup anemometer correction factor estimated from flow elevation and anemometer response curve, and percent of readings beyond +/- 40 deg elevation angle as a function of wind direction variability.
Additionally, over-speeding errors (Frenzen 1974; Papadopoulos et al. 2001) caused by the non-cosine response of a cup anemometer to wind elevation angle and also by their non-linear response to wind speed, i.e. cup anemometers respond faster to increases in wind speed than they do to decreases in wind speed. Over-speeding due to wind elevation is a static characteristic dependent on flow while response is a result of the sensor’s distant constant which can be related to anemometer design and air density. Based on the attack angle distribution of velocity, and a cup anemometer response curve for the model anemometer used in this experiment (Albers, Klug, & Westermann 2000), figure A.22, it is estimated that an underestimate of only about 1 percent would be incurred by a cup anemometer installed above the canopy, figure A.23.

Though some minor failings exist, cup anemometers provided a robust and inexpensive measure of run averaged wind speed values under most microclimatological conditions. The manufacturers specifications state an accuracy of 0.1 m s\(^{-1}\) for wind speeds up to 10 m s\(^{-1}\) and 1% for higher wind speeds. The stall speed of the cup anemometers was specified as 0.15 m s\(^{-1}\). Systematic calibration errors, assessed from the deviation from the calibration curve, were translated into maximum potential gain errors of -5 to +10%.

A single cup anemometer was located on the eddy covariance tower at a height of 12.2 m, Figure 3.2. The sensor was mounted on a horizontal arm 1 m to the east-southeast of the tower. This sensor was used in association with the sonic anemometer and profile cup anemometers for determination of mean above canopy wind speeds.

Within and above canopy profile measurements of wind speed were obtained using a suite of eight, cup anemometers. They were placed at elevations corresponding to the top eight levels of air temperature profile measurements. The cup anemometers on the profile tower were mounted on arms that extended approximately 1 m to the north of the tower. These sensors were used to determine canopy wind profiles and to estimate canopy roughness and zero plane displacement.
Cup anemometers were levelled to gravitational horizontal using a bubble level. Occasional checks were made of cup anemometer levels at which time any build up of algal growth was removed. Within the canopy, any shoot growth that caused obstruction to adjacent cup anemometers was trimmed. Accumulations of snow and ice were removed when observed.

The cup anemometer signals were converted from millivolts (or counts) to m s\(^{-1}\) as was appropriate, and half hour averages were recorded using a data logger (CR21X/CR10X, Campbell Scientific, Logan Utah, USA). Post processing of the values was carried out to apply any calibration changes and quality control. The run mean values of wind speed were quality checked for reasonable values (0 to 30 m s\(^{-1}\)) and by comparison with other anemometers. A final value of above canopy wind speed was obtained by averaging available above canopy values using equal weighting.

**A.5.2 Sensor comparison**

A comparison of all cup anemometer values with corresponding sonic anemometer values, figure A.24, shows improved correlation with both proximity and similarity in height of measurement, table A.4. The 12.2 m cup anemometer on the eddy covariance tower shows the best correlation with the sonic anemometer even though it is 2.3 m below the sonic. Some of the discrepancy in magnitude between the magnitudes of the two, cup anemometers mounted at 12.2 m may be caused by the greater canopy height around the profile tower, see chapter 2.

Some of the spread in the observed relationships between the sonic and various cup anemometers is the result of forest growth, figure A.24. The non-linear vertical profile of wind speeds through and above the canopy results in changes in the relationship between anemometers at different levels because of canopy growth. Much of the additional variability, between the 14.6 m and 7.9 m cup anemometers, within growth years observed in figure A.25 is a result of wind direction effects as will be discussed later.
Figure A.24 Comparison of cup anemometers with sonic anemometer wind speed measurements.

Table A.4 Coefficients of regression of cup anemometers against sonic anemometer.

<table>
<thead>
<tr>
<th>Cup anemometer</th>
<th>Offset</th>
<th>Slope (error)</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>14.6 m</td>
<td>-0.011</td>
<td>0.821 (0.0021)</td>
<td>0.91</td>
</tr>
<tr>
<td>12.2 m EC</td>
<td>-0.033</td>
<td>0.933 (0.0018)</td>
<td>0.95</td>
</tr>
<tr>
<td>12.2 m</td>
<td>0.024</td>
<td>0.724 (0.0021)</td>
<td>0.89</td>
</tr>
<tr>
<td>9.5 m</td>
<td>-0.080</td>
<td>0.554 (0.0021)</td>
<td>0.82</td>
</tr>
<tr>
<td>7.9 m</td>
<td>-0.071</td>
<td>0.476 (0.0021)</td>
<td>0.78</td>
</tr>
<tr>
<td>6.8 m</td>
<td>-0.189</td>
<td>0.374 (0.0021)</td>
<td>0.69</td>
</tr>
<tr>
<td>3.4 m</td>
<td>-0.131</td>
<td>0.122 (0.0021)</td>
<td>0.65</td>
</tr>
<tr>
<td>1.7 m</td>
<td>-0.060</td>
<td>0.039 (0.0021)</td>
<td>0.51</td>
</tr>
<tr>
<td>0.75 m</td>
<td>-0.045</td>
<td>0.025 (0.0021)</td>
<td>0.48</td>
</tr>
</tbody>
</table>
Figure A.25 Comparison of cup anemometer speeds for two profile levels, separated by growing year (June to June).

Figure A.26 Average cup and sonic anemometer wind speeds and differences between sonic and cup wind speeds for sonic wind speeds between 1 and 2 m s⁻¹. Data for the cup anemometer on the eddy covariance tower and the top anemometer from the profile tower are presented.

Tower obstruction (Hojstrup 2000; Wyngaard, Rockwell, & Friese 1985) proved to be greater problem than sensor specific errors in the measurement of wind speed.
Cup anemometer measurements when on the lee side of the tower exhibited speed reductions of from 0.2 to 0.5 m s\(^{-1}\), depending upon the upwind obstruction, figure A.26. While the effect of the eddy covariance tower on the cup anemometer was relatively small, the effect of the profile tower and radiation sensors mounted at the top of that tower had clear effects on the anemometer mounted at the top of that tower. It is believed that the effect of the profile tower on anemometers mounted lower on that tower will be similar to that observed on the eddy covariance tower sensor, though the effect of the canopy on lower layers made it difficult to determine such a relationship for these sensors.

A simple equation for correction of the top profile anemometer for tower shadowing was employed for wind directions between 110 and 230 degrees:

\[
U = U \cdot \left[ 1 + 0.35 \cdot \frac{(60 - |\theta - 170|)}{60} \right]
\] (A.11).

This correction was necessary for the implementation of zero plane displacement and roughness length calculations.

A.6 Wind direction

On very long time scales, the behaviour of variations in wind direction may be considered as random (Schulz, Schulz, & Trimper 2001). On shorter time scales, wind direction is organized by synoptic, meso, and micro scale variations in atmospheric processes (Davies & Thomson 1999). Conversely, wind direction variability can be closely related to wind speed at low wind speeds, while at high wind speeds may be related to surface characteristics. The prediction of wind direction can be obtained from synoptic forecast models but is beyond the scope of this thesis. Instead, insight is gained on how the existing wind directions and wind direction variability may have been affected by the particular environment associated with this experiment.
A.6.1 Measurement methods

Wind direction measurements were obtained from both a wind vane and a sonic anemometer. Both the wind vane and sonic anemometer were located on the eddy covariance tower, though at different levels.

The wind vane provided direct measurements of wind direction while those from the sonic anemometer were derived from the three orthogonal wind components output by that sensor. Neither the wind vane nor the sonic anemometer signal outputs were calibrated; the manufacturers calibration values were relied upon for both sensors. For both sensors, wind directions are given in units of degrees with 0 degrees representing north and 90 degrees representing east.

A.6.1.1 Wind vane

Wind direction measurements were obtained from a wind vane installed at the 12.2 m level on the eddy covariance tower. This sensor was oriented towards north. The wind vane reference and potentiometer output signals were converted to a run mean wind direction and wind direction standard deviation using the signal averaging routines available in the data logger (CR10X, Campbell Sci. Logan Utah). The half hour values of wind direction and wind direction standard deviation were quality checked for reasonable values (0 to 360 degrees) and by comparison with the sonic anemometer wind direction values.

A.6.1.2 Sonic anemometer

A description of the sonic anemometer deployment and use is given in chapter 3. Run mean wind directions were calculated near-real time from the sonic anemometers horizontal wind components using EdiSol software. During post processing, sonic anemometer raw data were recalculated to account for changes in anemometer orientation and to obtain wind direction standard deviation values. Wind direction measurements from the sonic anemometer were calculated from the orientation angle of the sonic probe and the unrotated, orthogonal, horizontal velocity components. The half hour values of wind direction and wind direction standard deviation were quality checked for reasonable values (0 to 360 degrees),
by comparison with the wind vane and by review of the component velocities of the sonic anemometer.

### A.6.2 Sensor comparison

A comparison of all five years of mean wind direction values from the wind vane and sonic anemometer, figure A.27, suggested a close correspondence between the two signals. The slight offset indicates a 7-degree discrepancy in the orientations of the sonic and wind vane, with the sonic’s orientation shifted 7 degrees counterclockwise with respect to the wind vane’s orientation. It is not know which orientation was correct and neither was adjusted for this difference.

![Graph showing comparison of mean wind direction values](image)

**Figure A.27** Comparison of five years of run mean wind direction values for the sonic anemometer and wind vane.

To investigate the numerous outliers of large magnitude observed in figure A.27, regressions between the two signals were calculated for 0.5 m s\(^{-1}\) wind velocity groupings up to a velocity of 2 m s\(^{-1}\). The results of these regressions, table A.5, suggested that the correspondence in wind direction between the two sensors weakened at velocities below 1 m s\(^{-1}\). It is known that wind vanes may be subject to stall errors at low velocities (Albers et al. 2000; Brock et al. 2001). Comparisons of the wind vanes under different conditions of atmospheric stability, horizontal velocity attack angle, and precipitation suggested no effects of these conditions...
upon the comparison between sensors. Because of the probable errors in the wind vane at low velocities and because of the close correspondence at higher wind velocities, sonic anemometer derived wind direction values were used as the preferential wind direction measurements. Wind vane measurements were employed if no sonic anemometer measurements were available.

Table A.5 Coefficients and $R^2$ values for the Linear regression of sonic anemometer wind direction against wind vane wind direction ($\theta_s = \text{offset} + \text{Slope} \times \theta_v$).

<table>
<thead>
<tr>
<th>Wind speed</th>
<th>Offset</th>
<th>Slope</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 – 0.5</td>
<td>-11.99</td>
<td>1.042</td>
<td>0.84</td>
</tr>
<tr>
<td>0.5 – 1.0</td>
<td>-9.49</td>
<td>1.022</td>
<td>0.96</td>
</tr>
<tr>
<td>1.0 – 1.5</td>
<td>-6.78</td>
<td>1.011</td>
<td>0.98</td>
</tr>
<tr>
<td>1.5 – 2.0</td>
<td>-5.36</td>
<td>1.005</td>
<td>0.99</td>
</tr>
<tr>
<td>&gt; 2.0</td>
<td>-6.45</td>
<td>1.010</td>
<td>0.99</td>
</tr>
</tbody>
</table>

Figure A.28 Comparison of difference between sonic anemometer and wind vane wind direction standard deviation values for five years of data.

Values of wind direction standard deviations were compared by examining the difference between sonic and wind vane standard deviations, figure A.28. It is
noted that wind vane values of the standard deviation of wind direction greatly underestimate corresponding values obtained from the sonic anemometer for wind speeds less than about 1.5 m s\(^{-1}\). At wind speeds higher than 1.5 m s\(^{-1}\) the relationship appears to tend towards an asymptote that suggests that the wind vane standard deviations are greater than those of the sonic anemometer. Both the underestimate at low wind speed and overestimate at high wind speed of the wind vane may be explained by errors associated with sensor inertia (Brock et al. 2001; Wyngaard 1981). As with wind direction, values of wind direction standard deviation were preferentially obtained from sonic anemometer measurements.

### A.7 Precipitation

Precipitation was measured to determine the amount of liquid water influx to the experiment site. The measurements, however, may not have accounted for fog deposition or condensation, or snowfall. Although measurements of precipitation approximate the total amount of water available to the experiment site, the amount of water available to the vegetation is reduced by canopy interception/evaporation and runoff. Values of canopy through fall, stem flow, and catchment drainage were measured in association with another project and are reported here for completeness.

### A.7.1 Measurement methods

#### A.7.1.1 Tipping bucket rain gauges

Two tipping bucket rain gauges were employed for continuous monitoring of precipitation. The first precipitation gauge was installed at the base of the eddy covariance tower from the beginning of the experiment until February 1998. This sensor's close proximity to over story vegetation made it susceptible to canopy interception of precipitation. In February 1998, this sensor was relocated to a position slightly south of the centre of a forest ride in the proximity of the power hut (see figure 3.1). In August 1997, a second tipping bucket was installed on a pole, approximately 1 meter above the solar panels mounted on the power hut. This location was chosen to minimize the interception effects of the surrounding trees, though it may have been affected by wind flow distortion caused by the solar panels.
Both sensors were calibrated *in situ* by slowly siphoning a volume of water from a two litre graduated cylinder into the tipping bucket. The calibration drain tube was placed close to the tipping buckets collection hole to minimize the effects of gauge wetting and a cover was placed over the bucket to eliminate effects of precipitation and evaporation. The number of tips recorded on the data logger was used with the volume change of the graduated cylinder to produce a calibration coefficient. The effect of resolution error of the tipping bucket on the calibration was small for the volume of calibration water used (~2%).

Both sensors were occasionally checked for sensor contamination or blockage. The elevated sensor, which was constructed of plastic and contained an integral filtering screen, exhibited more problems with blockage due to algal growth on the filter screen. The surface sensor, which was constructed of brass, only required occasional removal of plant material, which had fallen into the sensor. No attempt was made to remove or melt snow or ice from either sensor.

**A.7.1.2 Static rain gauges**

Static rain gauge measurements were made at 5 locations for the period spanning April 1, 2000 to December 5, 2000 as part of a nitrogen deposition study. These measurements were made using manual precipitation gauges, which were recorded on a weekly to biweekly basis. One gauge was placed adjacent to the automatic gauges at the centre of the experiment site while the other four gauges were placed adjacent to the road surrounding the site at approximately NNW, WSW, SW, and NE relative to the centre of the site, confer figure 2.4.

**A.7.1.3 Throughfall and stem flow gauges**

In addition to the static rain gauge measurements, measurements of stem flow and throughfall were conducted in three plots for the period April 1, 2000 to December 5, 2000. Throughfall and stem flow measurements were made on six trees in each of the three plots.

Four throughfall measurements were made for each sample tree using 30 cm diameter straight walled plastic containers with a fabric mesh covering to exclude
plant matter. The containers were placed at a radial distance of 1 meter from the tree. Some error will have arisen by evaporation from the fabric mesh and from the exposed water surface of the buckets. No estimate of these errors was attempted.

Stem flow was estimated by spirally encircling each of the sample tree boles with split tubing, which acted as a conduit and guided water flowing down the bole into an adjacent holding container. Gaps between the tree and the guide tubing were stopped with silicone rubber.

Both the through fall and stem flow volume measurements were manually recorded on a weekly to biweekly basis. Throughfall volume was converted to depth using the dimensions of the collection containers. Stem flow measurements were converted to equivalent depths by scaling the average per tree volume to an area average volume using the mean planting density.

### A.7.1.4 Stream flow gauges

Stream flow measurements were conducted by Xiangqing Ma and Kate Heal and the results presented here reflect their efforts. Stream flow was measured in a stable, laminar flow cross section of Cultillich Burn from April-December 2000. Stream flow measurements at this location captured drainage from the entire upstream catchment area, which encompasses the experiment site. Stream flow was estimated from river stage using a ratings curve derived from velocity-area measurements made on 12 occasions between April-December 2000 in a range of flows, using an OTT C2 current meter. River stage was measured every 10 minutes using a pressure transducer corrected for atmospheric pressure. The ratings curve had a strongly significant fit between logarithm of the pressure transducer reading in mV and the logarithm of the stream flow ($R^2 = 0.975$, $p<0.001$). Stream flow volumes were scaled to an average catchment depth, in mm, using a catchment size of 4.7 km$^2$. 

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A.7.2 Sensor comparison

A comparison of the two automated rain gauges employed in this experiment, figure A.29, shows a strong relationship ($R^2 = 0.81$), which indicates that the surface mounted gauge received approximately 25% less precipitation than the elevated gauge.

![Figure A.29](image)

**Figure A.29** Comparison of diel total precipitation for the elevated and surface automated rain gauges. Because of the low frequency of large rain events data are plotted on a log-log plot.

The effect will have been greater during precipitation periods when the wind was from the southeast if the low values of the surface mounted sensor were a result of interception by the forest. (The surface mounted sensor was placed, it was located approximately 10 m from the forest edge to the southeast and 15 m from the forest edge to the northwest). The run average precipitation by wind direction, figure A.30, indicates that this effect was occurring. We also observe that precipitation was heaviest under southeasterly flow conditions. The placement of the elevated precipitation gauge closer to the forest edge to the north suggests that any interception errors in the elevated sensor would have been small.
Figure A.30 Run averaged precipitation for the elevated and surface automated rain gauges for 10 deg wind direction bins.

Figure A.31 Comparison of total precipitation for the automated and manual precipitation gauges. Error bars indicate one standard deviation.

As a cross check, automated gauge values were summed to correspond with the manual rain gauges, which had been operated from April to December, 2000. A comparison of the average totals from the two collection methods, figure A.31, indicates excellent agreement between the two methods. Cumulative totals over the period of manual sampling, figure A.32, indicate consistent behaviour of the different gauges throughout the period, with the two automated gauges reading higher and lower than the manual gauges. Considering the observed losses by the surface mounted precipitation gauge and the potential for losses in the manual
gauge it is suggested that the elevated gauge provided the most accurate precipitation measurements.

![Diagram showing cumulative precipitation comparison](image)

**Figure A.32** Comparison of cumulative precipitation for individual automated and manual precipitation gauges.

### A.8 Atmospheric pressure

#### A.8.1 Measurement methods

Atmospheric pressure was measured with an aneroid barometer (model PTB101, Vaisala, Helsinki, Finland). The sensor was situated within an enclosure at the base of the profile tower, which also housed the profile tower data loggers and IRGA manifold. The sensor was housed in a flexible container along with desiccant to prevent condensation. The sensing port was exposed to the interior of the enclosure via a 6.4 mm diameter vent tube. There was no evidence that dynamic pressure fluctuations would have affected the sensor when deployed in this manner. The manufacturers calibration was used in converting the sensors signal from millivolt to kPa. Half hour averages of the signal were recorded using a data logger (CR21X, Campbell Scientific, Logan Utah, USA). The recorded mean atmospheric pressure values were retrieved weekly and maintained in database of results. The half hour values air temperature were quality checked for reasonable values (950 to 1020) and by inspection.
A.9 Soil moisture

A.9.1 Measurement methods

A.9.1.1 Gravimetric

Gravimetric methods provided the simplest approach to determining soil moisture. However, few such samples were collected because of the labour intensive nature of this method. When samples were collected, they were usually sampled in association with other soil sampling procedures. Samples were collected in association with the soil bulk density transects (see section 2.5.2). The sample handling is described in their respective sections. Volumetric soil moisture, $\theta_v$, was determined gravimetrically from the following equation (Hewlett 1982):

$$\theta_v = B_s \frac{m_w}{m_s}$$

where $B_s$ is the soil’s bulk density, $m_w$ is the mass of water in a sample of soil, and $m_s$ is the sample’s mass of dry soil.

A.9.1.2 Time domain reflectometry

TDR measurements of soil moisture make use of the dielectric properties of water to infer the amount of water in close proximity to a TDR probe. An electrical pulse is transmitted down a wire and along a probe inserted into the soil. As the pulse travels along the soil probe, water adjacent to the probe reflects a portion of the pulse. The amount of signal reflected as the pulse travels through the soil is proportional to the soil’s volumetric water content (Massman & Lee 2002).

An empirical conversion relates the reflected signal to the soil’s volumetric water content. This conversion is a function of soil type, salinity, and temperature.

There were two periods of TDR probes deployment. During the first period Oct 1997 to Nov 1998 a cable tester (model 1502C, Tectronix, Beaverton Oregon, US)
and TDR multiplexer (model SDMX50 Campbell Sci., Logan Utah, USA) were used to sample between 4 and 8 pairs of 30 cm stainless steel soil probes via equal lengths of cable (Belden 9090, St Louis, Mo, USA). Each pair of probes was inserted into the soil at an angle of either 90 or 60 degrees, corresponding to sampling of either the top 30 cm or top 15 cm soil layer. The probes were placed in the open canopy plot adjacent to the profile tower (see figure 3.3). Two sets of probe pairs were located in the ridge, furrow, and unploughed regions of the plot. At each location, probes were placed to sample both the 15 and 30 cm depth layers. Each probe pair was inserted to its full length using an insertion die to maintain a 5 cm spacing between soil probes. The cable tester employed was not designed for continuous field use so that it was necessary to remove the system after eight months due to moisture damage.

During the second period, dedicated TDR probes (model CS615, Campbell Scientific, Logan, Utah, USA) associated with a data logger were deployed adjacent to the eddy covariance tower. These probes are sealed and weather resistant. Each probe consists of a parallel pair of 30 cm stainless steel probes attached to a sealed electronics package. Only two of these probe pairs were installed at this location. One probe was installed in a ridge while the second was installed in an unploughed region. Both sensors were installed at an angle of 60 degrees in order that the top 15 cm of soil was sampled.

A.9.2 Sensor comparison

No continuous comparison of soil moisture measurement methods is possible because of the temporal separation of the sampling strategies. However, table A.6 gives estimates of variability associated with the soil moisture content estimates based on the gravimetric and two TDR sampling approaches. The smaller variability observed in the 1998 TDR measurements may reflect the restricted sampling taken with these devices or it may reflect on a greater potential for error associated with the manual measurements gravimetrically determined soil moisture. Further testing would be required to resolve this question.
Table A.6 Comparison of spatial variability observed in gravimetric determined and TDR determined volumetric determinations of soil moisture content. The error (± one standard deviation) associated with the TDR measurements were obtained as statistics derived from repeated measures of the eight sampling locations.

<table>
<thead>
<tr>
<th>Volumetric water content</th>
<th>Average g cm⁻³</th>
<th>Std dev g cm⁻³</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transect sampling</td>
<td>0.50</td>
<td>0.22</td>
<td>92</td>
</tr>
<tr>
<td>TDR in 1998</td>
<td>0.37±0.04</td>
<td>0.057±0.007</td>
<td>8  (2677)</td>
</tr>
</tbody>
</table>

A.10 Soil heat flux

The flux of heat into the soil may be estimated by measuring soil temperature profiles and inferring the heat transport down that profile based on soil properties (Monteith 1973; Rosenberg, Blad, & Verma 2003). A more common method is to measure the heat flux using soil heat flux plates. These devices are sensitive thermopiles that are capable of measuring their temperature difference between their top and bottom sides. Because the plates have a known conductively, the heat transfer through the plate will be proportional to the temperature difference across the plate. Soil heat flux plates measurements assume that the thermal conductivity of the soil is identical to that of the plate. Errors in this method may exist if the soil is of different conductivity than the plate. Errors may also be caused by vapour transport from below the level of the plate, advective transfer of heat by air or water, and due to selective heat transfer paths, (Kimball & Jackson 1999).

A.10.1 Measurement methods

A.10.1.1 Soil heat flux plates

Soil heat flux plates were employed as the primary estimate of the flux of heat into the soil. In order to capture some of the potential variability, seven plates were installed in two plots located near the profile tower. The plates were installed in the plots in an effort to characterise the plot soil surface variability. Three of these plates where installed in a plot that was near canopy closure. In this plot, one plate was installed in a furrow and two plates were installed in an unploughed area. Four plates were installed in a second plot in which the canopy was still open. The
canopy in this plot did not reach closure until near the end of the experiment. In the open canopy plot, one plate was installed in a furrow and one in an unploughed area; the two remaining plates were installed in an unploughed area shaded by a small tree.

Plates were installed horizontally at a depth of 5 cm. The depth of installation was measured from relative to the top of the litter layer, which had been compressed by hand. During installation, a block of soil slightly larger than the plate was removed to a depth of slightly more than 5 cm and a horizontal cut was then made in one side at the bottom of the resulting hole. The plate was inserted into the horizontal cut and the soil block replaced.

It was assumed that there was no horizontal heat advection and no vertical advection of evaporated soil water from below the level of the installed sensor. The assumption of evaporation will have produced an underestimate that would have been most significant during summer when the water table was well beneath the surface. Horizontal heat advection because of topography or variability in soil texture or water content will have had unpredictable effects on the soil heat flux (Aase, Jackson, & Idso 1975).

A.10.1.2 Soil temperature profiles

When available and in the absence of soil heat flux plates, soil temperature profiles were used to estimate values of soil heat flux. Using the relationship

\[ G = K \cdot \frac{\partial T_s}{\partial z} \]  

(A.13)  

in which \( K \) is the thermal conductivity of the soil.

It is known that values of \( K \) will vary with soil type and soil moisture conditions. Because the soil temperature profiles and soil heat flux plates were installed in close proximity, it was assumed that soil composition and soil moisture was similar for both the measurements of \( G \) and \( T_s \). With this assumption, it was possible to determine values of \( K \) for different soil moisture conditions by rearranging equation
A.17. Average values of $G$ and $\partial T_z/\partial z$ were used to calculate $K$, which was then grouped by 5% bins of volumetric soil moisture content. The median value for each bin was chosen as the representative value of $K$, figure A.33.

The resulting soil thermal conductivities are close to the conductivities of water (0.586 W m$^{-2}$ K$^{-1}$) at high soil water contents. At lower soil water content, the conductivities fall to values between 0.28 and 0.35 W m$^{-2}$ K$^{-1}$. These values are slightly higher than the values specified for peat soil by (Monteith 1973; Van Wijk & Derksen 1966) which may be consistent with a greater mineral content at this depth, see figure 2.22.

![Graph showing soil thermal conductivity calculated from site average soil heat flux and temperature difference between the 5 and 6 cm depths.](image)

Figure A.33 Soil thermal conductivity calculated from site average soil heat flux and temperature difference between the 5 and 6 cm depths.

A.10.2 Sensor comparisons

Comparison of open and closed canopy plates indicate closed canopy soil heat flux was significantly more positive ($n=24452$, $p<0.001$), though the difference in medians was small, table A.7. Contrary to expectations, the quartile range observed in the closed canopy was greater than that observed in the open canopy. The cause of the greater range of soil heat flux under a closed canopy is likely an effect of under sampling, see table 3.1, as the number of sample taken would have been insufficient for determining the small expected difference between open and closed canopy locations. Paired comparisons of individual plates revealed significant
differences between all but two comparisons. However, insufficient information on the physical characteristics of locations was available to determine the cause of these differences.

<table>
<thead>
<tr>
<th>Canopy</th>
<th>Median</th>
<th>25% quartile</th>
<th>75% quartile</th>
</tr>
</thead>
<tbody>
<tr>
<td>Open canopy all</td>
<td>0.165</td>
<td>-1.847</td>
<td>2.170</td>
</tr>
<tr>
<td>Closed canopy all</td>
<td>0.218</td>
<td>-2.350</td>
<td>2.800</td>
</tr>
</tbody>
</table>

Table A.7 Median and quartile values of average open and closed canopy soil heat flux values.
B Derivations, equations, and models

This appendix contains derivations, models and equations that were considered too lengthy, trivial or not crucial to the body of the thesis. Derivations of length were supplied in this appendix to maintain a more consistent flow to the body of the thesis itself. Final derived equations are repeated in the body of the thesis. The models presented may or may not be explicitly mentioned in the thesis. The models presented were considered secondary to the main aim of the thesis or were used for minor gap filling or gap filling of secondary variables. The equations presented are those used for calculations and conversions during analysis of the data set. A brief description of equations is given, and references to their source, if appropriate. Derivations are presented first, followed by a mixture of models and equations.

B.1 Derivation of corrections to sonic temperature

A more general presentation of this topic is covered in appendix H. The derivation presented in this section is an expansion on that presented in appendix H and is given for the purpose of clarity.

The standard corrections to sonic temperature are outlined in (Kaimal & Gaynor 1991; Schotanus, Nieuwstadt, & De Bruin 1983). This derivation follows the methods outlined in those papers. Differences in the final equation result from the inclusion of actual path length and changes in velocity between subsequent sound pulse samples.

The initial equations (Kaimal et al. 1991) define the sonic temperature, $T_s$, for dry still air based on an assumed speed of sound in air, $c_s$.

$$ T_s = \frac{c_s^2}{403} \quad \text{(B.1)} $$

An assumed speed of sound, $c_{sa}$, may also be determined using a sonic anemometer, from the transit time of two sound pulses, $t_1$, $t_2$, over an assumed known distance, $l_a$. 

\begin{align*}
\text{(B.1)}
\end{align*}
If liquid water is present on the faces of the transducers, the transit time of the sound pulses in air, $t_{a1}$, $t_{a2}$, will be related to the measured transit times, $t_1$, $t_2$, by accounting for the transit time through liquid water, $t_w$.

$$t_{a1} = t_1 - t_w$$  \hspace{1cm} (B.3)

$$t_{a2} = t_2 - t_w$$

From figure 1 in appendix H we can determine the relationship of $t_{a1}$ and $t_{a2}$ to the sound path geometry following Kaimal and Gaynor (Kaimal et al. 1991).

$$t_{a1} = \frac{l_m - l_w}{c_s \cdot \cos(\gamma_{a1}) + V_d}$$  \hspace{1cm} (B.4)

$$t_{a2} = \frac{l_m - l_w}{c_s \cdot \cos(\gamma_{a2}) - (V_d + \Delta V_d)}$$

The difference in this formulation lies in the inclusion of terms for the incorporation of actual path length, $l_m$, depth of liquid water on the transducer faces, $l_w$, the increase $\Delta V_d$, in the along path wind velocity, $V_d$, and the in sound path deviations, $\gamma_{a1}$, $\gamma_{a2}$, caused by changes in the cross path wind velocity, $\Delta V_n$:

$$\gamma_{a1} = \sin^{-1} \left( \frac{V_n}{c_s} \right)$$  \hspace{1cm} (B.5)

$$\gamma_{a2} = \sin^{-1} \left( \frac{V_n + \Delta V_n}{c_s} \right)$$

Incorporating equation B.5 into equation B.4 Results in:

$$t_{a1} = \frac{l_m - l_w}{\sqrt{c_s^2 - V_n^2} + V_d}$$  \hspace{1cm} (B.6)

$$t_{a2} = \frac{l_m - l_w}{\sqrt{c_s^2 - (V_n + \Delta V_n)^2} - (V_d + \Delta V_d)}$$

which may be rearranged to move the transit times to the denominator:
\[ \sqrt{c_s^2 - V_n^2} + V_d = \frac{l_m - l_w}{t_{s1}} \]  
(B.7)

\[ \sqrt{c_s^2 - (V_n + \Delta V)^2} - (V_d + \Delta V^2) = \frac{l_m - l_w}{t_{s2}} \]

By combining equations B.7:

\[ \sqrt{c_s^2 - V_n^2} + V_d + \sqrt{c_s^2 - (V_n + \Delta V_n)^2} - (V_d + \Delta V_d) = \frac{l_m - l_w}{t_{s1}} + \frac{l_m - l_w}{t_{s2}} \]  
(B.8)

and rearranging:

\[ \sqrt{c_s^2 - V_n^2} + \sqrt{c_s^2 - (V_n + \Delta V)^2} - \Delta V = (l_m - l_w) \cdot \left( \frac{1}{t_{s1}} + \frac{1}{t_{s2}} \right) \]  
(B.9)

we obtain a formulation that may be substituted into equation B.2

\[ \sqrt{c_s^2 - V_n^2} + \sqrt{c_s^2 - V_n^2 - 2 \cdot V_n \cdot \Delta V_n - \Delta V^2} - \Delta V_d = (l_m - l_w) \cdot \left( \frac{1}{t_{s1}} + \frac{1}{t_{s2}} \right) \]  
(B.10)

Before doing so we perform a binomial expansion of the second term on the left hand side of equation B.10

\[ (a + b)^n = a^n + n a^{n-1} b + \frac{n(n-1)}{2!} a^{n-2} b^2 + \frac{n(n-1)(n-2)}{3!} a^{n-3} b^3 + \ldots \]

\[ a = \left( c_s^2 - V_n^2 \right) \]

\[ b = \left( -2 \cdot V_n \cdot \Delta V_n - \Delta V^2 \right) \]

\[ n = \frac{1}{2} \]

\[ \left( c_s^2 - V_n^2 \right)^{1/2} - \frac{2 \cdot V_n \cdot \Delta V_n + \Delta V^2}{2 \left( c_s^2 - V_n^2 \right)^{1/2}} - \left( \frac{2 \cdot V_n \cdot \Delta V_n + \Delta V^2}{8 \left( c_s^2 - V_n^2 \right)^{3/2}} \right)^2 + \ldots \]

Expanding the terms of the binomial expansion produces the equation:
\[
\left( c_s^2 - V_n^2 \right)^{1/2} - \frac{V_n \cdot \Delta V_n}{\left( c_s^2 - V_n^2 \right)^{1/2}} - \frac{\Delta V_n^2}{2 \left( c_s^2 - V_n^2 \right)^{1/2}} - \frac{\left( V_n \cdot \Delta V_n \right)^2 + V_n \cdot \Delta V_n^3 + 4 \Delta V_n^4}{2 \left( c_s^2 - V_n^2 \right)^{3/2}} \quad (B.12)
\]

Eliminating higher order terms incorporating the value \( \Delta V_n \) reduces this equation to:

\[
\left( c_s^2 - V_n^2 \right)^{1/2} - \frac{V_n \cdot \Delta V_n}{\left( c_s^2 - V_n^2 \right)^{1/2}} \quad (B.13)
\]

This equation is then incorporated into equation B.10:

\[
2\sqrt{c_s^2 - V_n^2} - \frac{V_n \cdot \Delta V_n}{\sqrt{c_s^2 - V_n^2}} - \Delta V_d = \left( l_m - l_w \right) \cdot \left( \frac{1}{l_{a1}} + \frac{1}{l_{a2}} \right) \quad (B.14)
\]

With some slight rearrangement this equation may be substituted into equation B.2, replacing the bracketed term on the right hand side to produce an equation for the speed of sound:

\[
c_{sa} = \frac{l_s}{2 \cdot \left( l_m - l_w \right)} \left[ 2\sqrt{c_s^2 - V_n^2} - \frac{V_n \cdot \Delta V_n}{\sqrt{c_s^2 - V_n^2}} - \Delta V_d \right] \quad (B.15)
\]

Multiplying through by \( \frac{1}{2} \) gives:

\[
c_{sa} = \frac{l_s}{\left( l_m - l_w \right)} \left[ \sqrt{c_s^2 - V_n^2} - \frac{V_n \cdot \Delta V_n}{2\sqrt{c_s^2 - V_n^2}} - \frac{\Delta V_d}{2} \right] \quad (B.16)
\]

We further reduce the equation by removing the third term from the right hand side, which is justified considering its small magnitude relative to the last term:

\[
c_{sa} = \frac{l_s}{\left( l_m - l_w \right)} \left[ \sqrt{c_s^2 - V_n^2} - \frac{\Delta V_d}{2} \right] \quad (B.17)
\]
This equation may then be substituted into equation B.1 to obtain a definition of the sonic temperature, which is uncorrected for the previously described effects, $T_{su}$:

$$T_{su} = \left( \frac{l_s}{(l_m - l_w)^2} \left( \frac{\sqrt{c_s^2 - V_n^2 - \Delta V_d^2}}{403} \right) \right)^2$$

(B.18)

This resulting equation is expanded and simplified in the following steps:

Square terms

$$T_{su} = \frac{l_s^2}{(l_m - l_w)^2} \left( \frac{c_s^2 - V_n^2 - \Delta V_d \cdot \sqrt{c_s^2 - V_n^2 + \frac{\Delta V_d^2}{4}}}{403} \right)$$

(B.19)

Apply denominator to bracketed terms.

$$T_{su} = \frac{l_s^2}{(l_m - l_w)^2} \left( \frac{c_s^2 - V_n^2 - \Delta V_d}{403} \cdot \sqrt{\frac{c_s^2}{403^2} - \frac{V_n^2}{403^2} + \frac{\Delta V_d^2}{4 \cdot 403}} \right)$$

(B.20)

Convert speed of sound terms to sonic temperature and rearrange.

$$T_{su} = \frac{l_s^2}{(l_m - l_w)^2} \left( \frac{c_s^2 - V_n^2}{403} \cdot \sqrt{\frac{\Delta V_d}{403} \cdot \frac{T_s - V_n^2}{20.1} - \frac{\Delta V_d}{4}} \right)$$

(B.21)

Again perform a binomial expansion, of the square root term and ignore the higher order expansion results:

$$T_{su} = \frac{l_s^2}{(l_m - l_w)^2} \left( \frac{c_s^2 - V_n^2}{403} \cdot \frac{\Delta V_d}{403} \left[ 20.1 \cdot \frac{T_s - V_n^2}{20.1} - \frac{\Delta V_d}{4} \right] \right)$$

(B.22)

Expanding this equation gives:

$$T_{su} = \frac{l_s^2}{(l_m - l_w)^2} \left( \frac{c_s^2 - V_n^2}{403} \cdot \frac{\Delta V_d \cdot \sqrt{T_s}}{20.1} + \frac{\Delta V_d \cdot V_n^2}{8100 \cdot \sqrt{T_s}} + \frac{\Delta V_d^2}{1612} \right)$$

(B.23)

from which the last two terms are dropped to produce:
\[ T_{su} = \frac{l_s^2}{(l_m - l_w)^2} \left[ T_s - \frac{V_n^2}{403} - \frac{\Delta V_d \cdot T_{su}^{1/2}}{20.07} \right] \] 

(B.24)

By assuming \( T_{su} = T_s \) for the last term on the right hand side this equation may be rearranged to obtain a definition of the sonic temperature incorporating the actual path length, the depth of liquid water, the cross sound path wind velocity and the along path change in wind velocity between subsequent sound pulses.

\[ T_s = \left( \frac{l_m - l_w}{l_s} \right)^2 T_{su} + \frac{V_n^2}{403} + \frac{\Delta V_d \cdot T_{su}^{1/2}}{20.07} \] 

(B.25)

This may be further modified, following (Kaimal et al. 1991; Schotanus et al. 1983) to obtain the true air temperature, \( T_a \), by incorporating the density effects of water vapour on the corrected sonic temperature:

\[ T_a = \left( \frac{l_m - l_w}{l_s} \right)^2 T_{su} + \frac{V_n^2}{403} + \frac{\Delta V_d \cdot T_{su}^{1/2}}{20.07} - 0.00032 \cdot T_s \cdot q \] 

(B.26)

### B.2 Derivation/revision of mean vertical velocity effects of density variations

Consider a volume with a unit surface area below the height of flux measurement. We initially define this volume to be impermeable to exchange but having a source of heat, pressure and gases at its bottom surface. We also assume that the volume can expand only in the vertical direction. We would like to know the rate at which this volume’s height changes.

We specify the heat added to this volume to be equivalent to the rate of sensible heating of the atmosphere corresponding to the surface sensible heat flux. We also specify the source of gas molecules to correspond with the surface latent heat flux.
The amount of heat added to the column can be determined from the sensible heat flux using the equation:

\[ \Delta T = \frac{\Delta t \cdot H \cdot A}{\rho \cdot C_\rho \cdot V_L} = \frac{\Delta t \cdot H}{\rho \cdot C_\rho \cdot h_1} \]  
(B.27)

Similarly, the number of moles water vapour molecules may be determined from the equation:

\[ \Delta n = \frac{\Delta t \cdot \lambda E \cdot A}{\lambda \cdot M_v} = \frac{\Delta t \cdot \lambda E}{\lambda \cdot M_v} \]  
(B.28)

Going back to the initial volume below the height of flux measurement, it may be defined in terms of the ideal gas law as:

\[ V_{L0} = \frac{R_g \cdot T_0}{P_0} \cdot \sum n_i \]  
(B.29)

Where \( R_g \) is the gas constant, \( T \) is temperature in degrees Kelvin, \( P \) is atmospheric pressure in Pascals, \( n \) is the number of moles of gas species, where the particular species is designated by the index \( i \).

Over some measurement period this initial volume will be altered as a result of changes in temperature, pressure, or moles of component species.

\[ V_{L1} = \frac{R_g \cdot (T_0 + \Delta T)}{(P_0 + \Delta P)} \cdot \sum (n_i + \Delta n_i) \]  
(B.30)

The change in volume can be determined by calculating the difference between the final and initial volumes.

\[ \Delta V_L = \frac{R_g \cdot (T_0 + \Delta T)}{(P_0 + \Delta P)} \cdot \sum (n_i + \Delta n_i) - V_{L0} \]  
(B.31)

Do a binomial series expansion of the pressure term in the denominator, keeping only the first two terms.

\[ \Delta V_L = \frac{R_g}{P_0} \cdot (T_0 + \Delta T) \cdot \left( 1 - \frac{\Delta P}{P_0} \right) \cdot \sum (n_i + \Delta n_i) - V_{L0} \]  
(B.32)
We next employ equation B.29 to replace the ratio R/P,

$$\Delta V_L = \frac{V_{10}}{T_0} \sum_i \frac{n_i}{n_{i0}} \cdot (T_0 + \Delta T) \cdot \left(1 - \frac{\Delta P}{P_0}\right) \cdot \sum_i (n_{i0} + \Delta n_i) - V_{L0} \quad (B.33)$$

Dividing both sides by the initial volume give the volume change as a fraction of the initial volume.

$$\frac{\Delta V_L}{V_{L0}} = \frac{1}{\sum_i n_{i0}} \cdot \left(\frac{T_0 + \Delta T}{T_0}\right) \cdot \left(1 - \frac{\Delta P}{P_0}\right) \cdot \sum_i (n_{i0} + \Delta n_i) - 1 \quad (B.34)$$

By rearranging the temperature and mass terms we obtain an equation describing the fraction change in volume in terms of fractional changes in volume temperature, pressure and number of moles of gas.

$$\frac{\Delta V_L}{V_{L0}} = \left(1 + \frac{\Delta T}{T_0}\right) \cdot \left(1 - \frac{\Delta P}{P_0}\right) \cdot \left[1 + \frac{\sum_i \Delta n_i}{\sum_i n_{i0}}\right] - 1 \quad (B.35)$$

Because we are assuming a volume above a unit area we may convert the volume ratios to height ratios. This also requires us to assume horizontal homogeneity to prevent lateral expansion of the volume.

$$\frac{\Delta h}{h_{i0}} = \left(1 + \frac{\Delta T}{T_0}\right) \cdot \left(1 - \frac{\Delta P}{P_0}\right) \cdot \left[1 + \frac{\sum_i \Delta n_i}{n_i}\right] - 1 \quad (B.36)$$

Dividing through by the length of the measurement period, in seconds, and moving the height denominator to the RHS, we obtain the rate of change of volume height. For the conditions we have assumed this will represent the vertical velocity of the top of the volume.

$$\frac{\Delta h}{\Delta t} = \frac{h_{i0}}{\Delta t} \left[1 + \frac{\Delta T}{T_0}\right] \cdot \left(1 - \frac{\Delta P}{P_0}\right) \cdot \left[1 + \frac{\sum_i \Delta n_i}{n_{i0}}\right] - 1 \quad (B.37)$$

While equation B.37 should hold true for an impermeable volume, we know that in the atmosphere that this is not the case because energy and mass are transported across the top of the volume. To deal with this effect we determine the vertical
velocity components associated with storage, \( \bar{w}_s \) and with transport across the upper plane of the volume \( \bar{w}_f \).

If we alter our assumptions slightly and assume that instead of an impermeable volume we have a volume containing a constant number of moles of gas while allowing transport of heat, pressure, and trace gases across the upper boundary we obtain a slightly altered version of equation B.37 which is representative of the vertical velocity associated with temperature and pressure changes within the volume, \( w_S \).

\[
\bar{w}_S = \frac{\Delta h_0}{\Delta t} \left[ \left( 1 + \frac{\Delta T_S}{T_0} \right) \left( 1 - \frac{\Delta P_S}{P_0} \right) - 1 \right]
\]

(B.38)

In this equation the mole change term has disappeared by our assumption, while the temperature and pressure change terms represent the observed volume averaged temperature and pressure changes over the period of observation.

As shown in equation B.38, because we are assuming horizontal homogeneity we may associate volume changes with a value of vertical velocity. Therefore, to obtain the vertical velocity associated with the fluxes of mass and energy we would like to know the transport of volume associated with the transfer of heat, trace gases and pressure across the top of the specified volume.

To determine the amount of volume lost to transport out of the volume we revert again to the ideal gas law. Because we are interested in transport we have restated equation B.29 in terms of mean and fluctuating components.

\[
\bar{V}_L + V' = \frac{R}{P} \left( \bar{T} + T' \right) \left( 1 - \frac{P'}{P} \right) \left( \bar{n} + n' \right)
\]

(B.39)

We again rearrange the temperature and mole terms in order to convert the ratio of \( R/P \) to volume.

\[
\bar{V}_L + V' = \bar{V}_L \left( 1 + \frac{T'}{T} \right) \left( 1 - \frac{P'}{P} \right) \left( 1 + \frac{n'}{n} \right)
\]

(B.40)
We next expand the bracketed terms on the RHS of equation B.40.

\[ \overline{V_L} + \overline{V'_L} = \overline{V_L} \cdot \left( 1 + \frac{n'}{n} + \frac{T'}{T} - \frac{P'}{P} + \frac{n'T'}{nT} - \frac{n'P'}{nP} - \frac{P'T'}{PT} - \frac{n'P'T''}{nP'T} \right) \]  

(B.41)

Multiplying both sides of the equation by the vertical velocity fluctuation and dividing both sides by the mean volume gives:

\[ \frac{w'V'}{V_L} = \frac{w'n'}{n} + \frac{w'T'}{T} - \frac{w'P'}{P} + \frac{w'n'T'}{nP} - \frac{w'n'P'}{nP'} - \frac{w'P'T'}{PT} - \frac{w'n'P'T''}{nP'T} \]  

(B.42)

From which we exclude the triple and quadruple prime terms as being insignificant to give the relation of flux of volume per total volume.

\[ \frac{w'V'}{V_L} = \frac{w'n'}{n} + \frac{w'T'}{T} - \frac{w'P'}{P} \]  

(B.43)

Again we extract the constant unit surface area to obtain the flux of height per volume height.

\[ \frac{w'h'}{h} = \frac{w'n'}{n} + \frac{w'T'}{T} - \frac{w'P'}{P} \]  

(B.44)

which has units corresponding to the vertical velocity associated with the transport of energy and mass across the top of the volume, \( w_f \).

\[ \frac{w'}{w_f} = \frac{w'n'}{n} + \frac{w'T'}{T} - \frac{w'P'}{P} \]  

(B.45)

In this equation the numerators on the RHS respectively represent the fluxes of molecules, temperature, and pressure across the top of the volume while the denominators represent the corresponding run mean values.

We then combine the vertical velocities associated with storage within the volume and flux across the top of the volume into a single vertical velocity term.

\[ \overline{w} = \overline{w_s} + \overline{w_f} \]  

(B.46)

Which combined gives the vertical velocity caused by storage and flux across the top of the volume in terms of changes in the column variables and fluxes.
\[
\frac{-w}{T} = \frac{h_{0}}{\Delta T} \left[ \left( \frac{1 + \Delta T}{T} \right) \left( 1 - \frac{\Delta P}{P_0} \right) - 1 \right] + \frac{w' n'}{n} + \frac{w' T'}{T} - \frac{w' P'}{P} \quad (B.47)
\]

### B.3 Radiation equations

The following equations are used to model the radiation conditions for the Griffin experiment site, and were used in calculation of percent potential radiation values. The elevation and azimuthal position of the sun with respect to the earth's surface are required for solar radiation modelling. The following equations, used to determine the solar azimuth and elevation were obtained from Iqbal (1983).

#### Day angle

Day angle, \( \alpha_d \), is a function of day of year, \( d_y \), and describes the position of the earth in its orbit about the sun, and has units of radians.

\[
\alpha_d = \left( \frac{2 \cdot \pi}{365} \right) \left( d_y - 1 \right) 
\quad (B.48)
\]

#### Solar declination

Solar declination angle, \( \delta \), is a function of day angle and defines the angular difference between the equatorial plane of the earth and the line between the centre of the sun and the centre of the earth. Solar declination as described by this equation has units of radians.

\[
\delta = 0.006918 - 0.399912 \cdot \cos(\alpha_d) + 0.070257 \cdot \sin(\alpha_d) - 0.006758 \cdot \cos(2 \cdot \alpha_d) + 0.000907 \cdot \sin(2 \cdot \alpha_d) - 0.002697 \cdot \cos(3 \cdot \alpha_d) + 0.001480 \cdot \sin(3 \cdot \alpha_d) 
\quad (B.49)
\]

#### Equation of time

The equation of time, \( E_t \), is a function of solar declination angle and accounts for the variation of solar day length compared to day length defined by the earth’s rotation.
This difference arises from the elliptical orbit of the earth around the sun and the fact that the earth’s equator is not in the same plane as its solar orbit.

\[ E_i = 229.18 \cdot \left[ 0.000075 + 0.001868 \cdot \cos(\delta) - 0.032077 \cdot \sin(\delta) \right] \]

Local apparent time
The local apparent time, \( t_L \), adjusts the local time \( t \) for the difference between the site longitude, \( \Omega \), and the longitude of the time standard, \( \Omega_0 \). The adjustment also includes an adjustment for the equation of time.

\[ t_L = t + \frac{4.0 \cdot (\Omega_0 - \Omega) + E_i}{60} \]

Hour angle
Hour angle, \( \alpha_h \), is a function of local apparent time and describes the position of a point on the earth’s surface with respect to its position at local solar noon and has units of radians.

\[ \alpha_h = -0.2618 \cdot (t - 12.0) \]

Solar elevation angle
The solar elevation angle, \( \beta \), describes the angle of the sun above the tangential plane at a given latitude, solar declination and hour angle.

\[ \beta = \arcsin \left( \sin(\delta) \cdot \sin(\Xi) + \cos(\delta) \cdot \cos(\Xi) \cdot \cos(\alpha_h) \right) \]

Solar azimuth angle
The solar azimuth angle, \( \alpha \), describes the angle of the sun relative to the plane defined by a position on the earth’s surface and the rotational axis of the earth and is a function of latitude, solar elevation angle and solar declination angle.

\[ \alpha_s = \arccos \left( \frac{\sin(\beta) \cdot \sin(\Xi) - \sin(\delta)}{\cos(\beta) \cdot \cos(\Xi)} \right) \]
B.4 Potential short wave radiation model

The amount of solar radiation received at the top of the earth's atmosphere per unit area perpendicular to the rays of the sun is referred to as the solar constant, $I_0$, and has a magnitude of $1368 \pm 7$ W m$^{-2}$. A similar value for PPFD of 2550 \mu mol m$^{-2}$ s$^{-1}$ assumes an extraterrestrial ratio of $R_g$ PPFD$^{-1} = 0.5333$.

If the earth had no atmosphere we could define a potential short wave radiation, $R_p$, received at the earth's surface purely as a function of solar radiation geometry using the solar elevation angle, $\beta$.

$$R_p = \sin(\beta) \cdot I_o \quad (B.55)$$

However, more realistic estimates of $R_p$ have been obtained assuming cloudless conditions and a typical atmosphere. Although complicated models of such values are available (Paltridge & Platt 1976), a simplified model, such as that given by Linacre (1992) is adequate.

$$R_p = I_o \cdot \tau_s \left( \tau_g \tau_o + \tau_w - 1 \right) \cdot \sin(\beta) \quad (B.56)$$

In this formulation, the solar constant is affected by both the solar elevation angle, and a set of atmospheric transmissivity values, which account for adsorption of radiation by a standard dry atmosphere, $\tau_g$, as a function of the relative atmospheric air mass, $m_z$.

$$\tau_g = 1 - 0.0043 \cdot m_z \quad (B.57)$$

water vapour, $\tau_w$

$$\tau_w = 1 - \frac{1}{2.03 + \left( \frac{8.5}{w^{0.365}} \right)} \quad (B.58)$$

for which the atmospheric water content is defined in terms of atmospheric pressure, $P$, in kPa and surface dew point temperature, $T_d$ in deg. C,

$$w = m_z \cdot e^{(2.257+0.0545-T_d)} \left( \frac{101.3}{P} \right)^{0.75} \quad (B.59)$$
the transmissivity of aerosols \( \tau_a \) is also estimated as

\[
\tau_a = 0.95
\]  

and the transmissivity associated with ozone, \( \tau_o \), is defined as

\[
\tau_o = 1 - \left( \frac{0.00212 \cdot X}{1 + 0.0042 \cdot X} \right) - 0.013 \cdot X^{0.0195}
\]  

for which

\[
X = 3.5 \cdot m_z
\]  

An alternate method employing Linke turbidity factors more specific to the UK was obtained from Page and Lebens (2002). This model provided an estimate of global radiation, as a combination of direct and diffuse radiation components.

\[
R_g = R_b + R_d
\]  

The estimate of direct beam radiation, \( R_b \), was determined as a function of the solar elevation angle adjusted Linke turbidity factor, \( L_T \), the Rayleigh optical thickness \( \delta_R \), the relative atmospheric air mass, the solar elevation angle, and the solar constant with an adjustment for the Earth Sun distance, \( K_S \).

\[
R_b = I_0 \cdot K_S \cdot e^{-\delta_R \cdot m_z} \cdot \sin(\beta_e) \]  

The estimate of diffuse radiation, \( R_d \), is a function of solar elevation angle, solar constant, direct beam radiation, as well as an atmospheric scattering coefficient, \( \xi \), and an atmospheric transmissivity factor, \( \tau_a \).
\[ R_d = 0.5 \cdot \sin(\beta_s) \cdot \zeta \left( I_0 \cdot r_s^{m} - R_b \right) \]  

(B.66)

The adjustment in solar constant for Sun Earth distance \( K_s \), is defined as a function of day angle

\[ K_s = 1 + 0.03344 \cdot \cos(\alpha_d - 2.8) \]  

(B.67)

while the solar elevation adjusted Linke turbidity factor, \( L_T \), was determined as a function of solar elevation and Linke turbidity factor, \( L_T_0 \) (which was obtained from a lookup table of regional \( L_T_0 \) values (2002)).

\[ L_T = L_T_0 - 0.85 + 2.25 \cdot \sin(\beta) - 1.11 \cdot \sin^2(\beta) \]  

(B.68)

The Rayleigh optical thickness value was a function only of the relative atmospheric air mass

\[ \delta_R = \frac{1}{0.9 \cdot m + 9.4} \]  

(B.69)

Both the atmospheric transmittance coefficient, \( r_s \), and the atmospheric scattering coefficient \( \xi \), for diffuse radiation were a function of the Linke turbidity factor and solar elevation

\[ r_s = (0.506 - 0.01079 \cdot T_s) \cdot \left[ \begin{array}{c} 1.294 \\ +2.4417 \cdot 10^{-2} \cdot \beta_s \\ -3.973 \cdot 10^{-4} \cdot \beta_s^2 \\ +3.8034 \cdot 10^{-6} \cdot \beta_s^3 \\ -2.2145 \cdot 10^{-8} \cdot \beta_s^4 \\ +5.8332 \cdot 10^{-11} \cdot \beta_s^5 \end{array} \right] + (T_L - 5) \cdot \left[ \begin{array}{c} 0.927173 \\ +1.86002 \cdot 10^{-2} \cdot \beta_s \\ -5.37651 \cdot 10^{-4} \cdot \beta_s^2 \\ +5.51224 \cdot 10^{-6} \cdot \beta_s^3 \\ -1.50178 \cdot 10^{-8} \cdot \beta_s^4 \\ -3.81556 \cdot 10^{-11} \cdot \beta_s^5 \end{array} \right] \]  

(B.70)

\[ \xi = \left[ \begin{array}{c} 1.294 \\ +2.4417 \cdot 10^{-2} \cdot \beta_s \\ -3.973 \cdot 10^{-4} \cdot \beta_s^2 \\ +3.8034 \cdot 10^{-6} \cdot \beta_s^3 \\ -2.2145 \cdot 10^{-8} \cdot \beta_s^4 \\ +5.8332 \cdot 10^{-11} \cdot \beta_s^5 \end{array} \right] + (T_L - 5) \cdot \left[ \begin{array}{c} -0.190432 \\ +1.82259 \cdot 10^{-2} \cdot \beta_s \\ -6.01334 \cdot 10^{-4} \cdot \beta_s^2 \\ +1.10146 \cdot 10^{-5} \cdot \beta_s^3 \\ -1.00432 \cdot 10^{-7} \cdot \beta_s^4 \\ +3.53849 \cdot 10^{-10} \cdot \beta_s^5 \end{array} \right] \]  

(B.71)
B.5 Model of missing short wave radiation, wind speed, and air and soil temperatures

When determinant variables such as $Q_{pg}$, or $R_n$ were available they were used as a method to more directly estimate $R_g$. However, when no data were available for modelling of missing values alternate approaches to replacement of those values were employed. Diel curves of $R_g$ were determined by month for mean diel radiation level increments of 50 W m$^{-2}$. If no data were available for an entire day the preceding day was used to estimate the radiation level. If more than one consecutive day were missing, the earliest missing days were estimated days from either 4 or 7 days preceding. Lagged correlation analysis of mean diel values of $R_g$ suggest that 1, 3.5 and 7 days preceding a day will best represent that days mean diel value of $R_g$, as shown in figure B.1.

![Figure B.1 Lagged correlation analysis of global short wave radiation using all data (heavy line) and data separated by year (thinner lines).](image)

Missing values of air temperature, soil temperature, and wind speed were also replaced using monthly and average short wave radiation incrementally averaged values.

B.6 Model of diffuse radiation

Because of the limited data availability it was necessary to determine an empirical model of diffuse radiation, $R_d$, based on measured values of global and modelled
values of potential short wave radiation. The data used in this relationship were obtained from periods when diffuse radiation measurements were deemed to be acceptable. This generally limited diffuse radiation measurements to a period of three to ten days after a successful adjustment of the sensor shade band. The assessment of a successful adjustment was necessarily quite subjective because of the nature of radiation at the Griffin site and because of the limited window of adjustment, i.e. typically a few hours around midday.

Figure B.2 Relationship of the percent diffuse short wave radiation to the percent potential short wave radiation for four solar elevation angle classes. Heavy line represents least squares fit model for data in graph, thin line is corresponding model for the greater than 50 deg. solar elevation angle class.

A comparison of percent diffuse radiation with percent of potential short wave radiation for four ranges of solar elevation angle is shown in figure B.2. It is apparent that for lower solar elevation angles the relatively clear relationship between percent diffuse radiation and percent potential short wave radiation begins to break down with higher levels of diffuse radiation corresponding to high values of percent potential radiation. It was assumed, without proper justification, that this increased scatter at low solar elevation angles was the result of instrument measurement error, and an
The empirical model was based on values for higher solar elevation angles. The model and parameter values used are given in equation B.72.

\[
\frac{R_d}{R_g} = \frac{0.981}{1.277 - 1.06 \cdot \frac{R_g}{R_p}} \quad \text{for} \quad \frac{R_g}{R_p} < 0.27
\]

\[
\frac{R_d}{R_g} = \frac{0.981}{1.277 - 1.06 \cdot \frac{R_g}{R_p}} \quad \text{for} \quad \frac{R_g}{R_p} > 0.27
\]  

**B.7 Model of albedo and reflected short wave radiation**

Estimates of missing \( R_t \) were obtained as the product of measured or estimated \( R_g \) with albedo:

\[
R_t = R_g \cdot \alpha_c
\]  

(B.73)

When values of \( R_t \) were not available, the albedo of the canopy was estimated using an equation from Dickinson (Dickinson 1983).

\[
\alpha_c = \frac{\omega \cdot \cos(\beta_s) \cdot k_c \cdot (1 + d_e \cdot (2 \cdot \beta_s - 1))}{(\alpha_c \cdot \cos(\beta_s) \cdot k_c) \cdot (1 + d_e)}
\]  

(B.74)

where

\[
d_e = \frac{1 - \omega_1}{\alpha_c}
\]  

(B.75)

\[
\alpha_c = \sqrt{1 - \omega_1} \cdot \sqrt{1 - \omega_1 + 2 \cdot \beta_s \cdot \omega_1}
\]  

(B.76)

The parameters \( \beta_d \) and \( \beta_i \) represent the upward scatter parameters for diffuse and incident radiation, the parameter \( \omega_1 \) is the light scattering term consisting of both light transmission and reflection for leaves in the canopy and the parameter \( k_c \) is the canopy extinction coefficient.
Using constant parameter values for \( \omega \), \( k \), \( \phi_0 \) and \( \beta \) this model provided a reasonable representation of the diel pattern of albedo. However, annual variations in the magnitude of albedo indicated that the coefficients were changing over the year. Annual parameter estimates were obtained by holding three of the parameters constant and performing a non-linear regression (SAS, NLIN) on the fourth parameter on data grouped by month and 20% increments of \( R_p \). No strong relationships to \( \% R_p \) were obtained but annual patterns in \( \omega \) and \( k \) that followed lognormal peak distributions were observed:

\[
\omega = 0.1777 + 0.04341 \cdot e^{0.18698}
\]

(B.77)

\[
k = 15.48 - 10.57 \cdot e^{0.243}
\]

(B.78)

In these equations the parameter \( M \) is the decimal month, which varies between a value of 1 and 13 over the course of the year.

**B.8 Model of mean CO\(_2\) concentration**

The diel pattern of normalized CO\(_2\) concentration, as shown in the left panel of figure B.3 was fitted to a four parameter Weibull distribution given by equation B.79.

\[
N_i = a_w \left( \frac{c_w - 1}{c_w} \right)^{\alpha_w} \left( \frac{\Pi_d - \Pi_{d0}}{b_w} \right) \left( \frac{c_w - 1}{c_w} \right)^{\beta_w} \cdot e^{-1} \left( \frac{\Pi_d - \Pi_{d0}}{c_w} \right) \left( \frac{c_w - 1}{c_w} \right)^{\gamma_w} \cdot e^{\frac{(c_w - 1)}{c_w}}
\]

(B.79)

where the empirical parameters obtained by non-linear fitting (SAS NLIN) had the values; \( a_w = 1.0207, b_w = 77.8366, c_w = 2.2431, \Pi_{d0} = 61.1794 \) and \( \Pi_d \) is the value of percent of day or night.
Figure B.3 Diel pattern of CO₂ concentration normalized by diel range of CO₂ concentration and grouped by midday average \( R_g \).

The pattern of diel \( C_c \) variation was then un-normalized using the relationship between the diel range of \( C_c \) and the midday average \( R_g \) shown in the right panel of figure B.3. The equation describing this relationship, given by equation B.80, is also plotted in figure B.3.

\[
C_c = \left(0.5 - \frac{R_g}{69.4143 + R_g}\right) \frac{21.0037 \cdot \overline{R_g}}{R_g} + C_{co}
\]  

(B.80)

In this relationship, \( N_c \) is obtained from equation B.79, \( \overline{R_g} \) is the midday average value of \( R_g \), and \( C_{co} \) is the monthly average atmospheric CO₂ concentration obtained from Mace Head research station observations, and given in table B.1.

Table B.1 Monthly average atmospheric CO₂ concentrations from Mace Head monitoring station. (Conway, Tans, & Waterman 2004).

<table>
<thead>
<tr>
<th>Month</th>
<th>Jan</th>
<th>Feb</th>
<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO₂ conc.</td>
<td>360.7</td>
<td>360.1</td>
<td>359.9</td>
<td>361.3</td>
<td>360.2</td>
<td>357.2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Month</th>
<th>Jul</th>
<th>Aug</th>
<th>Sep</th>
<th>Oct</th>
<th>Nov</th>
<th>Dec</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO₂ conc.</td>
<td>352.1</td>
<td>347.3</td>
<td>348.0</td>
<td>353.1</td>
<td>356.7</td>
<td>358.4</td>
</tr>
</tbody>
</table>
B.9 Model of annual variation of canopy height, zero plane displacement and canopy biomass

The annual variation of canopy height and zero plane displacement were estimated using equation B.81. The value of $h_{co}$ is the canopy height at the beginning of the year and $d_y$ is the day of year. The values of $h_{co}$ are given in table B.2.

\[
h_c = h_{co} + \frac{0.95}{1 + \left(\frac{d_y}{182}\right)^{13}} \tag{B.81}
\]

The value of zero plane displacement was then obtained from $h_c$ using equation B.82.

\[d = 0.7 \cdot h_c \tag{B.82}\]

A wind direction dependent version of the zero plane displacement, appropriate for use with data from the eddy covariance tower was defined in relation to wind direction using equation B.83.

\[d_0 = d + \sin(2.2 \cdot \theta - 37.2) \tag{B.83}\]

The annual variation of aboveground canopy biomass was estimated using equation B.81. The value of $W_{ao}$ is aboveground canopy biomass at the beginning of the year and $d_y$ is the day of year. The values of $W_{ao}$ are given in table B.2.

\[W_a = W_{ao} + 0.8775 \cdot \frac{0.8775}{1 + \left(\frac{d_y}{182}\right)^{13}} \tag{B.84}\]

Table B.2 Values of beginning of year canopy height, $h_{co}$, and beginning of year above ground canopy biomass, $W_{ao}$.

<table>
<thead>
<tr>
<th>Year</th>
<th>1997</th>
<th>1998</th>
<th>1999</th>
<th>2000</th>
<th>2001</th>
</tr>
</thead>
<tbody>
<tr>
<td>$h_{co}$, m</td>
<td>5.75</td>
<td>6.7</td>
<td>7.65</td>
<td>8.6</td>
<td>9.55</td>
</tr>
<tr>
<td>$W_{ao}$, kg m$^{-2}$</td>
<td>5.57</td>
<td>6.45</td>
<td>7.33</td>
<td>8.21</td>
<td>9.08</td>
</tr>
</tbody>
</table>
B.10 Carbon dioxide flux model for unsupported gap filling

To fill carbon dioxide fluxes when no supporting information was available a date/time based model was employed, with coefficients determined by non-linear fitting (SAS, NLIN) using existing data.

\[
F_c = F_{co} + a_o \cdot e^{-0.5 \left( \frac{t - t_0}{t_o} \right)^2}
\]  

(B.85)

where \( t \) is the time in decimal hours and \( t_0 = 0.515 \) is a constant. The other parameters are defined as a function of fractional month, \( M \), using the following equations.

\[
F_{co} = 1.2775 + 5.2906 \cdot e^{-0.5 \left( \frac{M - 7.2064}{2.2623} \right)^{2.9065}}
\]  

(B.86)

\[
a_o = 7.6341 - 30.68 \cdot e^{-0.5 \left( \frac{M - 6.5602}{4.1956} \right)^2}
\]  

(B.87)

\[
b_o = 0.0337 + 0.1552 \cdot e^{-0.5 \left( \frac{M - 6.1723}{3.0112} \right)^{1.924}}
\]  

(B.88)

B.11 Primary statistics

In the processing of eddy covariance time series data, signal statistics were calculated using the following formulae. In the processing of other data sets different formulae may have been used, depending upon the software package used to perform the calculations.

Signal means

\[
\bar{\chi} = \frac{\sum_{i=1}^{n} \chi_i}{n}
\]  

(B.89)
Signal standard deviations

\[ \sigma_x = \sqrt{\frac{\sum_{i=1}^{n} (x_i - \bar{x})^2}{n-1}} \]  
\[ \text{(B.90)} \]

Signal variance

\[ \sigma_x^2 = \frac{\sum_{i=1}^{n} (x_i - \bar{x})^2}{n-1} \]  
\[ \text{(B.91)} \]

Signal skewness

\[ \text{Skewness} = \frac{\sum_{i=1}^{n} (x_i - \bar{x})^3}{n \cdot \sigma_x^3} \]  
\[ \text{(B.92)} \]

Signal kurtosis

\[ \text{Kurtosis} = \frac{\sum_{i=1}^{n} (x_i - \bar{x})^4}{n \cdot \sigma_x^4} - 3 \]  
\[ \text{(B.93)} \]

**B.12 Gas conversions**

General gas species conversions are in terms of an atmospheric component species designated with the subscript \( X \). If this subscript is not used the conversion is to be applied to atmospheric water vapour. Some common variables include air temperature, \( T_a \) (units of degrees C), air pressure, \( P \) (units of kPa), ideal gas constant, \( R_g \), species magnitude conversion factor, \( a_x \).

**B.12.1 Absolute density**

Absolute density in units of g m\(^3\) calculated from partial pressure, \( P_x \) in units of kPa.

\[ \rho_x = \frac{M_X \cdot P_x \cdot 1000}{R_g \cdot (T_a + 273.16)} \]  
\[ \text{(B.94)} \]
Absolute density in units of g m\(^{-3}\) calculated from concentration \(C_x\).

\[
\rho_x = \frac{C_x \cdot M_x \cdot P \cdot 1000}{a_x \cdot R_g \cdot (T_s + 273.16)}
\] (B.95)

Absolute density of water vapour in units of g m\(^{-3}\) calculated from wet bulb temperature in degrees C.

\[
\rho_v = \frac{M_v \cdot \left(1000 \cdot e_s - \left[0.66 \cdot (1 + 0.00115 \cdot T_w)\right] \cdot P \cdot (T - T_w)\right)}{R_g \cdot (T + 273.16)}
\] (B.96)

**B.12.2 Concentration**

Concentration calculated from molecular density (units of Moles m\(^{-3}\)).

\[
C_x = \frac{a_x \cdot R_g \cdot \theta_x \cdot (T_s + 273.16)}{1000 \cdot P}
\] (B.97)

Concentration calculated from partial pressure (units of kPa).

\[
C_x = \frac{P_x \cdot a_x}{P}
\] (B.98)

**B.12.3 Partial pressure**

Calculation of partial pressure (units of kPa) from concentration

\[
P_x = \frac{C_x \cdot P}{a_x}
\] (B.99)

Calculation to partial pressure (units of kPa) of water vapour from wet bulb temperature (units degrees C).

\[
e = e_s - 0.00066 \cdot P \cdot (1 + 0.00115 \cdot T_s) \cdot (T_s - T_w)
\] (B.100)

**B.12.4 Relative humidity**

Relative humidity calculated from partial pressure (units of kPa).
\[ \hat{h}_s = \frac{e \cdot 100}{e_s} \]  
\[(B.101)\]

### B.12.5 Density of moist air

Density of moist air calculated from partial pressures (units of kPa).

\[ \rho_s = \frac{M_a \cdot (P - 0.378 \cdot e) \cdot 1000}{R_g \cdot (T_a + 273.16)} \]  
\[(B.102)\]

### B.12.6 Saturation vapour pressure

Saturation water vapour pressures (units of kPa) calculated from air temperature, (units of degrees C). The second equation (B.104) is the enhancement factor for equilibrium of water vapour with a mixture of gases (1996).

\[ e'_s = 0.61121 \cdot e \left( \frac{17.502 \cdot T_a}{T_a + 246.97} \right) \]  
\[(B.103)\]

\[ e_s = e'_s \left( 1.00072 + \frac{P \cdot (3.2 + 0.00059 \cdot T_a^2)}{100000} \right) \]  
\[(B.104)\]

### B.12.7 Vapour pressure deficit

Vapour pressure deficit, (units of kPa) calculated from partial pressures (units of kPa).

\[ D = e_s - e \]  
\[(B.105)\]

### B.13 Miscellaneous calculations

#### B.13.1 Flux conversion factors

The following equations define the factors used to convert covariances of vertical velocity with scalars to energy flux values. Equation B.106 converts the covariance of vertical velocity and absolute density of water vapour to latent heat flux.
Equation B.107 and its subsequent definitions are used to convert the covariance of vertical velocity and air temperature to sensible heat flux. In this set of equations \( \rho_v \) refers to absolute density of water vapour.

\[
\rho_v \cdot c_p = \frac{c_{p\text{-moist}} \cdot \rho_v + c_{p\text{-dry}} \cdot (\rho_v - \rho_v)}{1000}
\]  \hspace{1cm} (B.107)

\[
C_{p\text{-dry}} = 1005.0 + \frac{(T_s + 23.12)^2}{3364.0}
\]  \hspace{1cm} (B.108)

\[
C_{p\text{-moist}} = 1859.0 + 0.13 \cdot h_t + T_s \cdot \left( \frac{193 + 5.69 \cdot h_t}{1000} \right) + T_s^2 \cdot \left( \frac{1 + 0.05 \cdot h_t}{1000} \right)
\]  \hspace{1cm} (B.109)

**B.13.2 Virtual temperature**

Calculation of virtual temperature from true air temperature, both in units of degrees C.

\[
T_v = \left( \frac{T_s + 273.16}{1 - \left( \frac{1 - 0.622}{0.622} \cdot \frac{e_s}{P} \right) e_s} \right) - 273.16
\]  \hspace{1cm} (B.110)

**B.13.3 Stability - Monin Obukhov**

The Monin-Obukhov stability factor was calculated using the following equation.

\[
\frac{z}{L} = \frac{0.4 \cdot (z - d) \cdot 9.81 \cdot H}{(T_v + 273.16) \cdot \rho_v \cdot C_p \cdot u^*^3}
\]  \hspace{1cm} (B.111)
B.14 Frequency response attenuation calculations

The following equations are employed in determination of frequency response corrections.

B.14.1 Tube attenuation coefficient

A coefficient describing the attenuation of the concentration fluctuations of a gas species as travels through a tube, \( G_T \), is described by equation B.112,

\[
G_T = \frac{4 \cdot \pi^2 \cdot l_T \cdot r_T \cdot \Lambda}{V_T^2}
\]  

(B.112)

where \( l_T \) is the tube length in meters, \( r_T \) is the tube radius in units of meters. The tube flow velocity \( V_T \) (units of m s\(^{-1}\)) is calculated from \( r_T \) and tube flow rate \( Q_t \) (units of LPM).

\[
V_T = \frac{Q_t}{1000 \cdot 60 \cdot \pi \cdot r_T^2}
\]  

(B.113)

For laminar flow (Re < 2300) the value of lambda was obtained following Leuning and Moncrieff (1990).

\[
\Lambda = \frac{0.0104 \cdot \nu \cdot Re}{D_e}
\]  

(B.114)

In which, Re is the Reynolds number, \( \nu \) is the viscosity of air, and \( D_e \) is the diffusivity of air. The Reynolds number was calculated from \( r_T \), \( V_T \), and viscosity

\[
Re = \frac{2 \cdot r_T \cdot V_T \nu}{D_e}
\]  

(B.115)

Where viscosity, \( \nu \), is calculated from atmospheric pressure, \( P \)

\[
\nu = (3.719815 - (0.0358737 \cdot P)) + \frac{0.0001342 \cdot P^2}{100000.0}
\]  

(B.116)

and diffusivity, \( D_e \), has a constant, species dependent, value.
For turbulent flow \((\text{Re} > 2300)\) the value of \(\Lambda\) was obtained from work by Massman (1991). If no value is specified for \(\Lambda\) and the Reynolds number is greater than 2300, a value for \(\Lambda\) was calculated from a polynomial, fit to the data of Massman (1991), of the form:

\[
\Lambda = c_0 + \frac{c_1}{\text{Re}} + \frac{c_2}{\text{Re}^2} + \frac{c_3}{\text{Re}^3} + \frac{c_4}{\text{Re}^4}
\]

where

for \(\text{H}_2\text{O}\)

\[
\begin{align*}
c_0 &= 0.490851295350703 \\
c_1 &= -12093.7897865916 \\
c_2 &= 325341636.671577 \\
c_3 &= -150205321310.87 \\
c_4 &= 2244127770065660.0
\end{align*}
\]

for \(\text{CO}_2\)

\[
\begin{align*}
c_0 &= 0.977616284023198 \\
c_1 &= -40511.5062176583 \\
c_2 &= 742694287.9160520 \\
c_3 &= -3394928055533.75 \\
c_4 &= 4991471292069380.0
\end{align*}
\]

The use of the polynomial fit in determining \(\Lambda\) resulted in inaccuracies that are dependent upon the value of Reynolds number for the specified tube and flow conditions, but were considered small.
B.14.2 Conversion of natural to normalized spectral frequency

Normalized spectral frequencies were used in frequency response attenuation and spectral models. The normalized spectral frequencies were obtained from natural frequencies, \( n \), height above zero plane displacement \( (z-d) \) and mean wind speed, \( U \).

\[
f = n \cdot \frac{z-d}{U}
\]  

(B.119)

B.14.3 Frequency response correction transfer functions

The transfer function for vector path averaging was obtained from Moore (1986).

\[
T_{PV} = \sqrt{\frac{2}{2\pi nlU}} \left( 2 + e^{\frac{2\pi nl}{U}} \right) - 3 \cdot \left( 1 - e^{\frac{2\pi nl}{U}} \right)
\]  

(B.120)

The transfer function for scalar path averaging was obtained from Moore (1986).

\[
T_{PS} = \sqrt{\frac{3 + e^{\frac{2\pi nl}{U}} - \frac{4}{2\pi nlU} \left( 1 - e^{\frac{2\pi nl}{U}} \right)}}
\]  

(B.121)

The transfer function for sensor time response was obtained from Moore (1986).

\[
T_i = \frac{1}{\sqrt{1 + (2\pi nl)^2}}
\]  

(B.122)

The transfer function for sensor frequency response mismatch was obtained from Zeller et al. (1989).
The transfer function for tube attenuation was obtained from Massman (1991).

\[ T_{\text{TA}} = e^{-\frac{\alpha}{U^2}} \]  \hspace{1cm} (B.124)

The transfer function for path separation between sensors was obtained from Moore (1986).

\[
T_s = \begin{cases} 
  e^{-\frac{\alpha}{U^2}} & \frac{n\cdot l}{U} \leq 4 \\
  0 & \frac{n\cdot l}{U} > 4 
\end{cases} \hspace{1cm} (B.125)
\]

The transfer function for block time averaging was obtained from Lenschow et al. (1994).

\[ T_B = 1 - \left( \frac{\sin \left( \frac{\pi \cdot n \cdot t_s}{\pi \cdot n \cdot t_s} \right)}{\pi \cdot n \cdot t_s} \right)^2 \hspace{1cm} (B.126) \]

**B.14.4 Equivalent frequency response correction transfer function time constants**

The following equivalent time constant values were obtained from Massman (2000).

The time constant for sonic anemometer horizontal line averaging for momentum flux is calculated as,

\[ \tau_{\text{PV1}} = \frac{l}{2.8 \cdot U} \hspace{1cm} (B.127) \]
The time constant for sonic anemometer vertical line averaging for momentum flux is calculated as,

\[ \tau_{PH} = \frac{l_s}{5.7 \cdot U} \]  

(B.128)

The time constant for sonic anemometer line averaging for scalar flux is calculated as,

\[ \tau_{PV2} = \frac{l_s}{8.4 \cdot U} \]  

(B.129)

The time constant for lateral sensor separation is calculated as,

\[ \tau_s = \frac{l_s}{1.1 \cdot U} \]  

(B.130)

The time constant for longitudinal sensor separation is calculated as,

\[ \tau_{lon} = \frac{l_s}{1.05 \cdot U} \]  

(B.131)

The time constant for longitudinal sensor separation of a first order instrument is calculated as,

\[ \tau_{lon1} = \frac{4}{3} \left( \frac{l_s}{2 \cdot U + \tau_l} \right) \]  

(B.132)

The time constant for a line averaging scalar sensor is calculated as,

\[ \tau_{PS} = \frac{l_s}{4.0 \cdot U} \]  

(B.133)

The time constant for volume averaging right circular cylinder is calculated as,

\[ \tau_{CM} = \left( 0.2 + 0.4 \left( \frac{2 \cdot r}{l_s} \right) \right) \left( \frac{l_s}{U} \right) \]  

(B.134)

The time constant for tube attenuation in turbulent flow is calculated as,
\[ \tau_{TA} = \left( \frac{\Lambda r}{\sqrt{\frac{l_x}{0.83} \left( \frac{l_y}{V_z} \right)}} \right) \]  

(B.135)

The time constant for spherical sampling volume averaging is calculated as,

\[ \tau_{sph} = \frac{r}{U} \]  

(B.136)

The time constant for high pass block averaging is calculated as,

\[ \tau_B = \frac{t_r}{2.8} \]  

(B.137)

**B.15 Spectral models**

Unless otherwise specified, cospectral models were obtained from Kaimal *et al.* (1972). Model cospectra were calculated for one hundred normalized frequency values defined by equation B.138.

\[ f(0) = 0.00001 \]

\[ f(i) = \begin{array}{ll}
99 & 1.2 \cdot f_{i-1} \\
i = 1 & 
\end{array} \]  

(B.138)

For neutral or unstable stabilities, scalar flux cospectra were calculated using equation B.139.

\[ \frac{f \cdot C_{ww}(f)}{w'x'} = \begin{array}{ll}
12.92 f & \text{for } f \leq 0.54 \\
\left(1 + 26.7 f\right)^{1.375} & \\
4.378 f & \text{for } f > 0.54 \\
\left(1 + 3.8 f\right)^{2.4} & 
\end{array} \]  

(B.139)

For neutral or unstable stabilities, momentum flux cospectra were calculated using equation B.140.
\[
\frac{f \cdot C_{ww}(f)}{u'w'} = \begin{cases} 
20.78 f & \text{for } f \leq 0.24 \\
(1 + 31f)^{1.575} & \\
12.66 f & \text{for } f > 0.24 \\
(1 + 9.6f)^{2.4}
\end{cases}
\]

(B.140)

For stable conditions, scalar flux cospectra were calculated using equation B.141.

\[
\frac{f \cdot C_{wx}(f)}{w'x'} = \frac{f}{\left[0.284 \cdot \left(1 + 6.4 \frac{u}{L}\right)^{0.75}\right] + \left[2.34 \cdot \left(0.284 \cdot \left(1 + 6.4 \frac{u}{L}\right)^{0.75}\right)^{-1.1}\right]} f^{2.1}
\]

(B.141)

For stable conditions, momentum flux cospectra were calculated using equation B.142.

\[
\frac{f \cdot C_{uw}(f)}{u'w'} = \frac{f}{\left[0.124 \cdot \left(1 + 7.9 \frac{u}{L}\right)^{0.75}\right] + \left[2.34 \cdot \left(0.124 \cdot \left(1 + 7.9 \frac{u}{L}\right)^{0.75}\right)^{-1.1}\right]} f^{2.1}
\]

(B.142)
Appendix C  Plot measurements

This appendix contains a table of measurements associated with each of the plots used in determination of site DBH. Sites with no associated with Measurements or Dates were not used. Locations 18 and 31 had a larger number of associated sample plot locations, which are referred to by the plot numbers in brackets. The date column refers only to the year(s) in which the measurements were made. Further details on the dates and times of measurements for plots 18 and 31 can be obtained from Appendix E.

Table C.1  List of sample plots, measurements made in the plots and the date range of those measurements.

<table>
<thead>
<tr>
<th>Plot</th>
<th>Measurements</th>
<th>Dates</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>2</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>3</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>4</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Litter</td>
<td>1999 - 2001</td>
</tr>
<tr>
<td></td>
<td>Branch photosynthesis</td>
<td>2000 - 2001</td>
</tr>
<tr>
<td></td>
<td>Sap flow</td>
<td>2000 - 2001</td>
</tr>
<tr>
<td></td>
<td>Isotopic measurement</td>
<td>2000 - 2001</td>
</tr>
<tr>
<td></td>
<td>Soil respiration</td>
<td>2000 - 2001</td>
</tr>
<tr>
<td></td>
<td>Water table</td>
<td>2000 - 2001</td>
</tr>
<tr>
<td></td>
<td>Soil water nutrients</td>
<td>2000</td>
</tr>
<tr>
<td>5</td>
<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>6, 7</td>
<td>No measurements</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>9</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>10</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Stem flow and through fall</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>(both amounts and nitrogen content)</td>
<td>2000</td>
</tr>
<tr>
<td>11</td>
<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>13</td>
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<td>1997</td>
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<td>14</td>
<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>15</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Litter</td>
<td>1998 - 2001</td>
</tr>
<tr>
<td>16</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Litter</td>
<td>1998 - 2001</td>
</tr>
<tr>
<td></td>
<td>Soil cores</td>
<td>2000</td>
</tr>
<tr>
<td>17</td>
<td>DBH</td>
<td>1997, 2001</td>
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<tr>
<td></td>
<td>Shoot photosynthesis</td>
<td>1997</td>
</tr>
<tr>
<td>17.5</td>
<td>Radon flux</td>
<td>2000 - 2001</td>
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<td>Measurement Details</td>
<td>Time Period</td>
</tr>
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<td>-----</td>
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<td>-------------</td>
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<tr>
<td>18</td>
<td>DBH</td>
<td>1997, 2001</td>
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<tr>
<td>19</td>
<td>DBH</td>
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<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>21</td>
<td>DBH</td>
<td>1997</td>
</tr>
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<td>22</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>23</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>24</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Stem flow and through fall (both amounts and nitrogen content)</td>
<td>2000</td>
</tr>
<tr>
<td>25</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Litter</td>
<td>1998 - 2001</td>
</tr>
<tr>
<td>26-28</td>
<td>No measurements</td>
<td></td>
</tr>
<tr>
<td>29</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>30</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>31</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>(51-54)</td>
<td>Litter</td>
<td>1998 - 2001</td>
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<tr>
<td></td>
<td>Respiration</td>
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</tr>
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<td></td>
<td>Radiation (above and below canopy)</td>
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<td>Air Temperature</td>
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<td></td>
<td>CO2 concentration</td>
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<td></td>
<td>Humidity</td>
<td>1996 - 2001</td>
</tr>
<tr>
<td></td>
<td>Wind speed</td>
<td>1996 - 2001</td>
</tr>
<tr>
<td></td>
<td>Wind direction</td>
<td>1996 - 2001</td>
</tr>
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<td></td>
<td>Soil moisture</td>
<td>2000 - 2001</td>
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<td>Soil cores</td>
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<td>Tree height</td>
<td>1997, 2001</td>
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<td></td>
<td>Leaf area</td>
<td>1998</td>
</tr>
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<td>32</td>
<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>33</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td></td>
<td>Soil cores</td>
<td></td>
</tr>
<tr>
<td>34,35</td>
<td>No measurements</td>
<td></td>
</tr>
<tr>
<td>36</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>37</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
<tr>
<td>38</td>
<td>DBH</td>
<td>1997, 2001</td>
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<tr>
<td></td>
<td>Stem flow and through fall (both amounts and nitrogen content)</td>
<td>2000</td>
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<tr>
<td>39</td>
<td>No measurements</td>
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</tr>
<tr>
<td>40</td>
<td>DBH</td>
<td>1997, 2001</td>
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<td>41</td>
<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>42</td>
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<td></td>
</tr>
<tr>
<td>43</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
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<tr>
<td>---</td>
<td>---</td>
<td>---</td>
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<tr>
<td>44</td>
<td>DBH</td>
<td>1997, 2001</td>
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<td>45</td>
<td>DBH</td>
<td>1997, 2001</td>
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<tr>
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<td>Sap flow</td>
<td>2000-2001</td>
</tr>
<tr>
<td>46</td>
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<td>1997</td>
</tr>
<tr>
<td>47</td>
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<td></td>
</tr>
<tr>
<td>48</td>
<td>DBH</td>
<td>1997</td>
</tr>
<tr>
<td>49</td>
<td>DBH</td>
<td>1997, 2001</td>
</tr>
</tbody>
</table>
Appendix D  Instrumentation and equipment

This appendix contains the tables of instrumentation and equipment used in the Griffin field experiment. For each sensor/instrument the manufacturer and item serial number are given and where appropriate sensor calibration coefficients are given. For sensors with varying calibration coefficients, the coefficients employed for real-time monitoring of the signal are listed. Table D.1 shows sensors and related equipment while Table D.1 lists other equipment used in the experiment.

Table D.1  List of sensors used in Griffin Experiment. Manufacturer, serial number, and sensor calibration equations are also given.

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Measured variable(s)</th>
<th>Calibration</th>
<th>Model/ Serial #(#s)</th>
<th>Manufacturer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sonic anemometer</td>
<td>wind vectors, air temperature, wind direction, signal A/D conversion</td>
<td>$u = 0.01 \cdot V$, $v = 0.01 \cdot V$, $w = 0.01 \cdot V$</td>
<td>1012R2 / #083</td>
<td>Gill Instruments</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$T = \left( \frac{0.02 \cdot V}{402.7} \right)^2$</td>
<td>1012R2 / #058</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Units $u$, $v$, $w$: $m \cdot s^{-1}$, $T$: °C, $V$: mV</td>
<td>1012R2 / #020</td>
<td></td>
</tr>
<tr>
<td>Cup anemometer</td>
<td>Wind speed - analogue</td>
<td>$U = 0.00421 \cdot \left( \frac{V}{0.2041} \right)$</td>
<td>A100 / #2647</td>
<td>Vector Instruments</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Denominator signifies effect of voltage divider.</td>
<td></td>
<td></td>
</tr>
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</table>

523
<table>
<thead>
<tr>
<th>( U )</th>
<th>( \frac{V}{0.2043} )</th>
<th>A100 / #2648</th>
</tr>
</thead>
<tbody>
<tr>
<td>( U )</td>
<td>( \frac{V}{0.2040} )</td>
<td>A100 / #2649</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{V}{0.2044} )</td>
<td>A100E / #2741</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{V}{0.2037} )</td>
<td>A100E / #2742</td>
</tr>
</tbody>
</table>

Wind speed - digital

<table>
<thead>
<tr>
<th>( U )</th>
<th>( \frac{\text{counts}}{I_s} )</th>
<th>Vector Instruments</th>
</tr>
</thead>
<tbody>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>A100R / #2005</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>A100R / #2120</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>A100R / #2121</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>A100R / #2122</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>A100R / #1033</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>A100R / #1848</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>MK3B / #276</td>
</tr>
<tr>
<td>( U )</td>
<td>( \frac{\text{counts}}{I_s} )</td>
<td>Cassella and Co. Ltd.</td>
</tr>
<tr>
<td>Instrument</td>
<td>Measurement Type</td>
<td>Formula</td>
</tr>
<tr>
<td>------------------------------------</td>
<td>-------------------</td>
<td>---------</td>
</tr>
<tr>
<td>Wind vane</td>
<td>Wind direction</td>
<td>$\theta = 360 \cdot \left( \frac{V}{V_{\text{ref}}} \right)$</td>
</tr>
<tr>
<td>Tipping Bucket rain gauge</td>
<td>Precipitation</td>
<td>$p_s = 0.46 \cdot \text{counts}$</td>
</tr>
<tr>
<td>Rain gauge</td>
<td>Precipitation</td>
<td>Direct measurement using graduated cylinder</td>
</tr>
<tr>
<td>Psychrometer (Thermistor)</td>
<td>Air temperature</td>
<td>$T = \frac{V}{1000}$</td>
</tr>
<tr>
<td>Platinum resistance thermometer</td>
<td>Soil temperature</td>
<td>Conversion according to DIN 43760 standard</td>
</tr>
</tbody>
</table>

**Formulas:**

- Wind vane: $U = 0.0572 \cdot \left( \frac{\text{counts}}{I_s} \right)$
- Tipping Bucket rain gauge: $p_s = 0.46 \cdot \text{counts}$
- Platinum resistance thermometer: $T = \frac{-R_o \cdot A + \sqrt{(R_o \cdot A)^2 - 4 \cdot B \cdot (R_o - R_f)}}{2 \cdot R_o \cdot B}$

**Conversion format obtained from Honeywell corp.**
<table>
<thead>
<tr>
<th>Thermo-couple</th>
<th>Air temperature</th>
<th>$T = \frac{V}{1000}$</th>
<th>Multiple probes but no serial #s available.</th>
<th>University of Edinburgh</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thermo-couple</td>
<td>Air temperature Soil temperature Bole temperature</td>
<td>Referenced to thermistor and converted using campbell datalogger software</td>
<td>RS components</td>
<td></td>
</tr>
<tr>
<td>Thermistor (thermocouple reference)</td>
<td>Reference temperature for thermocouples</td>
<td>Profile Bole 1 Bole 2 Soil Probe 1 Soil Probe 2 Soil Probe 3 Soil Probe 4</td>
<td>BetaTherm</td>
<td></td>
</tr>
<tr>
<td>IRGA (Conc. of H₂O, CO₂ cell pressure and cell temperature)</td>
<td>CO₂</td>
<td>$C_C = 250 + 50 \cdot \frac{V}{1000}$</td>
<td>LI6262 / #762 #660 #480 #350</td>
<td>Licor</td>
</tr>
<tr>
<td></td>
<td>H₂O</td>
<td>$C_Q = 6.0 \cdot \frac{V}{1000}$</td>
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</tr>
<tr>
<td></td>
<td>T</td>
<td>$T_{cell} = 0.01221 \cdot \frac{V}{1000}$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
\[ P = 58.974 + 0.01528 \frac{V}{1000} \]

<table>
<thead>
<tr>
<th>Pyranometer</th>
<th>Global short wave radiation</th>
<th>CM2 #662813</th>
<th>Kipp and Zonen</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>( R_g = 86.22 \cdot V )</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( R_g = 85.25 \cdot V )</td>
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<tr>
<td>Reflect short wave radiation</td>
<td>( R_g = 87.82 \cdot V )</td>
<td>CM2 #683239</td>
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<td>( R_g = 86.43 \cdot V )</td>
<td>CM5 #860240</td>
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<tr>
<td>Silicon cell sensor</td>
<td>Global PPFD</td>
<td>SD101QV #25</td>
<td>Licor</td>
</tr>
<tr>
<td></td>
<td>( Q_{pg} = 103.8 \cdot V )</td>
<td></td>
<td>Macam</td>
</tr>
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<td>( Q_{pg} = 1 \cdot V )</td>
<td>SD101QV #</td>
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<tr>
<td>Reflect PPFD</td>
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<td></td>
<td>( Q_{pr} = 173.8 \cdot V )</td>
<td>UoE #4</td>
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<td>( Q_{pr} = 101.7 \cdot V )</td>
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<td>APAR</td>
<td>( Q_{pt} = 102.4 \cdot V )</td>
<td>SD101QV #21</td>
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<td>( Q_{pt} = 98.4 \cdot V )</td>
<td>SD101QV #22</td>
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<td>( Q_{pt} = 99.6 \cdot V )</td>
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<td>$Q_{pt} = 101.4 \cdot V$</td>
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<td>Soil heat flux</td>
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<td>$Q_{pt} = 165.1 \cdot V$</td>
<td>UoE #39</td>
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<td>$Q_{pt} = 283.4 \cdot V$</td>
<td>UoE #108</td>
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<td>$Q_{pt} = 349.9 \cdot V$</td>
<td>UoE #112</td>
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<td>$Q_{pt} = 364.4 \cdot V$</td>
<td>UoE #131</td>
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<td>UoE #142</td>
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<td>$Q_{pt} = 191.8 \cdot V$</td>
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<td>$G = 39.1 \cdot V$</td>
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<td>HFT3 7906 / #H943020</td>
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<td>Measurement</td>
<td>Formula</td>
<td>Manufacturer</td>
</tr>
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<td>------------------</td>
<td>----------------------</td>
<td>------------------------------</td>
<td>---------------------------------------------------</td>
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<tr>
<td>TDR probes</td>
<td>Soil moisture</td>
<td>$\theta_v = -0.187 + 0.037 \cdot \tau + 0.335 \cdot \tau^2$</td>
<td>CS615 Campbell Scientific</td>
</tr>
<tr>
<td>Cable Tester</td>
<td>Soil moisture</td>
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<td>1502C Textronics</td>
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<td>Net Radiometers</td>
<td>Net radiation</td>
<td>$R_n = 13.8 \cdot V$</td>
<td>Q*6 REBS</td>
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<td>$R_n = 13.7 \cdot V$</td>
<td>#89003</td>
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<td>#87114</td>
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<td></td>
<td>$R_n = 13.7 \cdot V$</td>
<td>#92229</td>
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</tbody>
</table>
|                 |                      | $R_n = \begin{cases} 
11.44 \cdot V & \text{if } R_n \leq 0 \\
7.57 \cdot V & \text{if } R_n > 0 
\end{cases}$ | Q*7 #94119                                        |
<p>| Gauge Bara.     | Atmospheric pressure | $P = 60.0 + 0.0184 \cdot V$ | PTB101B Campbell Scientific                       |
|                  |                      |                              | #R0940048                                         |</p>
<table>
<thead>
<tr>
<th>Equipment Type</th>
<th>Description</th>
<th>Formula</th>
<th>Supplier</th>
<th>Model/Serial Number</th>
<th>Manufacturer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Differential Barometer</td>
<td>Pitot tube reference for cup anemometer calibration</td>
<td>$\Delta P = 0.02 \cdot V$</td>
<td>FCB 8810135</td>
<td>Furness Controls Ltd.</td>
<td></td>
</tr>
<tr>
<td>Shadow band</td>
<td>Diffuse radiation</td>
<td></td>
<td></td>
<td>University of Edinburgh</td>
<td></td>
</tr>
<tr>
<td>Mass flow controller</td>
<td>Sample air flow rate measurement and control</td>
<td></td>
<td>FC2901 #AA057039</td>
<td>Tylan General</td>
<td></td>
</tr>
<tr>
<td>Inclinometer</td>
<td>Tree height</td>
<td></td>
<td></td>
<td>Silva</td>
<td></td>
</tr>
<tr>
<td>Height pole</td>
<td>Tree height</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Scale</td>
<td>Weight determination (soil moisture, bulk density)</td>
<td></td>
<td>2204</td>
<td>Sartorius</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Weight determination (litter mass)</td>
<td></td>
<td>PJ360</td>
<td>Mettler</td>
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</tr>
<tr>
<td></td>
<td>Weight determination (carbon and nitrogen content)</td>
<td></td>
<td></td>
<td>AD-4 Auto-balance</td>
<td>Perkin-Elmer</td>
</tr>
<tr>
<td>C/N analyser</td>
<td>Carbon content</td>
<td></td>
<td>EMASyst 1106</td>
<td>Carlo Erba Strumentazione</td>
<td></td>
</tr>
<tr>
<td>--------------</td>
<td>----------------</td>
<td>----------------</td>
<td>------------------</td>
<td>--------------------------</td>
<td></td>
</tr>
<tr>
<td>DBH tape</td>
<td>Bole diameter at breast height</td>
<td></td>
<td></td>
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<tr>
<td>Data Logger</td>
<td>Low frequency data acquisition and calibration control</td>
<td></td>
<td>CR10X / #E1810 #EX07984</td>
<td>Campbell Scientific</td>
<td></td>
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<tr>
<td></td>
<td>Low frequency data acquisition</td>
<td></td>
<td>CR21X #2421 #E1018 #E1205/#OSX0 #E1211</td>
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<tr>
<td></td>
<td>Low frequency data acquisition and calibration control</td>
<td></td>
<td>CR10X #</td>
<td></td>
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<tr>
<td></td>
<td>Low frequency data acquisition and power system control</td>
<td></td>
<td>CR21X #</td>
<td></td>
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<tr>
<td>Dew point generator</td>
<td>Water vapour source for calibration</td>
<td></td>
<td>LI610 #DPG 278B</td>
<td>Licor Inc.</td>
<td></td>
</tr>
</tbody>
</table>
Table D.2 List of equipment used in Griffin Experiment. Manufacturer, serial number, and comment//description are given in adjacent columns.

<table>
<thead>
<tr>
<th>Equipment</th>
<th>Purpose</th>
<th>Comments</th>
<th>Model/Serial #</th>
<th>Manufacturer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tower</td>
<td>Instrument deployment</td>
<td></td>
<td>25G</td>
<td>Rohn</td>
</tr>
<tr>
<td>Wind turbine</td>
<td>Power generation</td>
<td>650 Watt, 3 blade turbine mounted at 12 m on guyed pole.</td>
<td>WT600</td>
<td>Proven engineering</td>
</tr>
<tr>
<td>Solar Panels</td>
<td>Power generation</td>
<td>Six, 75 Watt panels. Pairs of panels were connected in parallel with the three pairs connected in series to produce 24 volt output supply to 24 volt battery bank.</td>
<td>BP 275</td>
<td>BP solar</td>
</tr>
<tr>
<td>Generator</td>
<td>Power generation</td>
<td>1.2 kW generator converted to run on propane. Modified for autostart (see below)</td>
<td>GPX1100</td>
<td>Honda/Bellingham</td>
</tr>
<tr>
<td>Inverter/Charger</td>
<td>Power control</td>
<td></td>
<td>DR2424E #V7345</td>
<td>Trace Engineering</td>
</tr>
<tr>
<td>Generator Auto-start</td>
<td>Automated generator control</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>AC/DC converters</td>
<td>12 volt power supply</td>
<td></td>
<td>250-1246</td>
<td>RS</td>
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<tr>
<td>High capacity</td>
<td>24V Power</td>
<td>120 A hr, deep discharge lead acid batteries.</td>
<td>DL 33, 34</td>
<td>Varta</td>
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<tr>
<td>Component</td>
<td>Description</td>
<td>Storage</td>
<td>Manufacturer/Supplier</td>
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<td>---------------------------------</td>
<td>-----------------------------------------------------------------------------</td>
<td>-----------</td>
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<tr>
<td>Batteries</td>
<td>Storage</td>
<td>35, 51, 61, 63</td>
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<td>Compass</td>
<td>Plot layout Instrument layout and orientation</td>
<td>Ranger</td>
<td>Silva</td>
<td></td>
</tr>
<tr>
<td>Soil corer</td>
<td>Soil core extraction</td>
<td>Giddings Machine Company</td>
<td>University of Edinburgh</td>
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<tr>
<td>Eddy Covariance system</td>
<td>Eddy covariance flux measurements</td>
<td>University of Edinburgh</td>
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<tr>
<td>Multiplexer</td>
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<td>AM416 #2238 #4538 #4559 #4561</td>
<td>Campbell Scientific</td>
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<td>CSM1 #E1154</td>
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<td>High frequency data acquisition and processing</td>
<td>Armada 1110</td>
<td>Compaq</td>
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<td>Contura 420 CX</td>
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<td>Low frequency data retrieval</td>
<td>Contura</td>
<td>Compaq</td>
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</table>
Appendix E  Instrumentation time lines

This appendix contains timelines of instrument and equipment installation periods and where appropriate, system maintenance information is provided. For each installation, an ID of the instrument installed is given above the line designating the period of installation.

The timelines are broken down by the associated logging systems. The loggers are given in the following order:

1. Eddy covariance logging system
2. Flux tower data logger
3. Profile tower microclimate data logger instrument set #1
4. Profile tower microclimate data logger instrument set #2
5. Soil/tree temperature logger
6. Profile tower CO₂/H₂O logger
7. Power system control logger
<table>
<thead>
<tr>
<th>Time Line</th>
<th>Repaired</th>
<th>Rebuilt</th>
<th>Repair (swapped)</th>
<th>Repaired</th>
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<td>2649</td>
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<td>Wind Speed 5</td>
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<td>Wind Speed 6</td>
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<td>Wind Speed 7</td>
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<td>Wind Speed 8</td>
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<td>Repair</td>
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<td>Temp Air 3</td>
<td>Move from intake</td>
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<td>Temp Air 4</td>
<td>Move from intake</td>
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<td>BP 75W</td>
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<td>Honda 1200 (New Engine)</td>
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<tr>
<td></td>
<td>1200</td>
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**Note:** The timeline includes events such as precipitation, surface elevation, and various monitoring activities with specific dates and references.
# Appendix F Software

## F.1 Software packages

The following table contains the list of software used in the Griffin experiment. The primary use of the software, the software manufacturer and version number of the software used are specified. A comparison of software calculation quality for two primary packages employed in this experiment is given in section F.2.

### Table F.1 List of software used in the Griffin field experiment. The use of the software, its version number and the software’s manufacturer are given in adjacent columns.

<table>
<thead>
<tr>
<th>Software</th>
<th>Use</th>
<th>Version</th>
<th>Manufacturer</th>
</tr>
</thead>
<tbody>
<tr>
<td>PC208</td>
<td>Low frequency sampling data logger programming, signal processing and data retrieval</td>
<td></td>
<td>Campbell Scientific</td>
</tr>
<tr>
<td>Edisol</td>
<td>High frequency data sampling, signal processing and data storage</td>
<td>DOS</td>
<td>University of Edinburgh</td>
</tr>
<tr>
<td>EdiRe</td>
<td>High frequency data signal processing</td>
<td>1.4.129</td>
<td>University of Edinburgh</td>
</tr>
<tr>
<td>SAS</td>
<td>Data management</td>
<td>8e</td>
<td>SAS institute</td>
</tr>
<tr>
<td></td>
<td>Data processing</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Statistical analysis</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Data visualization</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Excel</td>
<td>Data processing</td>
<td>97</td>
<td>Microsoft</td>
</tr>
<tr>
<td></td>
<td>Statistical analysis</td>
<td>2000</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Data visualization</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sigma Plot</td>
<td>Figure preparation</td>
<td>5.0,</td>
<td>Jandel Scientific</td>
</tr>
<tr>
<td></td>
<td></td>
<td>8.0</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Data analysis</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>Data visualization</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Delphi</td>
<td>EdiRe and utility software Programming</td>
<td>4</td>
<td>Borland</td>
</tr>
<tr>
<td>DocToHelp</td>
<td>EdiRe manual preparation</td>
<td></td>
<td>Wextech</td>
</tr>
<tr>
<td>Intel Math libraries</td>
<td>EdiRe development</td>
<td>4.1 - 4.2</td>
<td>Intel</td>
</tr>
</tbody>
</table>

545
<table>
<thead>
<tr>
<th>Software</th>
<th>Category</th>
<th>Version</th>
<th>Developer</th>
</tr>
</thead>
<tbody>
<tr>
<td>HexEdit</td>
<td>File editing</td>
<td>1.0</td>
<td>Andrew W. Phillips</td>
</tr>
<tr>
<td>Boxer</td>
<td>File editing</td>
<td>10</td>
<td>Boxer software</td>
</tr>
<tr>
<td>Notepad</td>
<td>File editing</td>
<td>5.1</td>
<td>Microsoft</td>
</tr>
<tr>
<td>Corel</td>
<td>Figure preparation, Data analysis</td>
<td>7</td>
<td>Corel</td>
</tr>
<tr>
<td>MathCad</td>
<td>Theoretical development</td>
<td>2001</td>
<td>MathSoft Engineering &amp; Education, Inc.</td>
</tr>
<tr>
<td>Word</td>
<td>Document preparation</td>
<td>97 2000</td>
<td>Microsoft</td>
</tr>
<tr>
<td>Kedit</td>
<td>File editing</td>
<td>1.0</td>
<td>Mansfield Software Group Inc.</td>
</tr>
<tr>
<td>Access</td>
<td>Data management, Data processing, Data analysis</td>
<td>97 2000</td>
<td>Microsoft</td>
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<tr>
<td>IDRISI</td>
<td>Data analysis</td>
<td></td>
<td>Clark Laboratories</td>
</tr>
<tr>
<td>Reference Manager</td>
<td>Document preparation</td>
<td>10</td>
<td>ISI ResearchSoft</td>
</tr>
</tbody>
</table>
F.2 Computational validity of statistics

A test of the calculation accuracy of two frequently used software packages was performed following (Cook et al. 1999). In this test, means and standard deviations calculated for a reference data set were compared with the expected value using the quality metric:

\[ P = \log_{10} \left( 1 + \frac{|y_{\text{calc}} - y_{\text{ref}}|}{K \cdot \eta \cdot |y_{\text{ref}}|} \right) \]

where \( y_{\text{calc}} \) and \( y_{\text{exp}} \) represent the calculated and expected value, \( K \) is a difficulty factor, and \( \eta \) is the computational precision of the software. The values \( y_{\text{exp}} \) and \( K \) were identical for both software while \( \eta \) for Excel was calculated to be \( 7.11 \times 10^{-15} \) and was set to a more restrictive value of \( 1.00 \times 10^{-16} \) for EdiRe. For the EdiRe test, calculation results were output with a minimum of 15 decimal places. For these results, values of less than \( 10^{-15} \) were assumed to be equal to \( 10^{-16} \), representing the resolution limit of the 16 byte real values used to perform calculations in EdiRe. The results are given in figure F.1. In the figure, larger values can be interpreted as poorer performance. EdiRe performed better than Excel in all but one test, but did exhibit errors when calculating the standard deviations of the values with large means. The range of observed errors was considered to be negligible in relation to other errors associated with the data analysis performed.

Figure F.1 Quality metric of mean and standard deviation calculations for Excel and EdiRe software packages.
F.2 EdiRe reprocessing procedure list

The following text is the procedure processing list used by EdiRe in the main reprocessing of the Griffin flux data. Other procedure lists were used for more specialized results.

Location Output Files
Output File Calculations = C:\Flux procs\Flux Reprocess\Output\Output For Thesis 3.csv
Output File Spectral =
Output File Wavelet =
Output File Cross Correlation =
Output File Distribution =
Output File Quadrant =
Output File Reference =

Preprocessed Files
File <0> = C:\Flux procs\Flux Reprocess\Input\EdiCal.csv
File <1> =
File <2> =
File <3> =
File <4> =
File <5> =
File <6> =
File <7> =
File <8> =
File <9> =

Set Values
From Time =
To Time =
Number of Variables = 9
Storage Label = Press
Assignment value = 97.0 <0> Press
Storage Label = RH
Assignment value = 80 <0> RH_avg
Storage Label = e
Assignment value = 0.6 <0> e
Storage Label = Woff
Assignment value = 0 <0> ALPHA
Storage Label = PFalpha
Assignment value = 0 <0> BETA
Storage Label = PFbeta
Assignment value = 0 <0> GAMMA
Storage Label = Coff
Assignment value = 250 <0> C_Flx_off
Storage Label = Cgain
Assignment value = 60 <0> C_Flx_gain
Storage Label = Qoff
Assignment value = 0 <0> Q_Flx_off_new

Set Values
From Time =
To Time =
Number of Variables = 3
Storage Label = ZeroPlaneDisp
Assignment value = 5 <0> ZpD
Storage Label =
Assignment value =

548
Storage Label =
Assignment value =

Extract
From Time =
To Time =
Channel = 1
Label for Signal = U

Extract
From Time =
To Time =
Channel = 2
Label for Signal = V

Extract
From Time =
To Time =
Channel = 3
Label for Signal = W

Extract
From Time =
To Time =
Channel = 4
Label for Signal = T

Extract
From Time =
To Time =
Channel = 5
Label for Signal = C

Linear
From Time =
To Time =
Signal = U
1st Offset = 0
1st Gain = 0.01
1st Curvature = 0
2nd Offset = 0
2nd Gain = 1
2nd Curvature = 0

Linear
From Time =
To Time =
Signal = V
1st Offset = 0
1st Gain = 0.01
1st Curvature = 0
2nd Offset = 0
2nd Gain = 1
2nd Curvature = 0

Linear
From Time =
To Time =
Signal = W
1st Offset = 0
1st Gain = 0.01
1st Curvature = 0
2nd Offset = 0
2nd Gain = 1
2nd Curvature = 0

Sonic Temperature
From Time =
To Time =
Signal = T
1st Offset = 0
1st Gain = 0.02
1st Curvature = 0
2nd Offset = 0
2nd Gain = 1
2nd Curvature = 0

Linear
From Time =
To Time =
Signal = C
1st Offset = 0
1st Gain = 0.001
1st Curvature = 0
2nd Offset = Coff
2nd Gain = Cgain
2nd Curvature = 0

Linear
From Time =
To Time =
Signal = Q
1st Offset = 0
1st Gain = 0.001
1st Curvature = 0
2nd Offset = Qoff
2nd Gain = 6
2nd Curvature = 0

Attack Angle Correction - Polynomial
From Time =
To Time =
Signal (u) = U
Signal (v) = V
Signal (w) = W
Iterations = 2
Sine coefs - positive angles =
Sine coefs - negative angles =
Cosine coefs =
Wind direction coefs =
Coefficient delimiter =

Wind direction
From Time =
To Time = 12/01/1998 13:30:00
Signal (u) = U
Signal (v) = V
Orientation = 0
Wind Direction Components = U+N_V+E
Wind Direction Output = N_0_deg-E_90_deg
Storage Label Wind Direction = Wind_Direction
Storage Label Wind Dir Std Dev = SD_Wind_Direction

Wind direction
From Time = 12/01/1998 13:30:00
To Time = 01/01/2002
Signal (u) = U
Signal (v) = V  
Orientation = 180  
Wind Direction Components = U+N_V+E  
Wind Direction Output = N_0_deg-E_90_deg  
Storage Label Wind Direction = Wind_Direction  
Storage Label Wind Dir Std Dev = SD_Wind_Direction  

Sonic Temperature Path  
From Time =  
To Time =  
Signal (T) = T  
Signal (u) = U  
Signal (v) = V  
Signal (w) = W  
Transducer configuration = 1 non-vertical path  
Top transducer orientation, deg from u = -30  
Transducer zenith angle, deg = 45  
Transducer subsampling interval, sec = 0.006  

Rotation - Planar  
From Time =  
To Time =  
Signal (u) = U  
Signal (v) = V  
Signal (w) = W  
w offset = Woff  
Planar Alpha = PFalpha  
Planar Beta = PFbeta  

Cross Correlate  
From Time =  
To Time =  
Signal = W  
Signal which lags = C  
Correlation type = Covariance  
Output Correlation curve =  
Storage Label Peak Time =  
Storage Label Peak Value =  

Remove Lag  
From Time =  
To Time =  
Signal = C  
Min Lag (sec) = 6  
Lag (sec) =  
Max Lag (sec) = 8  
Below Min default (sec) =  
Above Max default (sec) =  

Cross Correlate  
From Time =  
To Time =  
Signal = W  
Signal which lags = Q  
Correlation type = Covariance  
Output Correlation curve =  
Storage Label Peak Time =  
Storage Label Peak Value =  

Remove Lag  
From Time =  
To Time =  
Signal = Q  
Min Lag (sec) = 6  
Lag (sec) =
Max Lag (sec) = 25
Below Min default (sec) =
Above Max default (sec) =

1 chn statistics
From Time =
To Time =
Signal = U
Storage Label Mean = MeanU
Storage Label Std Dev = StdDevU
Storage Label Skewness = SkewU
Storage Label Kurtosis = KurtU
Storage Label Maximum = MaxU
Storage Label Minimum = MinU
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

1 chn statistics
From Time =
To Time =
Signal = V
Storage Label Mean = MeanV
Storage Label Std Dev = StdDevV
Storage Label Skewness = SkewV
Storage Label Kurtosis = KurtV
Storage Label Maximum = MaxV
Storage Label Minimum = MinV
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

1 chn statistics
From Time =
To Time =
Signal = W
Storage Label Mean = MeanW
Storage Label Std Dev = StdDevW
Storage Label Skewness = SkewW
Storage Label Kurtosis = KurtW
Storage Label Maximum = MaxW
Storage Label Minimum = MinW
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

1 chn statistics
From Time =
To Time =
Signal = T
Storage Label Mean = MeanTv
Storage Label Std Dev = StdDevTv
Storage Label Skewness = SkewTv
Storage Label Kurtosis = KurtTv
Storage Label Maximum = MaxTv
Storage Label Minimum = MinTv
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

1 chn statistics
From Time =
To Time =
Signal = Q
Storage Label Mean = MeanQ
Storage Label Std Dev = StdevQ
Storage Label Skewness = SkewQ
Storage Label Kurtosis = KurtQ
Storage Label Maximum = MaxQ
Storage Label Minimum = MinQ
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

1 chn statistics
From Time =
To Time =
Signal = C
Storage Label Mean = MeanCconc
Storage Label Std Dev = StdevCconc
Storage Label Skewness = SkewCconc
Storage Label Kurtosis = KurtCconc
Storage Label Maximum = MinCconc
Storage Label Minimum = MinCconc
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

Partial pressure
From Time =
To Time =
Storage Label = ef
Apply to =
Apply by =
Variable type = Concentration
Measured variable = MeanQ
Min or QC =
Max or QC =
Temperature (C) = MeanTv
Min or QC =
Max or QC =
Pressure (KPa) = Press
Min or QC =
Max or QC =
Molecular weight (g/mole) = 18
Conc conv factor = 1000

Virtual Temperature Raw
From Time =
To Time =
Signal T(C) = T
Signal H2O = Q
Pressure, kPa = Press
Water vapour units = Concentration, mmol/mol
Temperature conversion = Calculate true from virtual-sonic

1 chn statistics
From Time =
To Time =
Signal = T
Storage Label Mean = MeanT
Storage Label Std Dev = StdevT
Storage Label Skewness = SkewT
Storage Label Kurtosis = KurtT
Storage Label Maximum = MaxT
Storage Label Minimum = MinT
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

Sensible heat flux coefficient
From Time =
To Time =
Storage Label = rhoCp
Apply to =
Apply by =
Vapour pressure (KPa) = e
Min or QC =
Max or QC =
Temperature (C) = MeanT
Min or QC =
Max or QC =
Pressure (KPa) = Press
Min or QC =
Max or QC =
Alternate rhoCp = 1230

Latent heat of evaporation
From Time =
To Time =
Storage Label = L
Apply to =
Apply by =
Temperature (C) = MeanT
Min or QC =
Max or QC =
Pressure (KPa) = Press
Min or QC =
Max or QC =
LE flux coef, L = 2500

User defined fast
From Time =
To Time =
Equation = Q*(Press*18.01)/(8.314*(MeanT+273.16))
Number of signals = 1
Signal = Q
Variable = MeanT
Variable = Press
Variable =

User defined fast
From Time =
To Time =
Equation = C*Press*1000/(8.314*(MeanT+273.16))
Number of signals = 1
Signal = C
Variable = MeanT
Variable = Press
Variable =

1 chn statistics
From Time =
To Time =
Signal = Q
Storage Label Mean = MeanQrho
Storage Label Std Dev = StdevQrho
Storage Label Skewness = SkewQrho
Storage Label Kurtosis = KurtQrho
Storage Label Maximum = MaxQrho
Storage Label Minimum = MinQrho
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

1 chn statistics
From Time =
To Time =
Signal = C
Storage Label Mean = MeanC
Storage Label Std Dev = StdevC
Storage Label Skewness = SkewC
Storage Label Kurtosis = KurtC
Storage Label Maximum = MaxC
Storage Label Minimum = MinC
Storage Label Variance =
Storage Label Turbulent Intensity =
Alt Turbulent Intensity Denominator =

Tube attenuation
From Time =
To Time =
Storage Label = C atten
Apply to =
Apply by =
Gas species = CO2
Tube pressure (KPa) = 90
Min or QC = 70
Max or QC = 110
Flow rate (LPM) = 6.0
Tube length (m) = 18.0
Tube ID (m) = 0.00615
Lambda coefficient = 3.5

Tube attenuation
From Time =
To Time =
Storage Label = Q atten
Apply to =
Apply by =
Gas species = H2O
Tube pressure (KPa) = 90
Min or QC = 70
Max or QC = 110
Flow rate (LPM) = 6.0
Tube length (m) = 18.0
Tube ID (m) = 0.00615
Lambda coefficient = 3.5

User defined
From Time =
To Time =
Storage Label = Qtimeconst
Apply to =
Apply by =
Equation = TAN(0.000053407*EXP(RH/13.8))/0.0062832
Variable = RH
Variable =
Variable =
Variable =

2 chn statistics
From Time =
To Time =
Signal = W

555
Signal = T
Storage Label Covariance =
Storage Label Correlation =
Storage Label Flux = H
Flux coefficient = \rho C_p

2 chn statistics
From Time =
To Time =
Signal = W
Signal = U
Storage Label Covariance = UW
Storage Label Correlation =
Storage Label Flux =
Flux coefficient =

2 chn statistics
From Time =
To Time =
Signal = W
Signal = Q
Storage Label Covariance =
Storage Label Correlation =
Storage Label Flux = LE
Flux coefficient = L

2 chn statistics
From Time =
To Time =
Signal = W
Signal = C
Storage Label Covariance =
Storage Label Correlation =
Storage Label Flux = Fc
Flux coefficient = \textit{L}

Comments
Comment =
Comment =
Comment =

Set Values
From Time =
To Time =
Number of Variables = 3
Storage Label = Hc
Assignment value = H
Storage Label = UWc
Assignment value = UW
Storage Label =
Assignment value =

Set Value
From Time =
To Time =
Storage Label = counter
Assignment value = 0

Comments
Comment =
Comment =
Comment =

U star
From Time =
To Time =
Storage Label = Ustr
Apply to =
Apply by =
uw covariance (m^2/s^2) = UWc
Min or QC =
Max or QC =
Stability - Monin Obhukov
From Time =
To Time =
Storage Label = ZoL
Apply to =
Apply by =
Measurement height (m) = 15.5
Zero plane displacement (m) = ZeroPlaneDisp
Virtual Temperature (C) = MeanT
Min or QC =
Max or QC =
H flux (W/m^2) = Hc
Min or QC =
Max or QC =
H flux coef, RhoCp = rhoCp
Min or QC =
Max or QC =
Scaling velocity (m/s) = Ustr
Min or QC =
Max or QC =
User defined
From Time =
To Time =
Storage Label = SP_uw
Apply to =
Apply by =
Equation = IIF(ZoL<0, 0.04, IIF(ZoL>4, 0.04, 0.08-0.01*ZoL))
Variable = ZoL
Variable = MeanU
Variable =
User defined
From Time =
To Time =
Storage Label = SP_wx
Apply to =
Apply by =
Equation = IIF(ZoL<0, 0.08, 0.08+0.13*ZoL)
Variable = ZoL
Variable = MeanU
Variable =
User defined
From Time =
To Time =
Storage Label = uw_broad
Apply to =
Apply by =
Equation = IIF(ZoL<0,0.65,IIF(ZoL>1,0.36,0.65-0.29*ZoL))
Variable = ZoL
Variable =
User defined
From Time =
To Time =
Storage Label = uw_Nu
Apply to =
Equation = IIF(ZoL<0, 1, IIF(ZoL>1, 2.4, 1 + 1.4*ZoL))
Variable = ZoL
User defined
From Time =
To Time =
Storage Label = wt_broad
Apply to =
Apply by =
Equation = IIF(ZoL>0, 0.4, 0.33)
Variable = ZoL
User defined
From Time =
To Time =
Storage Label = wt_Nu
Apply to =
Apply by =
Equation = IIF(ZoL>0, 2.1, 3)
Variable = ZoL
Load co-spectra
From Time =
To Time =
Load spectra of type = Model Horst/Massman
Spectral model = UW
Monin-Ohbukov stability = ZoL
Wind speed (m/s) = MeanU
Measurement ht (m) = 15.5
Zero plane (m) = ZeroPlaneDisp
Boundary layer ht (m) =
Spectral reference tag =
Reference condition =
Reference normalization =
Spectral peak frequency = SP_uw
Spectral broadness parameter = uw_broad
Spectral gain parameter = uw_Nu
Frequency response
From Time =
To Time =
Storage Label = UWFresp
Apply to =
Apply by =
Correction type = UW
Measurement height (m) = 15.5
Zero plane displacement (m) = ZeroPlaneDisp
Boundary layer height (m) = 1000
Stability Z/L = ZoL
Wind speed (m/s) = MeanU
Sensor 1 Flow velocity (m/s) = MeanU
Sensor 1 Sampling frequency (Hz) = 20.8333
Sensor 1 Low pass filter type =
Sensor 1 Low pass filter time constant =
Sensor 1 High pass filter type =
Sensor 1 High pass filter time constant =
Sensor 1 Path length (m) = 0.15
Sensor 1 Time constant (s) =
Sensor 1 Tube attenuation coef =

558
Sensor 2 Flow velocity (m/s) = MeanU
Sensor 2 Sampling frequency (Hz) = 20.8333
Sensor 2 Low pass filter type =
Sensor 2 Low pass filter time constant =
Sensor 2 High pass filter type =
Sensor 2 High pass filter time constant =
Sensor 2 Path length (m) = 0.15
Sensor 2 Time constant (s) =
Sensor 2 Tube attenuation coef =
Path separation (m) =
Get spectral data type = Loaded
Get response function from = model
Reference Tag =
Reference response condition =
Sensor 1 subsampled = x
Sensor 2 subsampled = x
Apply velocity distribution adjustment =
Use calculated distribution =
Velocity distribution std dev =
Stability distribution std dev =

Load co-spectra
From Time =
To Time =
Load spectra of type = Model Horst/Massman
Spectral model = WX
Monin-Obukov stability = ZoL
Wind speed (m/s) = MeanU
Measurement height (m) = 15.5
Zero plane (m) = ZeroPlaneDisp
Boundary layer height (m) =
Spectral reference tag =
Reference condition =
Reference normalization =
Spectral peak frequency = SP_wx
Spectral broadness parameter = wt_broad
Spectral gain parameter = wt_Nu

Frequency response
From Time =
To Time =
Storage Label = HFresp
Apply to =
Apply by =
Correction type = WX
Measurement height (m) = 15.5
Zero plane displacement (m) = ZeroPlaneDisp
Boundary layer height (m) = 1000
Stability Z/L = ZoL
Wind speed (m/s) = MeanU
Sensor 1 Flow velocity (m/s) = MeanU
Sensor 1 Sampling frequency (Hz) = 20.8333
Sensor 1 Low pass filter type =
Sensor 1 Low pass filter time constant =
Sensor 1 High pass filter type =
Sensor 1 High pass filter time constant =
Sensor 1 Path length (m) = 0.15
Sensor 1 Time constant (s) =
Sensor 1 Tube attenuation coef =
Sensor 2 Flow velocity (m/s) = MeanU
Sensor 2 Sampling frequency (Hz) = 20.8333
Sensor 2 Low pass filter type =
Sensor 2 Low pass filter time constant =
Sensor 2 High pass filter type =
Sensor 2 High pass filter time constant =
Sensor 2 Path length (m) = 0.15
Sensor 2 Time constant (s) =
Sensor 2 Tube attenuation coefficient =
Path separation (m) =
Get spectral data type = Loaded
Get response function from = model
Reference Tag =
Reference response condition =
Sensor 1 subsampled = x
Sensor 2 subsampled = x
Apply velocity distribution adjustment =
Use calculated distribution =
Velocity distribution standard deviation =
Stability distribution standard deviation =

Frequency response
From Time =
To Time =
Storage Label = LEFresp
Apply to =
Apply by =
Correction type = WX
Measurement height (m) = 15.5
Zero plane displacement (m) = ZeroPlaneDisp
Boundary layer height (m) = 1000
Stability Z/L = ZoL
Wind speed (m/s) = MeanU
Sensor 1 Flow velocity (m/s) = MeanU
Sensor 1 Sampling frequency (Hz) = 20.8333
Sensor 1 Low pass filter type =
Sensor 1 Low pass filter time constant =
Sensor 1 High pass filter type =
Sensor 1 High pass filter time constant =
Sensor 1 Path length (m) = 0.15
Sensor 1 Time constant (s) =
Sensor 1 Tube attenuation coefficient =
Sensor 2 Flow velocity (m/s) = MeanU
Sensor 2 Sampling frequency (Hz) = 20.8333
Sensor 2 Low pass filter type =
Sensor 2 Low pass filter time constant =
Sensor 2 High pass filter type =
Sensor 2 High pass filter time constant =
Sensor 2 Path length (m) = 0.01
Sensor 2 Time constant (s) = Qtimeconst
Sensor 2 Tube attenuation coefficient = Q atten
Path separation (m) = 0.1
Get spectral data type = Loaded
Get response function from = model
Reference Tag =
Reference response condition =
Sensor 1 subsampled = x
Sensor 2 subsampled = x
Apply velocity distribution adjustment =
Use calculated distribution =
Velocity distribution standard deviation =
Stability distribution standard deviation =

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Frequency response
From Time =
To Time =
Storage Label = FcFresp
Apply to =
Apply by =
Correction type = WX
Measurement height (m) = 15.5
Zero plane displacement (m) = ZeroPlaneDisp
Boundary layer height (m) = 1000
Stability Z/L = ZoL
Wind speed (m/s) = MeanU
Sensor 1 Flow velocity (m/s) = MeanU
Sensor 1 Sampling frequency (Hz) = 20.8333
Sensor 1 Low pass filter type =
Sensor 1 Low pass filter time constant =
Sensor 1 High pass filter type =
Sensor 1 High pass filter time constant =
Sensor 1 Path length (m) = 0.15
Sensor 1 Time constant (s) =
Sensor 1 Tube attenuation coef =
Sensor 2 Flow velocity (m/s) = MeanU
Sensor 2 Sampling frequency (Hz) = 20.8333
Sensor 2 Low pass filter type =
Sensor 2 Low pass filter time constant =
Sensor 2 High pass filter type =
Sensor 2 High pass filter time constant =
Sensor 2 Path length (m) = 0.01
Sensor 2 Time constant (s) = 0.1
Sensor 2 Tube attenuation coef = C atten
Path separation (m) = 0.1
Get spectral data type = Loaded
Get response function from = model
Reference Tag =
Reference response condition =
Sensor 1 subsampled = x
Sensor 2 subsampled = x
Apply velocity distribution adjustment =
Use calculated distribution =
Velocity distribution std dev =
Stability distribution std dev =
Mathematical operation
From Time =
To Time =
Storage Label = UWc
Apply to =
Apply by =
Measured variable A = UW
Operation = *
Measured variable B = UWFresp
Mathematical operation
From Time =
To Time =
Storage Label = Hc
Apply to =
Apply by =
Measured variable A = H
Operation = *
Measured variable B = HFresp

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Mathematical operation
From Time =
To Time =
Storage Label = LEc
Apply to =
Apply by =
Measured variable A = LE
Operation = *
Measured variable B = LEFresp

Mathematical operation
From Time =
To Time =
Storage Label = Fcc
Apply to =
Apply by =
Measured variable A = Fc
Operation = *
Measured variable B = FcFresp

Mathematical operation
From Time =
To Time =
Storage Label = counter
Apply to =
Apply by =
Measured variable A = counter
Operation = +
Measured variable B = 1

Repeat Previous Until
From Time =
To Time =
Previous items to repeat = 19
Maximum repeat iterations = 4
Repeat test 1 variable = counter
Repeat test 1 operator = >=
Repeat test 1 (lower limit) = 2
Repeat test 1 upper limit =
Repeat test union = Ignore test 2
Repeat test 2 variable =
Repeat test 2 operator =
Repeat test 2 (lower limit) =
Repeat test 2 upper limit =
Number of User defined equations in repeat = 6
Appendix G  Signal quality control ranges

This appendix specifies the ranges of values that were considered acceptable for the purposes of quality control of logger and flux system data.

Table G.1 Variable quality control ranges used with Griffin dataset.

<table>
<thead>
<tr>
<th>Signal</th>
<th>Variable</th>
<th>Mean</th>
<th>Standard Deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Minimum</td>
<td>Maximum</td>
</tr>
<tr>
<td>Global short wave radiation</td>
<td>$R_g$</td>
<td>-50</td>
<td>1600</td>
</tr>
<tr>
<td>Reflected short wave radiation</td>
<td>$R_r$</td>
<td>-200</td>
<td>200</td>
</tr>
<tr>
<td>Diffuse short wave radiation</td>
<td>$R_d$</td>
<td>-50</td>
<td>2000</td>
</tr>
<tr>
<td>Global PAR</td>
<td>$Q_{pp}$</td>
<td>-50</td>
<td>3000</td>
</tr>
<tr>
<td>Reflected PAR</td>
<td>$Q_{pr}$</td>
<td>-50</td>
<td>500</td>
</tr>
<tr>
<td>Surface PAR</td>
<td>$Q_{ps}$</td>
<td>-50</td>
<td>3000</td>
</tr>
<tr>
<td>Net radiation</td>
<td>$R_n$</td>
<td>-500</td>
<td>800</td>
</tr>
<tr>
<td>Air Temperature</td>
<td>$T_a$</td>
<td>-20</td>
<td>30</td>
</tr>
<tr>
<td>Wet bulb temperature</td>
<td>$T_w$</td>
<td>-20</td>
<td>30</td>
</tr>
<tr>
<td>Soil Temperature 5 cm</td>
<td>$T_s$</td>
<td>-5</td>
<td>30</td>
</tr>
<tr>
<td>Bole Temperature</td>
<td>$T_b$</td>
<td>-5</td>
<td>30,100</td>
</tr>
<tr>
<td>Water vapour concentration</td>
<td>$C_w$</td>
<td>0</td>
<td>25, 30</td>
</tr>
<tr>
<td>Vapour pressure</td>
<td>$e$</td>
<td>0</td>
<td>6</td>
</tr>
<tr>
<td>Vapour pressure deficit</td>
<td>$D$</td>
<td>0</td>
<td>6</td>
</tr>
<tr>
<td>CO2 concentration</td>
<td>$C_c$</td>
<td>275</td>
<td>650,1000</td>
</tr>
<tr>
<td>Wind speed</td>
<td>$U$</td>
<td>0, -2</td>
<td>30</td>
</tr>
<tr>
<td>Parameter</td>
<td>Symbol</td>
<td>Lower Limit</td>
<td>Upper Limit</td>
</tr>
<tr>
<td>---------------------------------</td>
<td>--------</td>
<td>-------------</td>
<td>-------------</td>
</tr>
<tr>
<td>Wind direction</td>
<td>$\theta$</td>
<td>0</td>
<td>360</td>
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<tr>
<td>Soil Moisture - TDR</td>
<td>$\theta_v$</td>
<td>0</td>
<td>20</td>
</tr>
<tr>
<td>Precip</td>
<td>$P_e$</td>
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<td>30</td>
</tr>
<tr>
<td>Pressure</td>
<td>$P$</td>
<td>90, 80</td>
<td>120</td>
</tr>
<tr>
<td>CO2 flux</td>
<td>$F_c$</td>
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<td>50</td>
</tr>
<tr>
<td>Sensible heat flux</td>
<td>$H$</td>
<td>-400</td>
<td>800</td>
</tr>
<tr>
<td>Latent heat flux</td>
<td>$\lambda E$</td>
<td>-100</td>
<td>800</td>
</tr>
<tr>
<td>Soil Heat Flux</td>
<td>$G$</td>
<td>-50</td>
<td>50</td>
</tr>
<tr>
<td>Momentum</td>
<td>$u'w'$</td>
<td>-3</td>
<td>3</td>
</tr>
<tr>
<td>Friction velocity</td>
<td>$u*$</td>
<td>0</td>
<td>3</td>
</tr>
<tr>
<td>Latent heat of evaporation</td>
<td>$\lambda$</td>
<td>2000</td>
<td>3000</td>
</tr>
<tr>
<td>Sensible heat flux coefficient</td>
<td>$\rho c_p$</td>
<td>1000</td>
<td>1500</td>
</tr>
<tr>
<td>Licor Cell temperature</td>
<td>$T_{cell}$</td>
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<td>100</td>
</tr>
<tr>
<td>Licor Cell pressure</td>
<td>$P_{cell}$</td>
<td>0</td>
<td>150</td>
</tr>
<tr>
<td>Licor flow rate</td>
<td>$Q_{cell}$</td>
<td>0</td>
<td>10</td>
</tr>
<tr>
<td>Licor intake tube temperature</td>
<td>$T_{tube}$</td>
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<td>30</td>
</tr>
<tr>
<td>Power usage</td>
<td></td>
<td>-5</td>
<td>50</td>
</tr>
<tr>
<td>Power production - solar</td>
<td></td>
<td>-5</td>
<td>50</td>
</tr>
<tr>
<td>Power production - wind</td>
<td></td>
<td>-5</td>
<td>50</td>
</tr>
<tr>
<td>Power production - wind and solar</td>
<td></td>
<td>-5</td>
<td>50</td>
</tr>
<tr>
<td>Battery Bank Voltage</td>
<td></td>
<td>0</td>
<td>50</td>
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</table>
H.1 Abstract

The correction of temperature measurements derived from sonic anemometry is reviewed. Additional terms are incorporated which adjust sensible heat fluxes derived from sonic temperature for flow acceleration and sensor head distortion. Respectively, these additional terms produce an adjustment to sensible heat flux of \( -4.8 \pm 1.5 \) W m\(^{-2}\) and \( +7.5 \pm 1.9 \) W m\(^{-2}\), for the data set tested. These compare with the standard adjustment for the fluctuations of water vapour and sensor path deflection of \( -7.3 \pm 1.4 \) W m\(^{-2}\). When these corrections are combined the resulting net adjustment of \( +2.7 \pm 1.3 \) W m\(^{-2}\) is observed.

H.2 Introduction

Sonic anemometers have simplified the field of micrometeorological flux measurement. They provide a durable and precise instrument with which the velocity signals required to calculate fluxes of both momentum and scalar quantities. A number of sensor configurations have been established in order to minimise weaknesses in the inherent design restrictions (Kaimal and Finnigan, 1994). Even so, most sonic anemometer designs allow for the measurement of all three wind component vectors within close proximity, and at a frequency greater than those at which most flux density exists. These measurement capabilities and the desirable quality of minimal user maintenance have resulted in widespread use of this sensor as a primary, eddy covariance measurement device.

Although sonic anemometers provide an excellent platform for measuring the fluxes of momentum, additional sensors are required to measure the fluxes of scalars. Such sensors must be installed in close proximity to the sonic
anemometer in order to avoid loss of higher frequency signals due to sensor separation (Moore, 1986). The distance within which such a sensor may be mounted is limited by its potential for alteration of the wind velocity components and mounting practicalities. Temperature constitutes a primary measurement because it determines the flux of sensible heat, which is a component of the energy budget as well as numerous derived variables and correction terms. Because of this importance, its accurate measurement is critical. Temperature is often measured with the use of fine-wire thermocouples (TC), platinum resistance thermometers (PRT), or derived from measurements of the speed of sound. While accurate temperature measurements can be made with both thermocouples and platinum resistance thermometers, they are both subject to effects of thermal inertia and solar heating at low wind speeds and are prone to breakage at high wind speeds (Jacobs and McNaughton, 1994), (Kaimal and Finnigan, 1994). In addition, precipitation interception and condensation/rime can alter their response characteristics (Schmitt K.F. et al., 1978), (Lawson and Cooper, 2001).

Sonic temperature measurements exploit the dependence of the speed of sound on the density, and hence temperature, of the air through which the sound travels. Because a sonic anemometer measures the travel times of sound pulses through air, it provides the ideal sensor with which to measure speed of sound derived air temperature. It has been pointed out that the resulting temperature measurement must be corrected to obtain a true air temperature (Schotanus et al., 1983). Errors in the measurement of sonic temperature due to deflection of the sound path and variations in the air density as a result of water vapour fluctuations have been described and correction values presented in earlier works on this subject (Schotanus et al., 1983) (Kaimal and Gaynor, 1991), (Hignett, 1992).

In addition to the now standard corrections to sonically derived temperature for advective sonic path lengthening and density fluctuations, Grelle & Lindroth (1996) have observed a wind speed related error. They observed, for a Gill Solent anemometer under high wind speeds, that sensible heat fluxes determined using sonic temperature were more negative than those
determined using a PRT. Their evidence demonstrates that the difference between the sonic and PRT temperatures occurred at all frequencies. They also show that this error persists in cospectra, implying that it is well correlated with vertical velocity. Their compiled fluxes for a distribution of wind speeds show a wind velocity threshold of about 7 m/s, above which the error in sensible heat flux was increasingly more pronounced. The authors also state that similar errors, with similar thresholds, were observed in other research with METEK, and Kaijo Denki anemometers. They conclude that the error resulted from deformation of the sonic anemometer's sensor head under high wind speeds and suggest that wind tunnel tests are required to ascertain an appropriate correction term.

We have observed similar differences in the comparison of sensible heat fluxes obtained using sonic, $H_{so}$, and thermocouple, $H_{tc}$, temperature measurements. Coincident high wind speeds suggested that this difference was also the result of the deformation of the sonic anemometer's head. In this paper we present further evidence in support of this hypothesis. The theory and geometry of the sonic anemometer employed in this study in relation to the potential cause of this error are reviewed and we present evidence supporting this theory. In addition, a further refinement of the sound path error is included which accounts for asynchronous sound pulse sampling. A suggested correction is applied to experimental data and its effect on resulting fluxes are analysed.

**H.3 Materials and methods**

Field measurements of sensible heat fluxes and supporting variables were collected at the Griffin EUROFLUX experiment site, latitude 56° 36'30"N, longitude 3° 47' 15"W. The site vegetation is a monoculture of Sitka Spruce (*Picea sitchensis*) that was planted in 1980-1981. The canopy height at the time of measurement was approximately 7 m, with a LAI of about 8. Data for this analysis were taken from the year 1998.
The eddy covariance instrumentation and processing methods follow the specifications laid out in Aubinet et al. (2000). Wind velocity components and the sonic temperature signals used in this experiment were measured using a Gill Solent 101R2 sonic anemometer, as part of a flux measurement system (Moncrieff et al., 1997). The anemometer was mounted at the top of a tri-pole tower at a height of 15.5 m. Coincident fast-response temperature measurements were obtained from a fine wire thermocouple (Krovetz et al., 1988) mounted with a 30 cm separation from the centre of the sonic anemometer. The thermocouple wire had a diameter of 0.004 cm, which lasted from hours to weeks depending upon environmental conditions. Thermocouple data were digitised using analogue input channels of the sonic anemometer. Raw data signals were collected at 20.833 Hz and stored on CD for further analysis.

Post processing of the data was carried out using the EdiRe software package developed by the author. In post processing, velocity signals were rotated using methods described by McMillen (1998), but no filtering or detrending of the data was performed. Potential errors due to sensor separation were minimised by removing any time lag between the thermocouple temperature and vertical velocity signal. Sonic temperatures were corrected as specified in the text. For the energy budget comparison, net radiation, soil heat flux and latent heat flux measurements were adjusted for sensor calibration. The soil heat flux was corrected for storage in the overlying soil layer following the method of Mayocchi and Bristow (1995) while latent heat fluxes were adjusted for canopy storage using within canopy profile measurements of humidity. Frequency response corrections were applied to the latent heat fluxes following the methods of Moore (1986) but were not applied to the sensible heat fluxes.

H.4 Theory

H.4.1 Definition of equation

The theory of measurement of air temperature using speed of sound as measured by a sonic anemometer have been well described (Kaimal and
Air temperature, $T$, may be measured using the speed of sound, $c$, in a gas of constant density, as in equation (1). The speed of sound may be determined from the travel times, $t_1$ and $t_2$, of sound pulses travelling in opposite directions along a path length, $L$, equation (2). The values of $t_1$ and $t_2$ are obtained from sonic anemometer measurements. If any of the variables involved in this calculation change, errors will be introduced into the sonic measurement of the speed of sound, and hence temperature.

$$ T = \frac{c^2}{403} $$

$$ c = \frac{L}{2} \left( \frac{1}{t_1} + \frac{1}{t_2} \right) $$

Following Kaimal & Gaynor (1991), and using the modified sound vector field as shown in figure 1, the sound pulse transit times may be described as:

$$ \frac{1}{t_1} = \frac{c \cdot \cos(\gamma_1) + V_d}{L_m} $$

$$ \frac{1}{t_2} = \frac{c \cdot \cos(\gamma_2) - (V_d + \Delta V_d)}{L_m} $$

where

$$ \gamma_1 = \sin^{-1}\left( \frac{V_n}{c} \right) $$

$$ \gamma_2 = \sin^{-1}\left( \frac{V_n + \Delta V_n}{c} \right) $$
These equations differ from those given by Kaimal & Gaynor (1991) by the replacement of the expected path length with the value of the path length at the time of measurement ($L_m$) and the inclusion of the change in velocity along the sound path $\Delta V_d$ and normal to the sound path $\Delta V_n$ as given in the descriptions of $t_2$ and $\gamma_2$.

In previous theoretical developments, the path length, $L$, has been assumed constant. With this value held constant, the speed of sound from equation (2) may be affected by the path length at the time of measurement ($L_m$) implicit in the transit times, equation (3). Here we have assumed that the value of $L_m$ is constant over the period required to measure the two transit times $t_1$ and $t_2$.

Previous theoretical developments have also not included the effect of flow acceleration, which appears in the equations for $t_2$ and $\gamma_2$. In still air the values of $t_1$ and $t_2$ will be identical. Similarly, for air at constant velocity, the values of $t_1$ and $t_2$ will average to their equivalent value for still air of the same temperature. However, for a parcel of air which is accelerating along the measurement path there will be an increase or decrease in the transit time $t_2$ which is not related to air temperature but is instead directly related to the change in velocity between the times of measurement of $t_1$ and $t_2$ resulting from the presence of $\Delta V_d$ and $\Delta V_n$.

Following Kaimal & Gaynor (1991) and including terms for flow acceleration over the measurement period, the speed of sound as measured by a sonic anemometer, $c_a$, may be written:

$$c_a = \frac{L}{2} \left[ \frac{c \cos(\gamma_1) + V_d}{L_m} + \frac{c \cos(\gamma_2) - (V_d + \Delta V_d)}{L_m} \right]$$

Substituting for the definition of $\gamma$ from equations (4) and combining terms gives:

$$c_a = \frac{1}{2} \frac{L}{L_m} \left[ \left( c^2 - V_n^2 \right)^{\frac{1}{2}} + \left( c^2 - V_n^2 - 2 \cdot V_n \cdot \Delta V_n - \Delta V_n^2 \right)^{\frac{1}{2}} - \Delta V_d \right]$$
Performing a binomial series expansion of the second term within the square brackets and retaining only the first two terms of that expansion but excluding terms which contain a squared velocity change, and then carrying through with the multiplication of 1/2, gives us an equation describing the speed of sound as measured by a sonic anemometer:

\[ c_a = \frac{L}{L_m} \left[ \left( c^2 - V_n^2 \right)^{\frac{3}{2}} - \frac{V_a \cdot \Delta V_n}{2 \cdot \left( c^2 - V_n^2 \right)^{\frac{3}{2}}} - \frac{\Delta V_d}{2} \right] \]  

(7)

The desired value from this equation is the speed of sound, \( c \), and the other variables must be determined in order to ascertain its value. The value of \( L \) in this equation is implicit in the manufacturers signal processing and is given in the sensors documentation, while the values of \( c_a, V_n, \Delta V_n, \) and \( \Delta V_d \) are available from the anemometer's output. The value of \( L_m \) is unknown, and will be discussed in the following section.

A typical value of the change in velocity between samples is on the order of 0.1 m s\(^{-1}\). For the anemometer used in this experiment, eight sets of six transit times are measured for each sample output, giving a change in velocity between \( t_1 \) and \( t_2 \) which is on the order of 0.021 times the change in velocity between samples. While the values of \( \Delta V_n \) are of the same magnitude as those of \( \Delta V_d \), the large denominator in the second term within the brackets suggests that this term will be about two orders of magnitude smaller than the third term, and may reasonably be ignored, giving:

\[ c_a = \frac{L}{L_m} \left[ \left( c^2 - V_n^2 \right)^{\frac{3}{2}} - \frac{\Delta V_d}{2} \right] \]  

(8)

Employing equations (1) and (8), we may define the temperature as measured by the sonic anemometer:
Expanding the numerator we eliminated the squared velocity difference term as being negligibly small. After rearranging, with the use of equation (1), we may then define the sonic temperature in terms of true temperature instead of speed of sound:

\[
T_s = \left( \frac{L}{L_m} \right)^2 \left[ \frac{T - \frac{V_n^2}{403} - \frac{\Delta V_d \cdot T_s^{1/2}}{20.07}}{403} \right]
\]

In so doing, we have approximated the value of \( T \) in the last term using the value of \( T_s \). This may be rearranged to describe the actual temperature in terms of \( T_s \).

\[
T = \left( \frac{L_m}{L} \right)^2 T_s + \frac{V_n^2}{403} + \frac{\Delta V_d \cdot T_s^{1/2}}{20.07}
\]

The final addition is to include the correction for the effect of changes in density due to constituent changes, primarily water vapour, as has been described by Schotanus et al. (1983) and Kaimal & Gaynor (1991). If uncorrected for water vapour, the resulting temperature is nearly equivalent to the virtual air temperature, and may be used in this form for the calculation of stability terms, which require the use of virtual temperature.

\[
T = \left( \frac{L_m}{L} \right)^2 T_s + \frac{V_n^2}{403} + \frac{\Delta V_d \cdot T_s^{1/2}}{20.07} - 0.00032 \cdot \bar{T} \cdot q
\]

In order to couch the equation in flux terms, we separate the variables into their mean and deviation components. The mean \( w \) and mean \( \Delta V_d \) have been excluded, as their values are assumed equal to zero. Expanding equation (12) and Reynolds averaging results in the flux equation, of which we have only retained the higher order products of the second term:
This formulation is similar to that given by Schotanus et al. (1983) and Kaimal & Gaynor (1991) with the exception of the retention of the 4th term, the addition of the 2nd and 5th terms, and the modifiers for path length differences in the first term. All the terms of equations (12) and (13) are measurable with the exception of the variables including $L_m$. Therefore, we next define the relationship of the variable $L_m$ to horizontal wind speed and direction, for the anemometer under consideration.

H.4.2 Geometry of velocity dependent path length changes

The geometry of the distortion of a sonic anemometer’s sensor probe will determine the value of $L_m$, resulting in confounding effects of sonic temperature measurement. As this effect is inherently sensor dependent, the resulting equations will apply specifically to the anemometer used in its derivation, in this analysis the Gill Solent 101R2. Some assumptions are also required to allow us to simplify the theory involved. We have assumed that any flow distortion resulting from probe design has been fully compensated by the manufacturer’s correction matrix. Grelle & Lindroth (1994) have discussed this topic and the weaknesses of this assumption will not be addressed further within this paper. More specific to this anemometer, we have assumed initially that the horizontal (aluminum) structure elements are inelastic relative to the vertical (carbon fibre) structure supports. We have assumed that the deformation of the sensor head is uniform for azimuthal changes in forcing and that there are no torsional forces acting on the sensor head. We assume the transducer support arms are inelastic and that the vertical structure support deformation acts such that the vertical supports are inelastic but the connections between the vertical supports and the horizontal supports are elastic. An alteration of this last assumption so that the vertical supports were considered elastic and deform in a sinusoidal manner showed insignificant
effects on the final results. Some values that we have assumed in these calculations include: the drag coefficient of the sensor head \((Cd = 2)\), the area of drag of the sensor head \((Area = 18 \text{ cm}^2)\), and the sonic's path length under calm conditions \((L = 15 \text{ cm})\).

Figure 2 shows the geometry describing the path between two transducers mounted at the end of support arms. The support arms are mounted with a separation, \(R\) (0.165 m). Each transducer support arm displaces its transducer, in opposing directions, from the sensors vertical axis, \(R\), by a distance \(X\) (0.0530 m) along the x-axis. Each support arm displaces its transducer vertically towards the centre of the probe, parallel to the z axis, by an amount \(Z\) (0.0295 m). Note that in figure 2, the bottom transducer is located at the origin so that the displacement of the upper transducer incorporates the displacements of both upper and lower transducer support arms. The resulting separation between the transducers, \(L_m\), is the path over which the sonic temperature is measured.

An ideal supporting structure would be rigid so that the path length \(L_m\) would be unresponsive to wind forces. Flexibility in the vertical support structure results in a displacement of the upper transducer arm by an angle \(\theta\) for wind from the direction \(\theta\). The anemometer employed in this analysis measures temperature using the transducer pair which lies along the axis defining the \(u\) wind component (the x axis in Figure 2). Therefore wind velocity and direction are calculated from the anemometer's unrotated horizontal wind components \(u\) and \(v\) as:

\[
V = \sqrt{u^2 + v^2} \tag{14}
\]

\[
\theta = \tan^{-1}\left(\frac{v}{u}\right) \tag{15}
\]

The angle \(\theta\) with which the upper transducer arm is displaced by a wind of given velocity can be estimated using a modification of the equation describing
the force applied to the anemometers upper transducer arm support structure (Welty et al., 1984):

\[ \omega = D_a \cdot \text{Area} \cdot C_d \cdot \frac{L}{2} \cdot V^2 \]  \hspace{1cm} (16)

in which \( \text{Area} \) (m\(^2\)) is area of horizontally projected area of the upper half of the sonic probe, \( C_d \) is the drag coefficient, \( \rho \) is the density of air (kg m\(^{-3}\)) and \( V \) is the wind speed (m s\(^{-1}\)). The standard equation describing the force of a fluid has then been altered by including a wind force dependent displacement angle \( D_a \) (deg s kg\(^{-1}\) m\(^{-1}\)), which determines the head displacement for a given wind force. If we determine the displacement angle associated with a given wind velocity, we may then describe the effective path length over which a corresponding sonic temperature measurement is made:

\[ L_m = \sqrt{(R \cdot \cos(\theta) \cdot \sin(\varphi) + 2 \cdot X)^2 + (R \cdot \sin(\theta) \cdot \sin(\varphi))^2 + (R \cdot \cos(\varphi) - 2 \cdot Z)^2} \]  \hspace{1cm} (17)

**H.4.3 Method of estimating the velocity dependent displacement angle**

From equation (9) defining the temperature as measured by the sonic anemometer, we know that the only missing variable required to define the effect of distortion of probe head is the measurement path length \( L_m \). Equation (17) describes \( L_m \) in relation to wind direction and wind speed, but leaves us with an alternate missing variable, the velocity dependent displacement angle, \( \omega \). Other empirical solutions to this problem exist, \( i.e. \) direct measurements of the effects of applied forces, and wind tunnel studies under controlled temperature conditions) and could be used to determine the desired displacement angle velocity relationship. However we have attempted an \textit{in situ} approach of comparing the sonic temperature with that measured by a fine wire thermocouple as it is an approach that could easily be replicated by other researchers.
This approach requires us to know the error term for the thermocouple temperature. We assume that the fine wire thermocouple temperature, $T_c$, is responding linearly to true temperature, $T$, but has an offset from true temperature, $\Delta T_c$. The relationship between true temperature and thermocouple temperature may then be defined as:

$$T = T_c + \Delta T_c$$  \hspace{1cm} (18)

Assuming that under calm conditions $L = L_m$, the sonic temperature will represent the true temperature when adjusted for humidity. Equations (12) and (18) may then be combined and rearranged to provide an estimate of $\Delta T_c$:

$$\Delta T_c = \left( \frac{T_s}{1 + 0.00032 \cdot q} \right) - T_c$$  \hspace{1cm} (19)

This estimate of $\Delta T_c$ is then used to define the path length at the time of measurement:

$$L_m = L \sqrt{\frac{(T_c + \Delta T_c - ^{\frac{V_s^2}{403}} \cdot \frac{\Delta V_a \cdot (T_c + \Delta T_c) ^{\frac{3}{2}}}{20.07} + 0.00032 \cdot (T_c + \Delta T_c) \cdot q}}{T_s}}$$  \hspace{1cm} (20)

We must assume that the thermocouple error term is reasonably constant in time and that it is not a function of either wind speed or humidity. This will not be entirely true, as thermal loading of the thermocouple will make it sensitive to fluctuations in low wind speed on sunny days. However, we evaluate $L_m$ for high wind speed conditions to minimize this problem.
H.5 Results

H.5.1 Determination of $\omega$ and $L_m$

We employed equation (20) to calculate the instantaneous values of $L_m$. These values were averaged by classes of wind direction and wind speed. For this analysis wind directions are specified relative to the path of the sonic temperature measurement, e.g. 0 degrees is for wind coming from the positive x axis and +90 degrees is for wind coming from the positive y axis in Figure 2. In Figure 3, we present these data for wind directions of 0 (+/− 10) and 180 (+/− 10) degrees. Consistent with theory, in this figure we observe a shortening of the sonic path for winds coming from the direction of the upper sonic transducer (0 degrees). A smaller lengthening of the path is observed for winds coming from the opposing direction. We would expect an equal and opposing effect on path length for winds from these opposing directions; the lack of such a response is unexplained. Nevertheless, we are still able to estimate the value of $Da$ from the difference in the responses for the two wind directions at high wind speeds. Using equation (17) and evaluating it for winds from 0 and 180 degrees relative to the sonic’s path, we can describe the differences in path lengths as:

$$L_{m2}^2 - L_{m1}^2 = -8 \cdot R \cdot X \cdot \sin(\omega)$$  (21)

From Figure 3, the difference between the 0 and 180 degree averaged path lengths at a velocity of 12 m s$^{-1}$ is approximately 0.6 mm which, because we are looking at opposing wind directions, should be twice the displacement for each of the two wind directions individually. From this displacement we calculated a value for $Da$ of 0.28 deg s kg$^{-1}$ m$^{-1}$. This value was used as an initial estimate in determining a function to accurately define values of $Da$ for use in equation (16). From Figure 3 we also inferred that the response of $Da$ to wind direction may not be uniform as assumed. We therefore reprocessed a subset of the sonic temperature data employing a range of constant values of $Da$. The heat flux error term (corrected $H_s - H_{ic}$) was grouped by wind speed and wind direction to give a rough directional response of $Da$. These results
were then fit with a sinusoidal model of $Da$ to take into account directional responses of $Da$:

$$Da = a + b \cdot \sin (c \cdot \theta + d) + e \cdot \sin (f \cdot \theta + g)$$  (22)

The model correction was applied to one month's data, and the differences between the model corrected $H_s$ and $H_{tc}$ were examined. This process was repeated numerous times and the residual error analysed with respect to wind direction until an apparently optimal form of equation (22) was determined. We observed that the sinusoidal model of $Da$ was better than any constant value of $Da$. However, wind tunnel tests may be needed to establish an ideal form and coefficients for equation (22). The resulting form for equation (22) was:

$$Da = -0.5 - 0.5 \cdot \sin ((\theta + 110) \cdot 0.8) - 0.02 \cdot \sin ((\theta + 135) \cdot 2)$$  (23)

Because these coefficients were determined for a data subset, we verified the effect of the sine function described in equation (23) by correcting the remainder of the years data. The results of this analysis are shown in following sections.

H.5.2 Proportional effects of sonic temperature correction for distortion and flow acceleration

The correction proposed in equation (12) incorporates the effects of two components. The proportional effects of those two components are described in Table 1. The data used in determining the component effects was the same as that used in determining the final correction effects. Table 1 gives the average, standard deviation, minimum and maximum values for the corrections to $H_s$ as proposed by Schotanus et. al. (1983) and as proposed in this paper. The combined effect of the individual acceleration and deformation corrections are given as well as the correction combining all terms.
The correction component values in Table 1 indicate that the corrections for acceleration and deformation are of the same order of magnitude as the standard corrections described by Schotanus et. al. (1983). For the data set to which these corrections were applied, the acceleration term had a net negative effect and the deformation had a net positive effect. Both the acceleration and deformation terms may be affected by the orientation of the sonic head with respect to the mean flow so that the average corrections stated here are not necessarily indicative of these corrections applied under different experimental conditions. It is quite plausible that for different wind conditions the net effect of these components could be negative, positive or zero. In this situation, however, the two additional correction components proposed are of opposing sign and result in a small positive adjustment in the net correction as compared to the standard correction.

H.5.3 Expected effects of revised sonic temperature correction

From equation (12) we know that sonic temperature will be less than the true temperature for higher wind velocities. Alternately, at higher humidity sonic temperature will be higher than true temperature. It has already been shown (Kaimal and Gaynor, 1991) and (Hignett, 1992) that correction for these effects typically results in a negative offset to $H_s$. Based on the theory derived above, increases in measurement path length and flow acceleration will both cause sonic temperature to be less than true temperature.

More precisely, from equation (17), we know that the sonic temperature measurement path length will increase for winds from around 0 degrees and decrease for winds from around 180 degrees. For strong correlations between $w$ and $u$, the measurement path length $L_m$ will be correlated with $w$ because of the velocity terms in equation (16). As a result, variations in $L_m$, should cause a positive heat flux error for winds from relative north and a negative error for winds from relative south. We do not have predefined expectation of the effect of the acceleration terms. It will obviously be a function of the correlation of
the characteristics of the along path flow with those of wind direction and vertical velocity.

The effect of the proposed additional corrections on sonic temperature statistics will be small relative to the effect on fluxes because the coherence between $T$ and the controlling wind velocity variables is smaller than that for $w$ and these variables. Indeed the observed effect of these corrections on sonic temperature means and standard deviations was on the order of 1% or less. Therefore the effect of the proposed correction on temperature statistics is not considered further. However, because of the autocorrelation between the variables controlling the magnitude of the proposed correction and the vertical velocity incorporated in the sensible heat flux, the potential effect of the proposed correction upon sensible heat flux is greater. In this paper we have assumed that the effect of noise on the correction terms is small in comparison to the natural correction term.

H.5.4 Observations on the effects of revised sonic temperature correction

The sample time series given in Figure 4, compares corrected and uncorrected forms of $H_s$ with $H_{tc}$ and a corresponding trace of wind speed. We observe a close relationship between $H_{tc}$ and uncorrected $H_s$. However, after the standard correction is applied, $H_s$ is much lower than $H_{tc}$. After correction for the model proposed, equation (12), there is better correspondence, at higher wind speeds, between the sensible heat flux measured with the two different temperature sensors.

In Figure 5 the difference between $H_{tc}$ and the various forms of $H_s$ are compared as a function of wind speed for eight wind direction categories. We observe that for winds from $+/-90$ to $+/-180$ (i.e. the top four panels in Figure 5) the uncorrected $H_s$ (unmarked line) exceeds $H_{tc}$ by about 10 W m$^{-2}$ for wind speeds near 5 m s$^{-1}$, as can be expected from equation (17). Similarly, for winds from 0 to $+/-90$ degrees (bottom four panels in Figure 5) and at wind
speeds near 5 m s\(^{-1}\), we observe an equivalent or slightly larger decrease of about -25 W m\(^{-2}\) in \(H_s\) as compared to \(H_{tc}\). This is roughly in agreement with our previous determination of \(L_m\) which suggested an error which was greater for a wind regime near 0 degrees.

From Figure 5 we also observe that the standard sonic temperature correction is wind direction independent. With this correction applied (open circles), \(H_s\) is more negative than the uncorrected sensible heat flux, by an approximately similar magnitude for all wind directions. The dependence of this correction on wind speed reflects the greater turbulent transport of latent heat and momentum at higher wind speeds, equation (13).

In contrast to the standard correction, we observe in Figure 5 that the model correction does exhibit a wind direction dependence, as is implied by equation (17). The proposed correction is positive for winds from +/-90 to +/-180 and negative for winds from 0 to +/-90 degrees. For all wind direction categories, except -135 to -90 degrees, the proposed correction improves the correspondence of \(H_s\) with \(H_{tc}\). We observe that the model correction does not give a perfect fit to \(H_{tc}\). This suggests that further work is needed in refining the model describing \(Da\), or determining other errors in the measurement of sonic temperature.

If we compare the results as shown in Figure 5 with those presented by Grelle and Lindroth (1996), we observe greater error in \(H_s\) at lower wind speeds. Their data indicate that the effect of sonic deformation is not notable until wind velocities on the order of 8 to 10 m s\(^{-1}\) are reached. Our analysis, however, indicates noticeable effects for wind velocities as low as 3 m s\(^{-1}\). This discrepancy likely arises from combining data when presenting their results, as merging the data from opposing wind directions would have the effect of combining errors of different signs, resulting in a small net error. Because, their data set has fewer data at higher wind speeds, it is possible that these data come from an unequal distribution of wind directions, and hence a more obvious distortion error effect.
To examine the net effect of the corrections being investigated we present the results for sensible heat fluxes from 1998 as cumulative differences ($\Sigma(H_{ic} - H_s)$) and cumulative absolute differences ($\Sigma |H_{ic} - H_s|$). These differences portray a visual representation of the accuracy and precision of the associated correction changes with time. These results are presented in Figure 6 and Figure 7, in which Figure 6a and Figure 7a illustrate results used for the development of the model $Da$, equation (23), while Figure 6b and Figure 7b contains results for the final model of $Da$ applied to the remainder of the years data. The data in these figures are plotted against consecutive runs instead of date/time in order to avoid the gaps in the time series caused by missing data.

The cumulative differences shown in Figure 6a and b provide a visual estimate of the accuracy of the various forms of $H_s$ in representing $H_{ic}$. In the development data of Figure 6a, the net accuracy indicates that uncorrected $H_s$ overestimate $H_{ic}$. However, the net accuracy is quite good for the verification data in Figure 6b (i.e. the end point of the uncorrected $H$ difference is quite close to zero). In both the development and verification data the standard correction applies a consistent negative offset, resulting in a $H_s$ that is consistently smaller than $H_{ic}$.

In Figure 6a, an example of the proposed correction with a constant value for $Da$ (-0.28 in this case) is presented to show how the correction responds to constant values. For this data set the constant value of $Da$ resulted in a net negative offset, causing the model to underestimate $H_{ic}$ even more than the standard correction. The model $Da$ applied to the development data produced a good agreement, however, when applied to the verification data, the correction using model $Da$ resulted in a $H_s$ that was more positive than the standard corrected $H_s$ but still considerably underestimated $H_{ic}$. It is possible that the choice of data used in developing the model values of $Da$ were in some way unrepresentative of the verification data. While some inaccuracy may stem from an innate inability to obtain a “true” model to define $Da$ under all circumstances, selection of a more representative data set when developing a model for $Da$ may result in a more accurate correction. We must also consider
the possibility that slight calibration inaccuracies in the thermocouple temperature sensor result in a net bias in the comparison of accuracy by this method.

The cumulative absolute differences between $H_{tc}$ and $H_s$ shown in Figure 7a and b, provide a visual comparison of the precision of the various forms of $H_s$ in representing $H_{tc}$. For the development data, Figure 7a, the effect of the standard correction was ambiguous. The standard correction exhibited periods of improved and worsened precision of the $H_s$ in comparison with $H_{tc}$. In the verification data, Figure 7b, the effect is more consistent in that the standard correction exhibits consistently lower precision, (i.e. greater cumulative absolute error), as compared to the uncorrected $H_s$. For the development data we again note that the modelled $Da$ values show considerable improvement over the use of a constant value of $Da$. When applied to the verification data the modelled $Da$ values exhibit an improvement in the precision as compared to the standard correction. The improvement in precision of the proposed correction is still less than the precision of the uncorrected $H_s$. Further refinement of the model describing $Da$ could improve upon the precision with which the corrected $H_s$ represents a true sensible heat flux.

When the corrected $H_s$ is incorporated into an energy budget comparison, Figure 8, we observe a 2% increase in the $H+LE$ term as a result of incorporation of the proposed correction. This increase is consistent with the proposed correction being 2.5 W m$^2$ greater than standard corrections, as shown in Table 1.

**H.6 Conclusions**

Fluxes of sensible heat determined from measurements of sonic anemometer speed of sound derived temperature are known to have errors caused by cross wind velocity and humidity fluctuations. The corrections for these errors are well known and are applied as standard practise to sonic temperature derived sensible heat fluxes Aubinet *et al.* (2000). The observation by other researchers
(Grelle and Lindroth, 1996) that an additional error exists as a result of sonic probe deformation is supported by the observations of the authors. We have developed a theory to further correct sonic temperature measurements based on the geometry of the anemometer employed in the analysis. In deriving this theory we have also incorporated a term approximating the effect of flow acceleration. Because it depends only on the determination of the acceleration of velocity signals along the path of sonic temperature measurement, the correction component for acceleration may be easily incorporated into the correction of sonic determined air temperature measurement. The correction for probe distortion, on the other hand, requires accurate supporting measurements of air temperature and empirical determination of a deformation coefficient ($Da$) relevant to the anemometer being corrected. While a procedure is described for the anemometer used in this experiment, the resulting theory may not apply directly to other models of sonic anemometer. We have estimated corrections for probe distortion and flow acceleration that are of the same order of magnitude as the standard corrections of Schotanus et al. (1983). The distortion component of this correction had an average effect of $+7.5 \text{ W m}^{-2}$ and the acceleration component had an average effect of $-4.8 \text{ W m}^{-2}$. The corresponding net effect of the proposed corrections increase the average flux by approximately $2.5 \text{ W m}^{-2}$, which corresponds to an improvement in energy budget closure of $2\%$ for the data presented. As the proposed correction is both wind velocity and wind direction dependent the observed improvement can not be assumed for application of this correction to other experimental data. The effect of this correction will be determined by the anemometer's construction, sampling frequency, and site wind characteristics. It may be beneficial to further determine the extent to which this deformation effect is applicable to and consistent with other sonic probes.
Table 1. Values of average, standard deviation, maximum, and minimum of the standard correction, components of the proposed correction model, and the proposed correction model.

<table>
<thead>
<tr>
<th>Method</th>
<th>Average</th>
<th>Std dev</th>
<th>Minimum</th>
<th>Maximum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Standard</td>
<td>-7.3</td>
<td>13.9</td>
<td>-141.0</td>
<td>109.5</td>
</tr>
<tr>
<td>Model acceleration</td>
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<td>15.6</td>
<td>-190.0</td>
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<tr>
<td>Model deformation</td>
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<td>19.2</td>
<td>-85.0</td>
<td>200.2</td>
</tr>
<tr>
<td>Model (standard + acc + def)</td>
<td>-4.6</td>
<td>13.7</td>
<td>-141.5</td>
<td>109.5</td>
</tr>
</tbody>
</table>
Figure 1. Schematic representation of sonic anemometer sound path vector travel. Schematic representation including effects of cross-path velocity component and asynchronous sampling of sound path travel times.

Figure 2. Schematic diagram demonstrating the sonic anemometer head deformation geometry. The heavy line ($L_m$) represents the sonic sensing path with the heavy dots at each end representing a sonic transducer. Note that the sonic transducer positions have been shifted so that the bottom transducer is located at the origin. The relevant deformation angle is $\omega$ and $\theta$ is the wind direction.

Figure 3. Averaged instantaneous estimations of measurement path length grouped by wind velocities from 0 to 12 m s$^{-1}$, for two opposing wind directions.

Figure 4. Sample time trace exemplifying the differences between sensible heat fluxes measured with sonic and thermocouple temperature. Standard and model corrections to sonic sensible heat flux are also shown, as is the corresponding horizontal wind speed.

Figure 5. Sensible heat flux difference (sonic – thermocouple) vs wind speed for different wind directions relative to the path of the sonic temperature.

Figure 6. The cumulative difference between forms of sonic sensible heat flux and thermocouple sensible heat flux: a) results for model development, b) results for the model applied to the remainder of the years data.

Figure 7. The cumulative absolute difference between forms of sonic sensible heat flux and thermocouple sensible heat flux. a) results for model development, b) results for the model applied to the remainder of the years data.
Figure 8. Energy budgets comparison using sensible heat flux determined from sonic temperatures, with standard (dark circles) and model corrections (open circles) applied.
Figure 1

\[ (V_d + \Delta V_d) t_2 \]

\[ (V + \Delta V) t_2 \]

\[ \Delta V_n \]

\[ V_n \]

\[ V_d \]

\[ \Delta V_d \]

\[ L_m \]
Figure 2
Figure 3

Measurement path, $L_m$ (m)

Instantaneous wind velocity (m s$^{-1}$)
Figure 4
Figure 5
Figure 6
Figure 7
(LE + H model) = 0.6565 (Rn-G) - 15.794
$R^2 = 0.9133$

(LE + H standard) = 0.6705 (Rn-G) - 14.234
$R^2 = 0.912$

Figure 8