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Marc Aubinet
Timo Vesala
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Editors

Eddy Covariance

A Practical Guide
to Measurement and Data Analysis

 Springer

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A Practical Guide to Measurement and Data
Analysis

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Editors

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Preface

As soon as the first eddy covariance networks developed, in the mid-1990s, the need for standardisation became clear. Standardisation concerned not only material but also data treatment, corrections, computation. In order to harmonise these procedures in the frame of the EUROFLUX network, some software intercomparison exercises were developed: “golden files” were circulated between teams and treated with different software packages with the aim to compare the computation results. It rapidly appeared that beyond some bugs that appeared in new software and were corrected immediately, important differences remained between different computations that were due to the use of different hypotheses. The necessity to clarify these choices, and to propose a standardised (even if perfectible) eddy covariance flux computation procedure, led us to publish a first methodological paper (Aubinet et al. 2000). Eleven years later, this paper remains an often cited reference in the field.

However, as the theory and measurement techniques progressed since, and since the eddy covariance techniques is becoming also a monitoring exercise and not more only a purely scientific activity, the necessity of an update of this paper and of creating something that could help to install an eddy covariance site and manage it correctly grew. In December 2008, during a meeting at the Hyytälä Forestry Field Station (Finland) celebrating the tenth anniversary of the EUROFLUX network constitution, the idea was launched (originally by Samuli Launiainen) to produce such an update. However, it appeared rapidly that if we wanted to produce a self-standing document, useful to eddy flux practisiers, we could not limit its size to those of a paper.

We thus decided to tackle the edition of a book with the general objective to give to eddy flux practisiers the theoretical and practical information necessary in order to develop eddy covariance measurements, from site installation to data treatment. After preparing a book plan, structured in 17 chapters, we chose different first authors, known for their skills in the field and asked them to constitute a team of co-authors and prepare their chapters. The present book is the result of the two and half year long work that followed.

After a first chapter recalling the theoretical bases on which eddy covariance method relies, Chap. 2 describes technical requirements of the eddy covariance set-up: tower positioning and dimensioning (height, position, system positioning on the tower), sonic and gas analysers, dimensioning, calibration and maintenance.

Chapter 3 describes the general procedure used in order to get “uncorrected” fluxes and to discuss the pros and cons of different computation alternatives. This implied especially a description of the data acquisition set-ups, and a detailed discussion on flux computation (fluctuation computation, first quality control on raw data, time lagging, rotation and flux computation).

Chapter 4 concentrates on the different corrections procedures necessary in order to get good quality fluxes and on the quality tests on these fluxes.

Chapter 5 focuses on the problem of night flux underestimation, its causes and its impact on flux measurements. It described different screening or correction procedures and discussed their pros and cons.

Chapter 6 specifies the conditions when data gap filling is necessary and which precautions should be taken when performing data gap filling. It presented and compared the different data gap filling procedures and their (dis)advantages.

Chapter 7 identifies and quantifies the different causes of uncertainty in flux measurements and analyses how they combine during scaling up.

Chapter 8 describes the main footprint models and the way they could be combined with vegetation cover maps (in order to identify the sources/sinks of flux) or with quality tests (in order to evaluate the general quality of data).

Chapter 9 presents the different possibilities to partition eddy flux into ecosystem respiration and gross ecosystem photosynthesis. Different approaches based on night-time or on day-time data were described.

Chapter 10 focuses on disjunct eddy covariance technique, which is especially adapted to capture tracer gas.

Chapters 11–16 describe the specific requirements for flux measurements in specific ecosystems like forests, grasslands, croplands, wetlands, lakes or urban environment.

Finally, Chap. 17 describes the objectives of a data base, the way it should be maintained and managed. In addition, it proposes some policies for data use, exchange and publication.

The editors would like to thank the co-authors of the chapters for their enthusiasm and their involvement in this long (but, hopefully, useful) work that we hope can contribute to reinforce the links between the different eddy covariance networks in the world.

Dario Papale and Timo Vesala, although editors of this book, would like to thank M. Aubinet for taking care of the lion’s share of the editing job.

The book idea and preparation has been also supported by the IMECC EU project and the ABBA Cost Action.

This book is dedicated to all the field (often anonymous) technicians whose continuous system care, maintenance and follow up constitute an inestimable contribution to ecosystem studies and to the Ph.D. students that decide to base their work on these unique measurements.

Folks, mark already in your calendars “the 20th Anniversary of EUROFLUX” to be held around 10 December in 2018, once again in Hyttiälä. We do not know yet what will be the main product of the meeting then.

Marc Aubinet
Dario Papale
Timo Vesala

Contents

1	The Eddy Covariance Method	1
	Thomas Foken, Marc Aubinet, and Ray Leuning	
1.1	History	1
1.2	Preliminaries	2
	1.2.1 Context of Eddy Covariance Measurements	2
	1.2.2 Reynolds Decomposition	4
	1.2.3 Scalar Definition	5
1.3	One Point Conservation Equations	6
	1.3.1 Dry Air Mass Conservation (Continuity) Equation	6
	1.3.2 Momentum Conservation Equation	7
	1.3.3 Scalar Conservation Equation	8
	1.3.4 Enthalpy Equation	9
1.4	Integrated Relations	9
	1.4.1 Dry Air Budget Equation	10
	1.4.2 Scalar Budget Equation (Generalized Eddy Covariance Method)	10
1.5	Spectral Analysis	12
	1.5.1 Spectral Analysis of Turbulence	13
	1.5.2 Spectral Analysis of Atmospheric Turbulence	13
	1.5.3 Sensor Filtering	14
	1.5.4 Impacts of Measurement Height and Wind Velocity	15
	References	16
2	Measurement, Tower, and Site Design Considerations	21
	J. William Munger, Henry W. Loescher, and Hongyan Luo	
2.1	Introduction	21
2.2	Tower Considerations	22
	2.2.1 Theoretical Considerations for Tower Design	22
	2.2.1.1 Diverse Ecosystems and Environments	22
	2.2.1.2 Physical Effects on Surrounding Flows Due to the Presence of Tower Structure	22

2.2.1.3	Size of Horizontal Supporting Boom	26
2.2.1.4	Tower Deflection and Oscillations	27
2.2.1.5	Recirculation Zone at the Opening in a Tall Canopy	27
2.2.2	Tower Design and Science Requirements	28
2.2.2.1	Tower Location Requirements	28
2.2.2.2	Tower Structure Requirements	30
2.2.2.3	Tower Height Requirements	31
2.2.2.4	Tower Size Requirements	32
2.2.2.5	Instrument Orientation Requirements	33
2.2.2.6	Tower Installation and Site Impact Requirements	34
2.3	Sonic Anemometer	35
2.3.1	General Principles	35
2.3.2	Problems and Corrections	36
2.3.3	Requirements for Sonic Choice, Positioning, and Use	37
2.4	Eddy CO ₂ /H ₂ O Analyzer	40
2.4.1	General Description	40
2.4.2	Closed-Path System	41
2.4.2.1	Absolute and Differential Mode	41
2.4.2.2	Tubing Requirements for Closed-Path Sensors	42
2.4.2.3	Calibration for CO ₂	46
2.4.2.4	Water Vapor Calibration	47
2.4.3	Open-Path Systems	47
2.4.3.1	Installation and Maintenance	47
2.4.3.2	Calibration	48
2.4.4	Open and Closed Path Advantages and Disadvantages ...	48
2.4.5	Narrow-Band Spectroscopic CO ₂ Sensors	50
2.5	Profile Measurement	51
2.5.1	Requirements for Measurement Levels	53
2.5.2	Requirements for Profile Mixing Ratio Measurement ...	54
	References	54
3	Data Acquisition and Flux Calculations	59
	Corinna Rebmann, Olaf Kolle, Bernard Heinesch, Ronald Queck, Andreas Ibrom, and Marc Aubinet	
3.1	Data Transfer and Acquisition	60
3.2	Flux Calculation from Raw Data	65
3.2.1	Signal Transformation in Meteorological Units	66
3.2.1.1	Wind Components and Speed of Sound from the Sonic Anemometer	66
3.2.1.2	Concentration from a Gas Analyzer	67
3.2.2	Quality Control of Raw Data	67

3.2.3	Variance and Covariance Computation	71
3.2.3.1	Mean and Fluctuation Computations	71
3.2.3.2	Time Lag Determination	72
3.2.4	Coordinate Rotation	73
3.2.4.1	Requirements for the Choice of the Coordinate Frame and Its Orientation	73
3.2.4.2	Coordinate Transformation Equations	75
3.2.4.3	Determination of Rotation Angles	76
3.3	Flux Determination	79
3.3.1	Momentum Flux	79
3.3.2	Buoyancy Flux and Sensible Heat Flux	80
3.3.3	Latent Heat Flux and Other Trace Gas Fluxes	80
3.3.4	Derivation of Additional Parameters	80
	References	82
4	Corrections and Data Quality Control	85
	Thomas Foken, Ray Leuning, Steven R. Oncley, Matthias Mauder, and Marc Aubinet	
4.1	Flux Data Correction	86
4.1.1	Corrections Already Included into the Raw Data Analysis (Chap. 3)	86
4.1.2	Conversion of Buoyancy Flux to Sensible Heat Flux (SND-correction)	86
4.1.3	Spectral Corrections	87
4.1.3.1	Introduction	87
4.1.3.2	High-Frequency Loss Corrections	88
4.1.3.3	Low-Cut Frequency	96
4.1.4	WPL Corrections	97
4.1.4.1	Introduction	97
4.1.4.2	Open-Path Systems	97
4.1.4.3	WPL and Imperfect Instrumentation	99
4.1.4.4	Closed-Path Systems	99
4.1.5	Sensor-Specific Corrections	101
4.1.5.1	Flow Distortion Correction of Sonic Anemometers	101
4.1.5.2	Correction Due to Sensor Head Heating of the Open-Path Gas Analyzer LiCor 7500	103
4.1.5.3	Corrections to the Krypton Hygrometer KH20	103
4.1.5.4	Corrections for CH ₄ and N ₂ O Analyzers	104
4.1.6	Nonrecommended Corrections	105
4.1.7	Overall Data Corrections	106
4.2	Effect of the Unclosed Energy Balance	108
4.2.1	Reasons for the Unclosed Energy Balance	108

4.2.2	Correction of the Unclosed Energy Balance	111
4.3	Data Quality Analysis	112
4.3.1	Quality Control of Eddy Covariance Measurements	113
4.3.2	Tests on Fulfilment of Theoretical Requirements	114
4.3.2.1	Steady State Tests	115
4.3.2.2	Test on Developed Turbulent Conditions	116
4.3.3	Overall Quality Flag System	117
4.4	Accuracy of Turbulent Fluxes After Correction and Quality Control	119
4.5	Overview of Available Correction Software	125
	References	125
5	Nighttime Flux Correction	133
	Marc Aubinet, Christian Feigenwinter, Bernard Heinesch, Quentin Laffineur, Dario Papale, Markus Reichstein, Janne Rinne, and Eva Van Gorsel	
5.1	Introduction	133
5.1.1	History	133
5.1.2	Signs Substantiating the Night Flux Error	134
5.1.2.1	Comparison with Bottom Up Approaches	134
5.1.2.2	Sensitivity of Flux to Friction Velocity	134
5.1.3	The Causes of the Problem	135
5.2	Is This Problem Really Important?	136
5.2.1	In Which Case Should the Night Flux Error Be Corrected?	137
5.2.2	What Is the Role of Storage in This Error?	137
5.2.3	What Is the Impact of Night Flux Error on Long-Term Carbon Sequestration Estimates?	138
5.2.4	What Is the Impact of the Night Flux Error on Functional Relationships?	139
5.2.5	What Is the Impact of the Night Flux Error on Other Fluxes?	139
5.3	How to Implement the Filtering Procedure?	143
5.3.1	General Principle	143
5.3.2	Choice of the Selection Criterion	145
5.3.3	Filtering Implementation	145
5.3.4	Evaluation	147
5.4	Correction Procedures	148
5.4.1	Filtering + Gap Filling	148
5.4.2	The ACMB Procedure	149
5.4.2.1	History	149
5.4.2.2	Procedure	150
5.4.2.3	Evaluation	151
	References	152

6	Data Gap Filling	159
	Dario Papale	
6.1	Introduction	159
6.2	Gap Filling: Why and When Is It Needed?	160
6.3	Gap-Filling Methods	160
6.3.1	Meteorological Data Gap Filling	162
6.3.2	General Rules and Strategies (Long Gaps)	163
6.3.2.1	Sites with Management and Disturbances	164
6.3.3	Methods Description	165
6.3.3.1	Mean Diurnal Variation	165
6.3.3.2	Look-Up Tables	165
6.3.3.3	Artificial Neural Networks	167
6.3.3.4	Nonlinear Regressions	168
6.3.3.5	Process Models	168
6.4	Uncertainty and Quality Flags	169
6.5	Final Remarks	170
	References	171
7	Uncertainty Quantification	173
	Andrew D. Richardson, Marc Aubinet, Alan G. Barr, David Y. Hollinger, Andreas Ibrom, Gitta Lasslop, and Markus Reichstein	
7.1	Introduction	173
7.1.1	Definitions	175
7.1.2	Types of Errors	175
7.1.3	Characterizing Uncertainty	177
7.1.4	Objectives	177
7.2	Random Errors in Flux Measurements	178
7.2.1	Turbulence Sampling Error	179
7.2.2	Instrument Errors	179
7.2.3	Footprint Variability	180
7.2.4	Quantifying the Total Random Uncertainty	180
7.2.5	Overall Patterns of the Random Uncertainty	182
7.2.6	Random Uncertainties at Longer Time Scales	187
7.3	Systematic Errors in Flux Measurements	188
7.3.1	Systematic Errors Resulting from Unmet Assumptions and Methodological Challenges	188
7.3.2	Systematic Errors Resulting from Instrument Calibration and Design	190
7.3.2.1	Calibration Uncertainties	190
7.3.2.2	Spikes	194
7.3.2.3	Sonic Anemometer Errors	194
7.3.2.4	Infrared Gas Analyzer Errors	194
7.3.2.5	High-Frequency Losses	195
7.3.2.6	Density Fluctuations	195

7.3.2.7	Instrument Surface Heat Exchange	197
7.3.3	Systematic Errors Associated with Data Processing	197
7.3.3.1	Detrending and High-Pass Filtering	198
7.3.3.2	Coordinate Rotation	201
7.3.3.3	Gap Filling	201
7.3.3.4	Flux Partitioning	202
7.4	Closing Ecosystem Carbon Budgets	203
7.5	Conclusion	203
	References	204
8	Footprint Analysis	211
	Üllar Rannik, Andrey Sogachev, Thomas Foken, Mathias Göckede, Natascha Kljun, Monique Y. Leclerc, and Timo Vesala	
8.1	Concept of Footprint	211
8.2	Footprint Models for Atmospheric Boundary Layer	214
8.2.1	Analytical Footprint Models	214
8.2.2	Lagrangian Stochastic Approach	216
8.2.3	Forward and Backward Approach by LS Models	217
8.2.4	Footprints for Atmospheric Boundary Layer	219
8.2.5	Large-Eddy Simulations for ABL	223
8.3	Footprint Models for High Vegetation	224
8.3.1	Footprints for Forest Canopy	224
8.3.2	Footprint Dependence on Sensor and Source Heights	226
8.3.3	Influence of Higher-Order Moments	227
8.4	Complicated Landscapes and Inhomogeneous Canopies	229
8.4.1	Closure Model Approach	229
8.4.2	Model Validation	231
8.4.3	Footprint Estimation by Closure Models	233
8.4.4	Footprints over Complex Terrain	237
8.4.5	Modeling over Urban Areas	241
8.5	Quality Assessment Using Footprint Models	243
8.5.1	Quality Assessment Methodology	244
8.5.2	Site Evaluation with Analytical and LS Footprint Models	249
8.5.3	Applicability and Limitations	250
8.6	Validation of Footprint Models	252
	References	253
9	Partitioning of Net Fluxes	263
	Markus Reichstein, Paul C. Stoy, Ankur R. Desai, Gitta Lasslop, and Andrew D. Richardson	
9.1	Motivation	263
9.2	Definitions	264
9.3	Standard Methods	266
9.3.1	Overview	266

9.3.2	Nighttime Data-Based Methods	266
9.3.2.1	Model Formulation: Temperature – Measurements	269
9.3.2.2	R_{eco} Model Formulation	269
9.3.2.3	Challenges: Additional Drivers of Respiration	270
9.3.2.4	Challenges: Photosynthesis – Respiration Coupling and Within-Ecosystem Transport	271
9.3.3	Daytime Data-Based Methods	273
9.3.3.1	Model Formulation: The NEE Light Response	273
9.3.3.2	Challenges: Additional Drivers and the FLUXNET Database Approach	275
9.3.3.3	Unresolved Issues and Future Work	277
9.4	Additional Considerations and New Approaches	278
9.4.1	Oscillatory Patterns	278
9.4.2	Model Parameterization	278
9.4.3	Flux Partitioning Using High-Frequency Data	279
9.4.4	Flux Partitioning Using Stable Isotopes	279
9.4.5	Chamber-Based Approaches	281
9.4.6	Partitioning Water Vapor Fluxes	281
9.5	Recommendations	282
	References	283
10	Disjunct Eddy Covariance Method	291
	Janne Rinne and Christof Ammann	
10.1	Introduction	291
10.2	Theory	291
10.2.1	Sample Interval	292
10.2.2	Response Time	292
10.2.3	Definition of DEC	293
10.3	Practical Applications of DEC	294
10.3.1	DEC by Grab Sampling	294
10.3.2	DEC by Mass Scanning	297
10.3.3	Use of DEC to Reduce the Burden on Data Transfer and Storage	300
10.4	DEC in Spectral Space	300
10.5	Uncertainty Due to DEC	303
10.6	On the History of the DEC Approach	305
	References	306
11	Eddy Covariance Measurements over Forests	309
	Bernard Longdoz and André Granier	
11.1	Introduction	309
11.2	Flux Computation, Selection, and Dependence	310
11.2.1	Correction for High Frequency Losses	310

11.2.2	Rotation Method	310
11.2.3	Friction Velocity Threshold	311
11.2.4	Selection Based on Footprint	311
11.3	Additional Measurements	311
11.3.1	Vertical Profile of Concentration in Canopy Air	312
11.3.2	Leaf Area Index	312
11.3.3	Biomass Estimates	313
11.3.4	Sap Flow	315
11.3.5	Extractable Soil Water, Throughfall, and Stem Flow	315
11.3.6	Heat Storage	316
11.4	Impact of Ecosystem Management and Manipulation	317
	References	317
12	Eddy Covariance Measurements over Crops	319
	Christine Moureaux, Eric Ceschia, Nicola Arriga, Pierre Béziat, Werner Eugster, Werner L. Kutsch, and Elizabeth Pattey	
12.1	Introduction	319
12.2	Measurement System	322
12.2.1	Choice of the Site and Communication with the Farmer	322
12.2.2	Flux Tower and Meteorological Station Configuration ...	323
12.2.3	Measurement Height	324
12.2.4	Maintenance	325
12.3	Flux Calculation	326
12.4	Flux Corrections	326
12.4.1	Storage Term	326
12.4.2	Nighttime Flux Data Screening	327
12.5	Data Gap Filling and Footprint Evaluation	327
12.6	Cumulated Carbon Exchange	327
12.7	Additional Measurements	328
12.8	Future Experimentations	329
	References	330
13	Eddy Covariance Measurements over Grasslands	333
	Georg Wohlfahrt, Katja Klumpp, and Jean-François Soussana	
13.1	Historic Overview of Grassland Eddy Covariance Flux Measurements	333
13.2	Peculiarities of Eddy Covariance Flux Measurements over Grasslands	334
13.3	Estimating Grassland Carbon Sequestration from Flux Measurements	337
13.4	Additional Measurements	339
13.5	Other Greenhouse Gases	340
	References	341

14 Eddy Covariance Measurements over Wetlands 345
 Tuomas Laurila, Mika Aurela, and Juha-Pekka Tuovinen

14.1 Introduction 345

14.2 Historic Overview 346

14.3 Ecosystem-Specific Considerations 352

14.4 Complementary Measurements 354

14.5 EC Measurements in the Wintertime 356

14.6 Carbon Balances and Climate Effects 358

14.7 Concluding Remarks 360

References 360

15 Eddy Covariance Measurements over Lakes 365
 Timo Vesala, Werner Eugster, and Anne Ojala

15.1 Introduction 365

15.2 Existing Studies 367

15.3 Surface-Specific Siting Problems 368

15.3.1 Stratification of Lakes 369

15.3.2 Aqueous Chemistry of CO₂ 369

15.3.3 Land-Lake Interactions 370

15.3.4 Quality Control Procedures 372

15.3.5 Mounting Instruments 373

References 374

16 Eddy Covariance Measurements Over Urban Areas 377
 Christian Feigenwinter, Roland Vogt, and Andreas Christen

16.1 Introduction 377

16.1.1 Scales in Urban Climatology 378

16.1.2 The Urban Atmosphere 379

16.1.3 Exchange Processes in the Urban Atmosphere 380

16.1.4 Characterization of the Urban Surface–
 Atmosphere Interface 381

16.2 Conceptual Framework for Urban EC Measurements 382

16.2.1 Turbulence Characteristics 384

16.2.2 The Volume Balance Approach 384

16.2.2.1 Turbulent Heat Fluxes in the
 Context of Urban Energy Balance Studies 385

16.2.2.2 Evapotranspiration in the Context
 of Urban Water Balance Studies 386

16.2.2.3 CO₂ Fluxes in the Context of
 Urban Metabolism Studies 386

16.2.3 Other Trace Gases and Aerosols 387

16.3 Challenges in the Siting of Urban EC Stations 388

16.4 Implications of the Peculiarities of the Urban
 Boundary Layer on EC Measurements 389

16.4.1 Advection and Storage 389

16.4.2 Flow Distortion 391

- 16.4.3 Night Flux Problem, Gap Filling, and QC/QA 393
- 16.4.4 Service and Maintenance of Instruments 393
- 16.5 Summary and Conclusions 394
- References 395
- 17 Database Maintenance, Data Sharing Policy, Collaboration 399**
 - Dario Papale, Deborah A. Agarwal, Dennis Baldocchi,
Robert B. Cook, Joshua B. Fisher, and Catharine van Ingen
 - 17.1 Data Management 400
 - 17.1.1 Functions 401
 - 17.1.2 Flux Tower Repositories 403
 - 17.1.3 Regional Repositories 404
 - 17.1.3.1 One Example: The European Eddy
Covariance Flux Database System 404
 - 17.1.4 The FLUXNET Initiative and Database 406
 - 17.2 Data Practices 408
 - 17.2.1 Contributing Data and Reporting Protocols 408
 - 17.2.2 Common Naming/Units/Reporting/Versioning 409
 - 17.2.2.1 Enabling Cross-site Analysis: Site
Identifier, Variables, and Units 409
 - 17.2.2.2 Data Releases 410
 - 17.2.2.3 File Naming 411
 - 17.2.3 Ancillary Data Collection 412
 - 17.3 Data User Services 413
 - 17.3.1 Data Products: The Example of fluxdata.org 413
 - 17.3.1.1 Users and Use Cases 413
 - 17.3.1.2 The Public Access Area 415
 - 17.3.1.3 The Authorized User Support Area 415
 - 17.3.1.4 Measurement Site Scientist Support Functions 417
 - 17.4 Data Sharing and Policy of Uses 417
 - 17.4.1 Data Sharing Motivation 417
 - 17.4.2 Data Policy of Use 419
 - 17.4.3 Additional Credit Possibilities 422
 - References 423
- Symbol Index 425**
- Abbreviations and Acronyms 431**
- Index 433**

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Chapter 1

The Eddy Covariance Method

Thomas Foken, Marc Aubinet, and Ray Leuning

1.1 History

The eddy covariance method for measuring exchanges of heat, mass, and momentum between a flat, horizontally homogeneous surface and the overlying atmosphere was proposed by Montgomery (1948), Swinbank (1951), and Obukhov (1951). Under these conditions, net transport between the surface and atmosphere is one-dimensional and the vertical flux density can be calculated by the covariance between turbulent fluctuations of the vertical wind and the quantity of interest.

Instrumentation limitations hampered early implementation of this approach. In 1949, Konstantinonov (Obukhov 1951) developed a wind vane with two hot wire anemometers to measure the shear stress but the full potential of the eddy covariance method only emerged after the development of sonic anemometers, for which the basic equations were given by Schotland (1955). After the development of the first sonic thermometer (Barrett and Suomi 1949), a vertical sonic anemometer with a 1 m path length (Suomi 1957) was used during the O'Neill experiment in 1953 (Lettau and Davidson 1957). The design of today's anemometers was developed by Bovscheverov and Voronov (1960) and later by Kaimal and Businger (1963) and

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Mitsuta (1966). These phase shift anemometers have now been replaced by running time anemometers with delay time measurements (Hanafusa et al. 1982; Coppin and Taylor 1983).

Early micrometeorological experiments from the 1950s to 1970s were designed to study fundamental aspects of atmospheric turbulence over homogeneous surfaces, whereas studies in the 1980s investigated the turbulent fluxes of momentum, sensible, and latent heat over heterogeneous surfaces. Similar experiments were conducted in the United States (FIFE, Sellers et al. 1988), in France (HAPEX, André et al. 1990), and in Russia (KUREX, Tsvang et al. 1991). These experiments were to become the basis of many further micrometeorological experiments (Foken 2008) that needed researchers who were highly experienced in micrometeorology and sensor handling.

The possibility of continuous eddy flux measurements arose in the 1990s with the development of a new generation of sonic anemometers (see reviews by Zhang et al. 1986; Foken and Oncley 1995) and infrared gas analyzers for water vapor and carbon dioxide, together with the first comprehensive software packages for the eddy covariance method (McMillen 1988). In the early 1990s, the eddy covariance method became more and more widely used by the ecological community for the measurement of the carbon dioxide and water exchange between an ecosystem and the atmosphere. The first measuring towers of what later became the international FLUXNET network (Baldocchi et al. 2001) were installed, and introductions into techniques new for nonmicrometeorologists were written (Aubinet et al. 2000; Moncrieff et al. 1997a, b). In parallel, the development of new analyzer types allowed an extension of the investigated trace gas spectrum. In particular, Tunable Diode Laser and Quantum Cascade Laser spectrometers were used for the measurement of methane and nitrous oxide (Smith et al. 1994; Laville et al. 1999; Hargreaves et al. 2001; Kroon et al. 2010), Proton Transfer Reaction Mass Spectrometers for volatile organic compounds (Karl et al. 2002; Spirig et al. 2005), and Chemiluminescent sensors for Ozone (Güsten and Heinrich 1996; Gerosa et al. 2003; Lamaud et al. 1994, a.o.).

Some milestones in the development of the eddy covariance method are given in Table 1.1 with the reference to the Chapters of this book.

1.2 Preliminaries

1.2.1 Context of Eddy Covariance Measurements

Eddy covariance measurements are typically made in the surface boundary layer, which is approximately 20–50 m high in the case of unstable stratification and a few tens of meters in stable stratification (see Stull 1988; Garratt 1992; Foken 2008; for complete definitions of layers in the atmosphere). Fluxes are approximately constant with height in the surface layer; hence measurements taken in this layer

Table 1.1 History of the development of the eddy covariance method

Historical milestone	References	See chapter/ section
Theoretical basis of the eddy covariance method	Montgomery (1948), Swinbank (1951), Obukhov (1951)	Section 1.2
Three-dimensional sonic anemometer	Bovscheverov and Voronov (1960), Kaimal and Businger (1963), Mitsuta (1966)	Chapter 2
Instrumental requirements	McBean (1972)	Chapter 2
Gas analyzer for water vapor (UV)	Buck (1973), Kretschmer and Karpovitsch (1973), Martini et al. (1973)	
Gas analyzer for water vapor (IR)	Elagina (1962), Hyson and Hicks (1975), Raupach (1978)	Chapter 2
Correction of the effect of the air density	Webb et al. (1980)	Section 4.1
Gas analyzer for carbon dioxide (IR)	Ohtaki and Matsui (1982), Elagina and Lazarev (1984)	Chapter 2
Transformation of buoyancy flux into sensible heat flux	Schotanus et al. (1983)	Section 4.1
System of transfer functions for spectral correction	Moore (1986)	Section 4.1
Fetch conditions	Gash (1986)	Chapter 8
Real-time data processing software	McMillen (1988)	Chapter 3
Source regions for fluxes (footprint), based on Gash (1986)	Schmid and Oke (1990), Schuepp et al. (1990)	Chapter 8
Relaxed eddy accumulation method, based on Desjardins (1977)	Businger and Oncley (1990)	
Influence of tubing of closed path sensors	Leuning and Moncrieff (1990)	Section 4.1.3 Chapter 3
Theoretical basis for flux footprints and sampling strategies	Horst and Weil (1994), Lenschow et al. (1994)	Chapter 8
Addressing the problem of the unclosed energy balance at the surface	Foken and Oncley (1995)	Section 4.2
Quality tests for eddy covariance data	Foken and Wichura (1996), Vickers and Mahrt (1997)	Section 4.3
Addressing the problem of vertical advection	Lee (1998) and many others	Section 1.3, Chapter 5
Methodology for FLUXNET network (EuroFlux)	Aubinet et al. (2000)	All chapters
Gap filling in the FLUXNET network	Falge et al. (2001a, b)	Chapter 6
Organization of an international network (FLUXNET)	Baldocchi et al. (2001)	All chapters

Foken et al. (1995), Foken (2008), Moncrieff (2004), modified

are representative of the fluxes from the underlying surfaces which are desired to be known. Here atmospheric turbulence is the dominant transport mechanism, justifying the use of the eddy covariance approach to measure the fluxes.

Some preliminary definitions are necessary before discussing the eddy covariance approach in detail.

1.2.2 Reynolds Decomposition

The description of turbulent motions in the following theory sections requires the decomposition of the time-series of each variable ζ into a time-mean part, $\bar{\zeta}$, and a fluctuating part, ζ' , the so-called Reynolds decomposition (Fig. 1.1). This can be written as:

$$\zeta = \bar{\zeta} + \zeta' \quad (1.1a)$$

where:

$$\bar{\zeta} = \frac{1}{T} \int_t^{t+T} \zeta(t) dt \quad (1.1b)$$

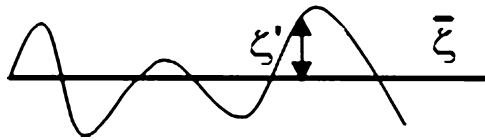
The application of Reynolds decomposition requires some averaging rules for the turbulent value ζ' which are termed Reynolds postulates:

$$\begin{aligned} I \quad & \overline{\zeta'} = 0 \\ II \quad & \overline{\zeta\xi} = \bar{\zeta}\bar{\xi} + \overline{\zeta'\xi'} \\ III \quad & \overline{\zeta\xi\xi} = \bar{\zeta}\bar{\xi}\bar{\xi} \\ IV \quad & \overline{a\zeta} = a\bar{\zeta} \\ V \quad & \overline{\zeta + \xi} = \bar{\zeta} + \bar{\xi} \end{aligned} \quad (1.2)$$

where a is a constant.

Stricto sensu, these relations are valid only when averages are by “ensemble” averaging (i.e., averaging over many realizations under identical conditions, Kaimal and Finnigan 1994). However, this is never possible in atmospheric measurements, so averages are most often computed on the basis of time series of statistical quantities by making use of the ergodic hypothesis which states that time averages are equivalent to ensemble averages (Brutsaert 1982; Kaimal and Finnigan 1994). To fulfil this assumption, the fluctuations have to be statistically stationary during the averaging time chosen (see Chap. 4).

Fig. 1.1 Schematic presentation of Reynolds decomposition of the value ζ (Foken 2008)



1.2.3 Scalar Definition

The following variables are commonly used in the literature (and throughout this book) to define the scalar intensity of an atmospheric constituent s : *density* (ρ_s , kg m^{-3}) and *molar concentration* (c_s , mol m^{-3}) represent the mass and the number of moles of s per volume of air, respectively. The *mole fraction* (mole mole^{-1}) is the ratio of the moles of s divided by the total number in the mixture (also equal to the ratio of the constituent partial pressure to the total pressure), the *molar mixing ratio* ($\chi_{s,m}$, mole mole^{-1}) is the ratio of the constituent mole number to those of dry air, and the *mass mixing ratio* (χ_s , kg kg^{-1}) is the ratio of the mass of the constituent to the mass of dry air. These variables are related by the perfect gas and the Dalton laws.

However, among these variables, only the molar and mass mixing ratios are conserved quantities in the presence of changes in temperature, pressure, and water vapor content (see Kowalski and Serrano-Ortiz (2007) for a more complete discussion). Unfortunately, the variables that are directly measured in the field by infrared gas analyzers are rather density and molar concentration, quantities that are not conserved during heat conduction, air compression/expansion or evaporation, and water vapor diffusion. Therefore, variations in these quantities may appear even in the absence of production, absorption, or transport of the component. The corrections that are necessary to take these effects into account were extensively discussed by Webb et al. (1980) and reexamined by Leuning (2003, 2007). They will be presented in Sect. 4.1.4.

The conservation equations developed in the section below are written using the mass mixing ratio but, for convenience, the other variables will also appear in this book. Conversion factors of one variable into another are given in Table 1.2.

Table 1.2 Conversion factors between different variables characterizing scalar intensity

Conversion factor	Molar mixing Ratio, $\chi_s =$	Mass mixing Ratio, $\chi_{sm} =$	Molar concentration, $c_s =$	Density, $\rho_s =$
Molar mixing ratio, $\chi_s \times$	1	$\frac{m_s}{m_d}$	$\frac{p_d}{R \bar{\theta}}$	$\frac{m_s p_d}{R \bar{\theta}}$
Mass mixing Ratio, $\chi_{sm} \times$	$\frac{m_d}{m_s}$	1	$\frac{m_d p_d}{m_s R \bar{\theta}}$	$\frac{m_d p_d}{R \bar{\theta}}$
Molar concentration, $c_s \times$	$\frac{R \bar{\theta}}{p_d}$	$\frac{m_s R \bar{\theta}}{m_d p_d}$	1	m_s
Density, $\rho_s \times$	$\frac{R \bar{\theta}}{m_s p_d}$	$\frac{R \bar{\theta}}{m_d p_d}$	$\frac{1}{m_s}$	1

Note that p_d corresponds to the dry air pressure (namely $p - p_v$). As a result, the exact conversion of mass or molar mixing ratio into concentration or density needs the knowledge of water vapor pressure (for details see list of symbols)

1.3 One Point Conservation Equations

The equation describing the conservation of any scalar or vector quantity ζ in the atmosphere may be written as

$$\underbrace{\frac{\partial \rho_d \zeta}{\partial t}}_I + \underbrace{\vec{\nabla}(\bar{u} \rho_d \zeta)}_{II} + \underbrace{K_\zeta \Delta(\rho_d \zeta)}_{III} = \underbrace{S_\zeta}_{IV} \quad (1.3)$$

where \vec{u} is the wind velocity vector, $\vec{\nabla}$ and Δ represent the divergence $\left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z}\right)$ and Laplacian $\left(\frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2}\right)$ operators, ρ_d is the dry air density, K_ζ is the molecular diffusivity of the quantity ζ , and S_ζ represents its source/sink strength. This equation is instantaneous and applies to an infinitesimal volume of air. It states that the *rate of change of the quantity* (I) can be due to its *atmospheric transport* (II) to *molecular diffusion* (III) or to its *production by a source/absorption by a sink* into the infinitesimal volume (IV). It can be applied to any scalar or vector quantity provided source terms are defined accordingly. In particular, if ζ is 1, Eq. 1.3 is the continuity equation, if ζ is air enthalpy, it is the enthalpy conservation equation, and if ζ is the mixing ratio of an atmospheric component (water vapor, carbon dioxide, etc.), it is the scalar conservation equation. If the quantity is a component of the velocity vector in one given direction, Eq. 1.3 expresses the conservation of the momentum component in this direction. The three equations describing the momentum conservation in the three directions constitute the Navier Stokes equations.

Application of these equations to the surface boundary layer requires application of the Reynolds decomposition rules: the variables ζ , ρ_d , \vec{u} , and S_ζ should each be decomposed into a mean and a fluctuating part according to Eq. 1.1, followed by application of the averaging operator, and appropriate rearrangement and simplification. This procedure will be applied to each equation below.

1.3.1 Dry Air Mass Conservation (Continuity) Equation

By replacing ζ by 1 in Eq. 1.3, one obtains

$$\frac{\partial \rho_d}{\partial t} + \vec{\nabla}(\bar{u} \rho_d) = 0 \quad (1.4)$$

as there is neither a source nor sink of dry air in the atmosphere. Application of the time- averaging operator gives immediately:

$$\overline{\frac{\partial \rho_d}{\partial t}} + \vec{\nabla}(\overline{\bar{u} \rho_d}) = 0 \quad (1.5)$$

1.3.2 Momentum Conservation Equation

By replacing ζ in Eq. 1.3 with the component of wind velocity in one given direction, u_i , one obtains the momentum conservation equation in this direction:

$$\frac{\partial \rho_d u_i}{\partial t} + \bar{\nabla} \cdot (\bar{u} \rho_d u_i) = S_i \quad (1.6)$$

In Eq. 1.6, the source/sink terms correspond to momentum source/sink, namely to forces. Forces that can act on air parcels in the atmospheric boundary layer are drag, pressure gradient, Coriolis forces, viscous forces, or buoyancy. The first three forces are considered negligible for a flat, horizontally homogeneous surface boundary layer above the roughness elements (i.e. not including vegetation) (Businger 1982; Foken 2008; Stull 1988). Buoyancy appears only in the equation for vertical momentum. The horizontal component of momentum parallel to the mean wind is dominant in the surface boundary layer and thus the buoyancy term is not considered. In a Cartesian coordinate system (x, y, z) where x corresponds to the horizontal, parallel to the average wind velocity, y to the horizontal, perpendicular to the average velocity, and z to the vertical; u, v, w are the $x, y,$ and z components of velocity, respectively, and this equation is written as

$$\frac{\partial \rho_d u}{\partial t} + \frac{\partial \rho_d u^2}{\partial x} + \frac{\partial \rho_d v u}{\partial y} + \frac{\partial \rho_d w u}{\partial z} = 0 \quad (1.7)$$

Application of the Reynolds decomposition to Eq. 1.7 and use of the following simplifications (Businger 1982; Stull 1988):

$$\begin{aligned} I \quad & |p' / \bar{p}| \ll |\rho'_d / \bar{\rho}_d| \\ II \quad & |p' / \bar{p}| \ll |\theta' / \bar{\theta}|, \\ III \quad & |\rho'_d / \bar{\rho}_d| \ll 1 \\ IV \quad & |\theta' / \bar{\theta}| \ll 1 \end{aligned} \quad (1.8)$$

where p is the pressure and θ the air temperature, leads to

$$\frac{\partial \bar{u}}{\partial t} + \bar{u} \frac{\partial \bar{u}}{\partial x} + \bar{v} \frac{\partial \bar{u}}{\partial y} + \bar{w} \frac{\partial \bar{u}}{\partial z} + \frac{\partial \overline{u'^2}}{\partial x} + \frac{\partial \overline{v'u'}}{\partial y} + \frac{\partial \overline{w'u'}}{\partial z} = 0 \quad (1.9)$$

Equation 1.8, III corresponds to the *Boussinesq-approximation* (Boussinesq 1877), which neglects density fluctuations except in the buoyancy (gravitation) term, because the acceleration of gravity is relatively large in comparison with the other accelerations in the momentum equation. By choosing a coordinate system such that