Simsphere Workbook

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Simsphere Workbook: Preface

Purpose

This book is intended as a hands-on introduction to understanding some land surface processes as they pertain to interactions between the soil, atmosphere and plant continuum. It was written over a period of about 15 years beginning in the 1990s and is intended not only to address questions posed in the text, but to help the readers formulate answers to their own questions, either for academic or professional purposes. The subject matter extends over a number of different fields (or disciplines)—meteorology, agronomy, horticulture, hydrology, forestry and geography. The material, either explicit or implicit, in the book may be of practical importance to practitioners and students alike.

A unique aspect of this book is the inclusion of a computer model (described below in part 3) along with the text. The subject matter does not cover all aspects of land surface processes but only those that can be described with the aid of this particular model. To make use of such a model, one must perform a simulation, so called because the execution is intended to simulate (or approximate) a natural process. What differentiates this book from others related to the atmospheric surface or boundary layer is that the simulations constitute an integral part of the text. Indeed, the model allows the reader to interact with the textual material and therefore become an active participant in the learning process. Through use of the model, the reader will obtain an understanding in a way that is not directly obtainable from any text. It is a Socratic style, except that the reader functions as his own Socrates, posing his own questions that are answered best by performing the appropriate simulation. Thus, the process is open ended by allowing the reader to explore aspects of land surface processes in ways that may not be conceived by the authors.

We envision that this text represents more than one level of training. One level is that of university students, including those who are exposed to formal courses in boundary layer meteorology, plant science, agronomy, or indeed in any of the various fields that are interwoven in this book. As a formal seminar course at Penn State, 'Biosphere-Atmosphere Interactions' has functioned successfully using a prototype of this book. In this course, we have worked with students from agricultural engineering, meteorology, geography, forest management and agronomy. Despite lack of knowledge in the other disciplines, graduate students taking this class have been treated as younger colleagues rather than as empty vessels, which require filling with knowledge as one would fill the gas tank of a car. The result has been a successful interaction with students that lead, on more than one occasion, to insightful revelations for the instructors, who are the authors of this book.

This book, like the seminar course, should serve as an introduction to some sophisticated concepts in land surface processes but without becoming highly technical. We believe that it is possible to understand the subject material and even to anticipate the outcome of the simulations by thinking physically and intuitively without recourse to complex physics and multitudes of equations. Therefore, we deliberately avoid use of equations (or express them in the simplest possible ways), specialized terminology or lengthy discussions of the physical, chemical or biological nature of the process. In that way a meteorologist, for example, will be able to understand how plants work, albeit on a somewhat superficial but highly informative level, whereas a plant scientist will gain an appreciation on a basic level how the atmosphere affects plants.

In combining material from diverse fields, we envision the end result to be a combination of subjects and components that exist hardly anywhere else in print. Such subjects are, for example, a discussion of the effect of carbon dioxide on transpiration or of thermal inertia on night time air temperatures, the importance of moisture at the earth's surface (as opposed to the root zone), the importance of feedback in mitigating or enhancing the importance of plants on microclimate, and other issues.

With its unique mix of scientific disciplines and its absence of excess terminology and mathematics, this book should provide a useful cross-disciplinary vista for the reader, providing novel insights to the workings of the plant, soil and atmospheric canopy and the interaction of the different components. To soften the odious problem of strange jargon, we will italicize all terms that represent technical terms in the common language of a particular discipline. Terms required for an understanding of the course are listed at the end of each chapter. Terms whose definitions are not fully accepted or are defined elsewhere are given in quotes. On the other hand, some technical terms introduced and discussed here are scarcely treated anywhere else, such as 'moisture availability' and 'auto-humidification'.

We intend this book for another level, the practicing specialists. By this, we refer to those who have professional interests in understanding how their discipline is related to the land surface processes, but whose domain falls in other disciplines. For example, a meteorologist might wish to learn more about how the plant canopy can influence the atmospheric boundary layer, but without the user becoming a plant scientist or obliged to read and understand a highly technical textbook. Similarly, an architect or city planner might wish to determine the solar energy absorbed on a roof of a certain colour or the effects on surface temperature by planting or removing trees. The text, however, does not necessarily provide explicit answers to these questions, although an understanding of the textual material, coupled with familiarity with the model, will allow the specialists to pose their own questions.

The chapters in the book must be both read and experienced. We recommend that one first read the material and then perform the simulations, chapter by chapter. Each chapter builds on the last, so that the best approach to understanding the material in this book is to undertake each chapter sequentially. Chapters divide into two sections, which we will simply call "Level 1" and "Level 2". In the first part of each chapter, Level 1, the material is introduced, concepts are discussed and a precise set of instructions are provided for carrying out the simulations in a step-by-step manner. The results of these simulations are discussed in the text and a set of questions at the end of the level are provided which are based on the output from the simulations. Level 2 introduces more complexity and suggests additional simulations and the possible outcome of these simulations but does not furnish a detailed treatment of the results. Moreover, these simulations may be open ended in the sense that their consequences may not be readily explainable by the text material, or that the results may be completely unforeseen even by the authors. The readers are encouraged to proceed to whatever level best suits their needs and expertise. In this respect, Levels 1 and 2 are roughly the equivalent of undergraduate and graduate level material.

It is our hope that the reader will not only learn from the material in the text but also appreciate the limitations of a physical model. Models are not nature but, at best, a somewhat realistic structure that allows one to gain some insight into the natural world. As the model constitutes an integral part of the text, the latter does not delve much deeper than the model is capable of describing. Therefore, use of external sources is encouraged in order that the reader may gain a much deeper insight into ideas addressed in the text. Several excellent texts already exist which cover a wide range of topics relevant to this book. These are included in the list of references, which are supplements to the material offered in this book. At times, the book adopts a light-hearted approach to its description of complex or contentious subjects. The authors offer an apology if we seem to have slighted or overly simplified these topics but we believe that it is important to demystify the subject matter to the extent that specialists in one field are not confused in trying to delve into someone else's field.

Finally, some plant scientists may object to our simplified and perhaps even facile depiction of plants as stick figures accompanied by explanations of the plant's behavior that may seem to 'dumb down' or incorrectly represent their true physiological responses. Besides the necessity of making extremely complex behavior clear to the non-specialist, our discussions are justified by and consistent with published measurements of sensible and latent heat fluxes over plant canopies.

Introduction

Our atmosphere has no real boundary except the surface of the earth. Move sideways and, unless one strikes a mountain or an object embedded in the earth, one will never encounter a boundary. The top of the atmosphere is bounded, literally, by almost nothing. Indeed, it is very difficult to precisely assign an altitude to the top of the atmosphere. It just gradually disappears with increasing height until one is left with a few ions whizzing about. The lower boundary, the earth's surface, can not precisely be defined either, but the effect of that surface on the atmosphere is real and profound. Looking down from above, however, one would be hard pressed to say what and where the surface is located. Is it the tops of the buildings, the surface of the vegetation, or the soil in and around the vegetation? Is it the faces of rocks, stones and other debris scattered about on the landscape? Yet, we can tell when our feet are on the ground and we can ask questions about how that ground affects the atmosphere even if we are not too sure of what and where that ground surface exists. We live at the surface and the atmospheric events that occur affect us as surface creatures in a variety of different ways.

The question is sometimes asked: "Why is it warm near the surface of the earth and cold at high altitudes, when the earth's surface is farther from the sun?" Of course, the sun really warms the earth, which then warms the atmosphere above it. Almost all of the solar energy not reflected away by clouds, the atmosphere or the earth's surface is absorbed at the ground or in material or objects that cover the ground, such as plants, houses, trees, etc. About 50% of the solar energy reaching the top of the atmosphere is ultimately absorbed at the surface of the earth on a sunny day at middle latitudes in summer. Thus, the radiator for maintaining the warmth of the atmosphere is the surface; the atmosphere itself is warmed only indirectly from the sun.

Still, the earth's surface would not exert as much variability on the atmosphere if it were everywhere uniform or if all the sun's radiation absorbed by the surface were used to heat the air in contact with it. Thus, partitioning of solar energy at the surface is highly variable and highly dependent on the nature of the surface. Some of the absorbed solar energy is stored in the ground or locked away in the molecules of evaporated water. Look around you at the landscape and you will see an infinite variety of objects comprising the ground surface. No two points on the ground are identical. Consider any region one square kilometer in area. It consists of a kaleidoscope of objects, textures, colours. The terrain is not flat. There are small hills, gullies, knolls. There are people, dogs, trees, bushes, grass, soil. Focus in on any 10 square meter area within that larger square and the surface is equally varied, perhaps with patches of vegetation, or a house, part of a road, a little stream. Focus in on a 1 square meter or 1 square centimeter area within that larger square and the variegated nature of the surface remains evident, albeit on a smaller scale. The soil is speckled with colour, rocks protrude, plants or bits of vegetation are visible. Leaves, some shiny, some dull, are orientated in various directions, deflecting and absorbing the sun's rays, trapping the heat or moisture within the vegetation canopy, shading the surface. The variations are endless no matter where we look.

These variations in the ground surface figure intimately in the distribution of the solar energy into its components, as well as in the reflection of that solar energy. Those components, which are given the names of sensible and latent heat—quantities which are to be described and defined in the text—profoundly affect the local climate, the so-called 'microclimate' (from the Greek root micro, which in its Anglicized form means small). Because the earth's surface is so varied, the microclimate is also endlessly varying on all scales from the very smallest to the size of a region. For example, the sea breeze is familiar to inhabitants along the New England coastline as a cool breeze which pushes inland from the ocean on warm days. This phenomenon is produced by the contrast between the heated land surface and the cooler water, which establishes a pressure gradient from sea to land that drives the cooler air inland. The sea-breeze front formed at the leading edge of this cool air may move several kilometers inland during the day and extend hundreds of kilometers along the coast. As the warm air is displaced, vertical currents of air develop which cause clouds, and occasionally heavy thunderstorms, to form along the sea breeze front. When the day's heating over land ceases in the evening, the sea breeze front recedes over the ocean, becoming a land breeze.

Yet on a much smaller scale, the same process that makes a sea breeze is occurring on a sunny afternoon between the expanse of macadam in a shopping center parking lot and a surrounding grassy meadow or between wet and dry sand along the seashore. One can easily verify the vast difference in surface temperature by walking barefoot across the parking lot or dry sand to the meadow or wet sand. Of course, it is unlikely that clouds and thunderstorms will form along the edge of a parking lot or between dry and wet sand, but it is nevertheless possible for differences in atmospheric pressure to be produced by such differences in temperature, which would induce horizontal and vertical currents of air. Such hot spots are also found in nature even when the surface has not been altered by humans. Birds soar in thermals produced over the heated face of a rocky outcrop along the sides of a mountain. Even different types of vegetation utilize the solar energy differently, so that subtle horizontal temperature variations may be produced over an inhomogeneous vegetated surface, e.g, a corn field adjacent to a forest. Regardless of the scale, wherever there are temperature gradients at the surface, there are also pressure gradients which force the air to move horizontally and vertically.

It is not always readily evident what processes are responsible for a particular microclimate. Consider the example of a white cement sidewalk aside a strip of green grass on a sunny summer's day. If one considers only the amount of absorbed solar energy, it would seem reasonable to conclude that the sidewalk should be cooler than the grass because it is reflecting a much greater fraction of the solar energy than the darker grass. It would be immediately evident to anyone crossing the sidewalk with bare feet, that the sidewalk is warmer than the grass. The reason, of course, is that the grass stays cool by evaporating water and the sidewalk has no means to cool itself. That is precisely how humans can maintain a reasonable body temperature while exercising on a hot summer day by their own sweat. Similarly, grass uses much of the solar energy to evaporate water rather than to raise the temperature of the grass and the air. By contrast, all of the sun's energy is used to heat the sidewalk and the air above it because there is no water available for evaporation. Yet it is also possible to imagine a sidewalk that reflects almost all of the incoming solar energy, in which case the sidewalk might well be cooler than the grass surface. We can deduce the physical reasoning behind the result that the sidewalk is warmer than the grass only by measurement or by experience but we are not sure, a priori, which process (evaporation or reflectivity) will dominate.

In fact, it is possible to contrive situations in which the intuition might prove faulty, such as in our imaginary scenario in which the sidewalk material is highly reflective. How could we be sure which surface would be the hotter, the sidewalk or the grass, in the absence of measurement? Further, imagine that the sidewalk is sloping at an angle to the sun and the grass is a horizontal surface. What is the answer now regarding which surface is hotter? The problem is even more complex in this case because the slope of the surface, the reflectivity and the evaporation are factors that must be considered.

This type of problem can be of considerable practical importance. For example, imagine a scenario in which the two sloping sides of a roof face at different angles to the sun at different times of the day and one wishes to determine which roof face absorbs the most sunlight and how much each face absorbs during the course of a day. Perhaps one would also wish to alter the reflectance of one side of the roof, such that the solar energy absorbed by both sides of the roof on a sunny day would be equal. The solution to these problems would be difficult to intuit but they could be obtained by calculation if we knew the properties of the surface.

As soon as one tries to calculate something, however, it is necessary to resort to an idealized 'model'. We may know the parameters in the problem, at least approximately, but our calculation would refer to an idealized sloping roof surface not the actual roof itself. We might choose a simple model of the roof—a sloping surface—and make some assumptions about parameters involved or the geometry of the roof. We might wish to ignore evaporation and consider only slope, azimuth and reflectivity of the roof. As we undertake this calculation we realize that many more factors are involved than just the slope, azimuth and reflectivity of the roof surfaces. For example, the thermal conductivity of the roof might need to be considered, or the wind speed and roughness of the roof surface.

Adding to the complexity is the interrelationship between parameters. The reflectivity or the thermal conductivity of a surface, such as the sand along the beach, might depend on the amount of water in the surface, as does the evapotranspiration. An increase in one of these parameters might actually cause a second one to decrease and a third one to increase, but the increase in the third parameter might then cause the second one to increase and the first one to decrease. For example, as the soil dries out it becomes lighter in color, reflecting more solar energy. Drying out of the soil causes it to become hotter, as the water content in the soil is gradually exhausted, but the whitening of the soil would reduce the solar energy absorbed in the sidewalk. The result is a tug of war between the drying out and the whitening, each affecting the heating of the sidewalk to a different extent. This tug of war between the different parameters in a system is an aspect of what is called feedback. Feedback processes constitute a central aspect of this book and a focus of the later chapters. An advantageous attribute of the model is that it allows one to investigate various feedback processes in an evolving system.

Feedback is a sophisticated concept, which has applications in some very technical areas, such as electronic signal processing and the design of high performance aircraft. How can one describe feedback between soil, atmosphere and plants qualitatively, that is in words, at the level appropriate to this book? The answer is that it cannot be done adequately without a model that allows the feedback processes to be implemented and studied by the reader. The formal expression of these mathematical interrelationships in this model is called a 'parameterization'. Parameterization simply means that one writes out equations governing these relationships between all the parameters; a set of parameterizations constitute a model.

The model's purpose is to simulate some of the processes and feedback in the soil-vegetation-atmosphere system. The model that describes the soil-vegetation-atmosphere continuum and the transfers of energy, momentum and water through it. Such a soil-vegetation-atmosphere transfer (or SVAT) model refers to a system that is almost infinitely complex. Yet, no model can be so complex in that it precisely describes the natural world. So, a simplification must be made. As one often does not know the correct mathematical description of all internal components in the SVAT system, the ultimate complexity of the SVAT model is limited as much by one's ignorance as one's tolerance of uncertainty in the results and the limitations imposed by cost in time and resources to construct the model. Moreover, were it possible to construct a model exactly as complex as nature itself, that model would be as difficult to understand as the un-parameterized physical world it represents. In that respect, nature is the ultimate model but such a model might be quite bewildering to apply and to comprehend even if inserted into a computer. Thus, the model can answer questions, even some not envisaged by the user. However, it is unable to answer all questions. Moreover, it can address specific problems only if the questions are well-posed. A silly or absurd result, or even an abortion of the simulation, may simply be the model's way of telling the user that the input parameters are not reasonable. Like a computer or an automobile, the model will function properly if operated properly. We often read of spectacular errors made by a bank or insurance company that are

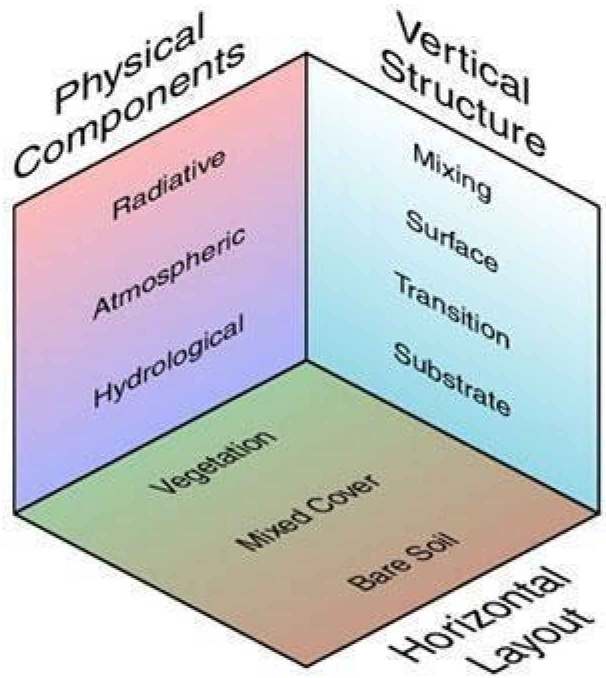


Figure P.1 Basic structure of the model

attributable to computer error. However, computers very rarely fail. Such errors are usually the result of improper use or incorrect data. To run the SVAT model, no special computer skills are required. The simple build and run process is described in a short user's guide provided with the program source code.

Structure of the Book

The simulation model

The Soil-Vegetation-Atmosphere Transfer (SVAT) computer simulation model is one-dimensional. It describes various processes in a column that extends from the root zone below the soil surface to a level well above the surface canopy. For convenience, let us refer to the surface canopy as the mix of vegetation with underlying soil and surrounding bare soil. We will also refer to the vegetation canopy by itself. The latter consists of just the vegetation and underlying soil. The processes and quantities described by the model evolve in time during a day and night (up to 24 hours). Output of the model consists of the surface energy (sensible, latent and stored heat) fluxes at the soil surface, in, around, and above the vegetation canopy, the flux of momentum above the canopy, the transfer of water in the soil and in the plants. It also includes the flux of carbon dioxide between the atmosphere and the plants (the carbon assimilation rate), the temperature of the soil, leaves and air above the vegetation canopy, the radiometric temperature of the vegetation and soil mix, the leaf temperature, the ground temperature beneath the leaves, and in the bare patches surrounding the vegetation, the temperature of the interleaf air spaces, the wind velocity, temperature and humidity of the air above the surface canopy and other parameters.

Although the model is limited to one dimension, it has an extra dimensionality in that it implicitly refers to a horizontal area of undefined size that contains a mix of bare soil and vegetation. Both vegetation and bare soil in the model function separately but they can interact. Thus, one can specify a fraction of vegetation cover and other attributes of the vegetation and bare soil fractions without having to know the size of the area. Interpretation of the results are, of course, a matter necessarily evaluated by the user. The model is limited, however, in its description of various aspects of the soil, plant, atmosphere continuum, such as the water transfer in the soil, the physiology of plant leaves, the structure of the roots, the description of the plant canopy during the night, etc. Because the model is an imperfect representation of nature, the user should always question the validity of the output in terms of whether the results make physical sense. Sometimes these results are counter-intuitive, but correct. The reader should always be cautious. Caveat Lector! This is a healthy attitude to adopt for any text and by any scientist toward either model simulations or measurements.

Contents of the chapters

Each of the 13 chapters contains a description of the subject, with the aim being to establish enough knowledge and critical thinking to execute and evaluate a set of computer simulations. The chapter begins with a short text discussing the nature of the subject at hand and introducing some background information and perhaps a few equations and some physics necessary to understand the simulations. Simulations follow the introductory text. These are carried out as exercises, in which the user modifies certain parameters in the input data, executes the model and then views the resulting output products. A standard input file is provided with the SVAT model distribution which is the starting point for these simulations. Should users become confounded in any way after making changes to the input file, the original stock version can be recovered from a fresh copy of the distribution to start over. The first set of simulations, 'Level 1,' are executed with close reference to the text. Input parameters are listed separately in a table. Output products must be evaluated. Level 1 concludes with a series of questions for the user. Level 2 is more free form, an adventure for the user. It begins with some more description, followed by suggestions for further simulations. Level 2 poses questions that might be answered by the user but which may have no obvious solution. In these cases, the design of the simulations and the exact choice of parameters are provided by the user. We list a series of terms to be looked up elsewhere or are particular to the text at the end of the chapter. Equations used in the text and a list of references for the material contained in the chapter are at the end of the text.

No detailed knowledge of any one of the disciplines comprising the text material is assumed. For example, long wave (thermal) radiation is mentioned, but the Planck black body relationships are not discussed in any detail, nor is the geometry of radiation or the concept of extinction coefficients. This type of material can be obtained in other texts, such as those referred to at the end of each chapter. If the reader has a casual acquaintance with thermal radiation, it will be possible to understand all that is required in the text with the aid of the simulations and some thought. Outside reading is highly recommended but not absolutely necessary for understanding.

Micrometeorology is a complex and often mathematical science, but much of the subject can be appreciated without recourse to mathematics. For example, we all know that when the sun shines, the ground warms, puddles dry up, the winds become gusty, thermals form allowing birds to soar, grass and flowers grow. At night temperatures decrease, but remain higher in cloudy conditions, puddles don't dry up as easily and the winds are calmer. What is not so intuitively obvious though are the non-linear interactions between physical components. We are interested in the nature of these interactions rather than a description of its components for their own sake. To gain an initial understanding of the resultant behavior of a system we eschew confusing details and concentrate on the resultant effects.

It is our hope that this course will thus prove useful for the non-specialist and, more precisely, to those who are interested in studying problems involving interactions between the vegetation and the atmosphere. We want to avoid the deeply technical side of many questions because equations and jargon are digressive and often divide rather than unite. Accordingly, we will avoid discussions of such topics as "higher-order closure theory" or Monin-Obukhov static stability functions and try to use intuition where possible. To address such issues we must begin with some elementary concepts such as the surface energy balance. Our ultimate goal is to sharpen up one's intuition so that we may begin to answer such questions without recourse to a computer.

Biosphere

Strictly speaking, the biosphere contains all living things and, as such, extends from deep inside the earth into the upper atmosphere. Conceptually, zones of interaction within the biosphere can be thought of as layers beginning with the ground (substrate) layer and extending upward through the ground-air transition layers, and finally into atmospheric layers above (Figure P.2).

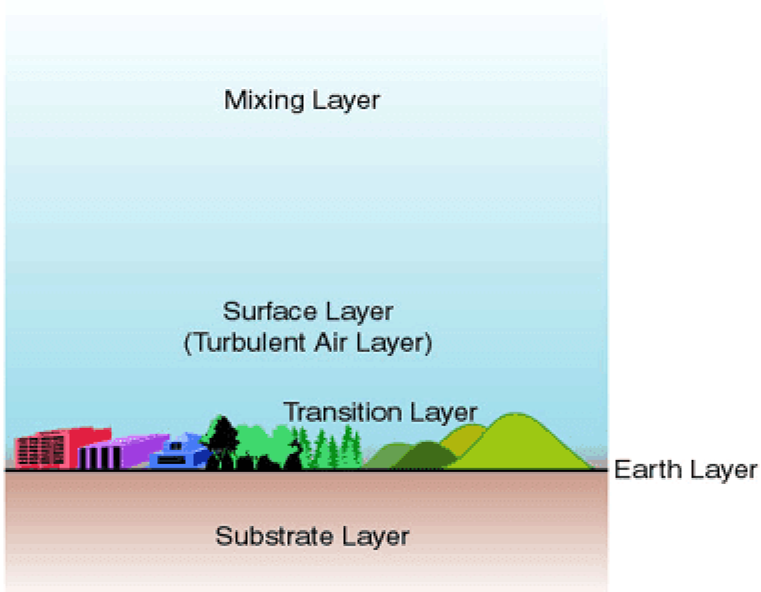


Figure P.2. Representation of mixing, surface, transition, earth and substrate layers.

We will restrict our definition of the biosphere to the domain most immediately affected by vegetation; a layer extending from just inside the ground (the root zone; say a meter or two) up to a height somewhat above the vegetation canopy, say a few tens or hundreds of meters above the ground. As such, the biosphere is where we spend most of our time and so, is of direct importance to us. Although we will need to start the course with a general discussion of the driving force for land surface processes, the surface energy balance, we will later address questions concerning how plants modulate the land surface processes and thereby how they affect the biosphere.

Feedback Systems

Feedback is a word that is used from time to time in this course. Feedback systems exist throughout nature and are inherent in complex, non-linear systems. The mathematical model upon which this course is based is highly non-linear and thus is able to capture a number of feedback processes found in nature.

Feedback is most often used in common language to describe the effect one encounters when sounds from a speaker are fed back into the amplifier, leading to a sudden, high-pitched squeal that usually sends the person in charge of the sound system scrambling to change the volume setting and orientation of the speaker. A thermostat is an example of feedback produced by a servo-mechanism, which reduces the heat generated, say by a furnace, when the temperature reaches a certain preset level, and it restarts the heating process when the temperature falls below the preset level. That type of feedback is referred to as *negative feedback* because the effect of rising temperature is to counteract the rising temperature. Suppose the thermostat were set such that an increase in temperature created more heat output from a furnace. In that case, the heating process would run away with itself and the room would become hotter and hotter. This is called *positive* feedback. Positive feedback is responsible for the squeal of sound in the speaker. Thus, negative feedback may nearly compensate for a reference input, reducing its effect on the output, and positive feedback amplifies the input.

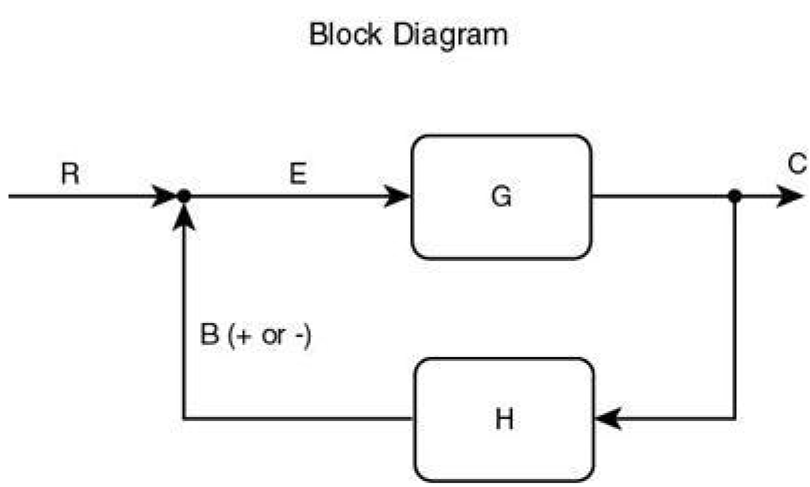


Figure P.3. Feedback loop.

We can illustrate the effect of feedback by Figure P.3 above. Let's demonstrate the figure using a somewhat facetious example. Suppose that a dispute takes place in a bar room, perhaps produced by a fleeting glance by Mr. Look at the girl friend of Mr. Grouch, who begins to get very angry at the wayward glance. The feedback loop consists of a single transfer function G, representing the forward component. We could imagine an input R, which is the wayward glance, anger by Mr. Grouch, which is the feed forward control element G, and some result of this possible encounter C. Now, let's say that Mr. Look has a friend who is respected by Mr. Grouch; We'll call him Mr. Sooth. Mr. Sooth mediates the situation through an intervention denoted by the letter H. Thus, Mr. Grouch is calmed somewhat by the effect of Mr. Sooth and the atmosphere returns to relative calm. This is a negative feedback situation, denoted by the feedback element B, which, in this case, has a negative sign. Now, suppose Mr. Grouch hated Mr. Sooth, so that the sound of Mr. Sooth's voice only made Mr. Grouch angrier. The result of intervention H would be a positive feedback (B positive), such that Mr. Grouch only becomes angrier by the intervention. Mr. Sooth does not take the hint but continues to ply Mr. Grouch with soft words that only make Mr. Grouch angrier. The result is a blow up—a fight.

Only in extreme cases, however, does negative feedback entirely compensate for the input, although we will later present a case of feed forward in plants in which the input is reversed, not simply mitigated. In other examples (we hope you do not discover these in running the model) the effect may be a positive feedback leading to a blow up (mathematically speaking, a catastrophe) in the solution of the model. Some of these pitfalls may be lurking in this model, waiting for the user to choose just the right combination of input parameters. If this should happen, we advise the user to re-evaluate the input parameters or change them slightly. Sometimes, unwanted feedback may occur because the user has chosen unreasonable input parameters. In this case, the model's behavior should be a signal to the user that something is wrong in the initial conditions.

The mathematics of feedback systems can become very complex. We do not intend to delve into feedback processes in any more depth than we have done in discussing Mr. Grouch's problem. However, we will describe some rather interesting feedback processes with regard to plant regulation of their environment.

Simsphere Workbook: Chapter 1

Surface Energy Balance in Simsphere

Introduction

This chapter addresses the issue of the surface energy balance. We begin with just a horizontal, bare soil surface and discuss the types of energy fluxes at the earth's surface as related to the absorbed solar energy, the nature of a flux, and the effect of the absorbed solar energy on the surface air temperature and the radiant temperature (in the thermal infrared) of the surface. We then ask what happens to these fluxes when the latitude is changed, the season is changed and the angle of slope and azimuth of that sloping surface is changed. Level 1 concludes with a series of simulations in which these parameters are varied and the output discussed. Level 2 addresses such things as the total amount of solar energy absorbed as a function of latitude, when applied to real-world situations, such as the example of the ski slope. Further simulations are then suggested. A set of default input parameters is furnished. If you make no changes to these parameters, the simulations proceed using this input. Changes in the input parameters are assessed by comparing the output fields with those obtained with the default input parameters.

The purpose of this chapter is to develop a feel for the shape and magnitude of the turbulent fluxes (*i.e.*, the partitioning of the net radiation) over a relatively simple surface; that of bare soil. Be sure to do this for differing locations, time of year and altering topography.

First, a word of caution. The expression 'all other things being equal' applies to these simulations. The purpose of using such a model is to pose simple questions that you can answer by making simple simulations. While the model is capable of capturing various feedback processes between differing components, the model is unable to answer all questions at once. For example, some simulations that you will make as part of this scenario show how the turbulent heat fluxes differ on different sides of a hillside *all other things being equal*. Although you may find more solar flux and more evapotranspiration on one side of a hillside than another, this flux will also depend on the soil moisture content, which you fix at the same value for all the simulations. In reality, greater solar flux on one side of a hill may cause that side of the hill to dry out more quickly, thereby resulting in less evapotranspiration than on the other side. Such an evapotranspiration differential could also lead to differing vegetation on the two sides of the hill. These vegetation differentials, in turn, affect the evapotranspiration. On the other hand, the reduction of latent heat flux on that side of the hill may result in greater sensible heat flux and stronger updrafts that could enhance the precipitation on that side of the hill, thereby replenishing the soil water content depleted by the greater solar demand, etc. Obviously, these processes lie outside the scope of the model and would require a full three-dimensional mesoscale model to capture all of the feedback mechanisms.

The Surface Energy Balance

Unlike government budgets, the surface energy budget must balance! Ignoring such trivial energy components such as friction, electrochemical interactions and photosynthesis, the remaining surface energy components - the net radiation $\mathbf{R_n}$, the latent heat flux $\mathbf{L_eE}$, sensible heat flux **H** and soil heat flux **G**, **should sum to zero**. If we consider an infinitely thin plane, representing the surface, we can assume that the plane surface has effectively no storage, simply because it has no mass. The balance for this hypothetical plane is shown in Figure 1.1.

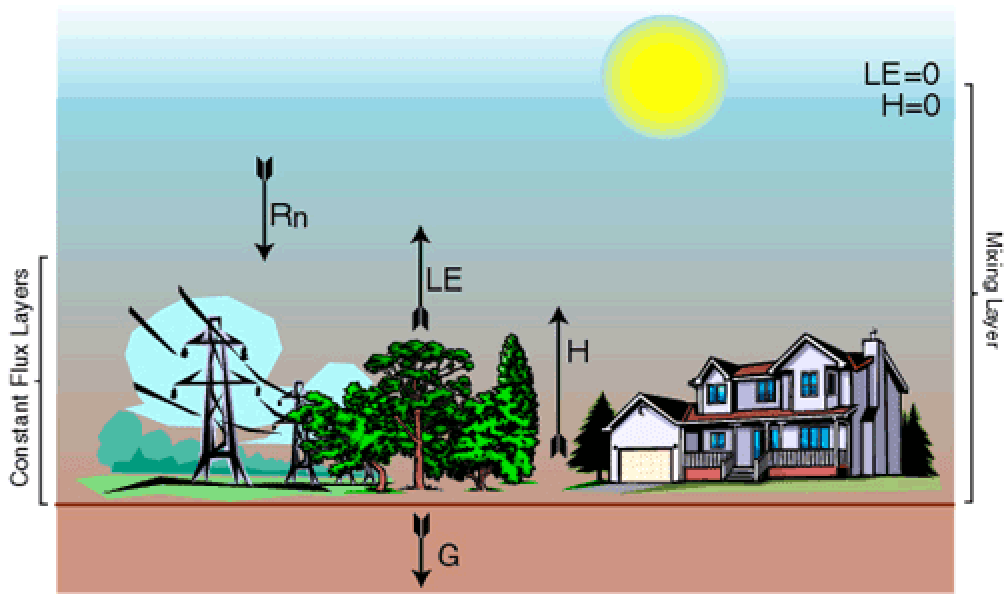


Figure 1.1. Schematic representation of surface energy fluxes: Net radiation (R_n), latent heat flux (LE), sensible heat flux (H) and ground heat flux (G). The fluxes listed at the upper right are zero at the top of the mixing layer.

A serious problem with the concept of the ground surface is that we have only to look about us to see that there is no planar ground surface: There are stretches of bare soil, but also crops, houses, trees, people, animals, cars, etc. The land surface is therefore a convenient artefact but it doesn't 'exist'. In imagining that it does for computational purposes, we have created a 'model'.

We will simply consider the ground surface, for the present, to be the layer over which the radiant energy from the sun is absorbed. This layer will be a few millimeters thick for 'hard' surfaces, but may be meters thick in depth for still water or for that matter in a dense plant canopy. To deal with this vast difference in absorbing depth we ignore water surfaces but we will eventually treat vegetation once the simple aspects of our model are explored.

Solar Radiation and Albedo

We will also bypass, for the most part, the attenuation of radiant energy by the atmosphere (although Simsphere does account for this attenuation using a rather simple algorithm) and start by imagining a flux of solar radiant energy ('irradiance') on the imaginary planar ground surface. Incident downward solar flux is given by the symbol $S\downarrow$ and upward (reflected) solar flux at the ground surface by the symbol $S\uparrow$. The ratio of incident solar to reflected solar energy is called the 'albedo'; i.e. $\frac{S\uparrow}{S\downarrow}$. Albedo is a tricky concept - not only does it vary with surface type, but also with sun angle (even over vegetation). Values obtained from the literature for albedos of various surfaces may be misleading since they may or may not refer to that surface as a pure entity. For instance, the albedo of some shiny leaves may be as high as 25%, although a canopy consisting entirely of that same plant may reflect much less than 25% of the incident solar flux, because of.Internal reflections between leaves. The leaf may absorb 25% every time it is struck by a sunbeam so each successive bounce attenuates the original ray; (the atmosphere, too, may reflect energy back to the earth). Alternately stated, your shag-rug may have a lower albedo than that for individual fibers, or a smooth rug made of the same material. We will also consider the surface to be 'Lambertian', which is to say that it reflects equally in all directions; in fact, reflected energy can be highly directional.

Nevertheless, it is often convenient (a key word in modelling) to consider the absorbed solar energy at the surface to be the incident solar flux multiplied by 1 - α_0, where α_0 is the surface albedo; (here, we will ignore, for the present, back reflections from the atmosphere).

Long Wave Radiation

Plank's law says that blackbody radiation is proportional to the fourth power of the temperature. The ratio of emitted radiation for an object to the blackbody radiation is called the 'emissivity'. An emissivity for a surface is actually equal to the absorptivity of that surface, according to Kirchoff's rule, but it is also equal to one minus the reflectivity of the surface. Reflectivity is actually the same parameter as albedo, but we usually speak of the latter only in reference to solar reflectivity. Since once minus the albedo is the absorptivity of a surface, one minus the emissivity also corresponds to an albedo, note though that the term albedo is usually reserved for the shorter (solar) wavelengths and emissivity for the longer (thermal) wavelengths. For a surface in thermal equilibrium, absorptivity and emissivity are equivalent; this is an aspect of Kirchoff's law.

Wein's law says that all bodies emit radiation at a wavelength inversely proportional to temperature. Because the sun is very hot (about 10,000 K on the outer surface), the solar energy spectrum is at low wavelengths (about 0.5 μ, which is close to the optimum sensitivity for the human eye). Surfaces such as the earth emit radiation at wavelengths close to 10 μm, as is consistent with its lower temperature (about 300 K). The symbol for this 'long wave' radiation is $L\uparrow$ for upwelling long wave radiation.

The atmosphere also emits thermal radiation because the atmosphere contains gases that are good absorbers of long wave radiation. Most solid objects, including water and cloud, are nearly perfect black bodies, which is to say that their emissivities are close to 1.0. The atmosphere, however, is mostly transparent to solar and long wave (thermal) radiation except for three constituent gasses: carbon dioxide, water vapor and methane – the so-called greenhouse gasses. Long wave radiation emitted from the surface is partially absorbed by these gasses, which then radiate upwards and downwards long wave radiation at the atmospheric temperature, which is generally lower than that at the ground. The greater concentration of these greenhouse gasses, the greater the back radiation toward the ground and the smaller the amount of surface radiation that passes through the atmosphere to space. Accordingly, the greater the concentration of greenhouse gasses, the warmer is both the atmosphere and the earth's surface.

Clouds will also absorb radiation from below and emit from their tops. Downward long wave radiation from cloud bases may greatly offset the upward long wave radiation from the earth, resulting in reduced nocturnal radiational cooling below cloud level and at the ground. If the cloud bases are warmer than the surface of the earth, as may happen during inversion conditions, night-time temperature may rise, rather than decrease with time. The symbol for downwelling thermal radiation as $L\downarrow$.

Net Radiation

After summing the absorbed solar radiant energy (by convention, positive), the outgoing surface thermal irradiance (negative) and the downwelling thermal irradiance (positive) we arrive at the net radiant energy (Eqn. 1.1 below). This energy constitutes a key driver for heating the atmosphere and the ground. Because the earth warms as the result of solar heating, a temperature gradient develops between the surface and the substrate below and the atmosphere above. According to the diffusion theory of heat transfer, sensible heat (that which is used to change the temperature) flows down the temperature gradient (ΔT), either away from the earth (by definition, positive in Eqn. 1.2 below) or towards the earth. Thus, at night the earth surface cools and the temperature gradient becomes reversed. Heat flows downward toward the earth. The direction of the sensible heat flux depends on the direction of the temperature gradient; (we will ignore counter-gradient flows, such as occur as the result of convection because resolution of this complication can only add unnecessary detail without changing our basic premise, which is that of gradient flow. It is best to remember, however, that the diffusive flow model is not always valid in nature. (You might like to ponder why this is so.)

The Representation of Fluxes

The division of net radiation into turbulent fluxes of heat and water vapor constitutes a central theme of this course. We will usually express water vapor flux E (kg m^{-2} s^{-1}) as latent heat flux (the former multiplied by the latent heat of vaporization L_e); this energy flux is designated by the symbol L_eE. Fluxes represent the amount of *stuff* (say water molecules or energy) that moves through a unit area (say one meter square) per unit time (one second). A unit of energy is a Joule and the flux of energy is expressed as Joules per square meter per second or watts per square meter; (one watt equals one joule per second). Fluxes are driven by gradients (the change in a quantity per distance). For example, the flux of water vapor is driven by a change in water vapor concentration over some distance. Moisture gradients in the atmosphere are expressed in units of 'specific humidity' q or 'vapor pressure' e. Sensible heat fluxes are driven by differences in temperature. The ratio of sensible heat flux to latent heat flux to latent heat flux is also an important parameter and is called the 'Bowen Ratio' β.

Some of the radiant energy serves to evaporate liquid water in the soil. Water vapor flux (or, alternately, latent heat flux), constitutes another component of the surface energy balance (by convention, positive upward). In principle, water vapor can flow down the vapor gradient and therefore, on occasion, towards the surface. However, the earth's surface does not moisten as the result of downward flux of water vapor, nor does dew form directly as the result of downward latent heat flux; (if forms because air cools below the saturation temperature as the result of long wave emission to space).

Except for the layer of air immediately in contact with the 'surface', the atmospheric fluxes are accomplished by eddying motions, rather than by the molecules. The efficiency of the atmospheric eddies (the eddy conductivity) in conveying sensible heat and water vapor away or toward the surface is very much greater than for molecular processes. The magnitude of these fluxes depend both on the wind speed (the mechanical shear) and the intensity of the heating (the convection); the total of which are limited by the net radiation R_n. In the case of heat and water vapor flux, the origin of the fluxes is virtually at the surface. Immediately adjacent to the surface turbulent eddies are relatively inefficient in conveying heat or water vapor and, accordingly, both molecular and eddy processes are co-equal in importance very close to the surface. This means that the flow of sensible and latent heat must first pass through a layer which is not very conducive to conducting it along.

Unlike heat and water vapor, the 'origin' for the flux of momentum from the air to the surface is above the surface at a fictitious level called the "roughness height" z_o. The roughness height reflects the irregularity of the surface and thus depends on the height and configuration of surrounding obstacles such as plants, buildings, etc. Even the individual boundary layers of leaves well inside a plant canopy are affected by the turbulent effects of the wind above the canopy and therefore by the roughness height.

Resistance and Conductance

Without going into much detail we would like to introduce another couple of terms related to the way in which we can view heat and water vapor being transported from the surface to the atmosphere: These are the 'resistance' (r) and 'conductance' (g). The resistance is simply defined as the ratio of the differential in the quantity being flux to the flux itself, *e.g.*, for sensible heat flux, H, which is driven by a temperature drop between two points, the resistance is defined as $\frac{\Delta T}{H}$. Here the temperature drop is expressed not as a gradient, which is the ratio of the change in the quantity per unit distance, but simply as a change in temperature alone. We can think of a resistance as the parameter that regulates the flux of something—in this case sensible heat H—across the gap over which a gradient could be calculated if one knew the distance across an imaginary gap between two points, much as an electrical resistance regulates the flow of current (flux) across a voltage drop. Conductance is simply the inverse of the resistance.

Similarly, a resistance for latent heat flux is $\frac{\Delta q}{L_* E}$. Resistances customarily have the units of time per length (*e.g.*, s m^{-1}). Conductance has the inverse units, or that of a speed. Given these definitions, can you find the appropriate constants that allow resistance to have the units of s m^{-1} for heat and for water vapor, given the atmospheric air density (kg per cubic meter), the latent heat of vaporization (Joules per gram) and the specific heat of dry air at constant pressure (Joules per kg per degree C)? Note that in order for these units for resistance to be consistent with the aforementioned equation, other quantities (largely constants) must be included in the formula, such as density, latent heat of evaporation, etc.

The concept of resistance comes from electrical circuit theory and was first proposed by Monteith (1975). In this formulation, flux is analogous to the current **i** and the potential drop of specific humidity, temperature, etc. to the voltage drop **V**. The flux equation $\Delta T = Hr$ is analogous to the current equation **V = i r**, where V is the voltage, i the current and r the resistance. Within this conceptual framework it is possible to construct series and parallel resistors (and even elaborate bridge circuits) to represent fluxes in various systems, such as plants. (It is even possible to represent capacitance as a storage term and thus to design RC circuits that depict fluxes through a system). Many plant scientists favor the conductance notation. The advantage of resistance over conductance notation is that the system can be treated mathematically in the same manner as electrical circuits. For example, in such a system resistances sum along a series; *e.g.*, the resistance to heat or moisture flux between the surface and 10 m above the surface as the sum of the resistances between the surface and 2 meters plus that between 2 meters and 10 meters.

Note, do not confuse conductance with 'conductivity', for instance, the thermal conductivity. Conductivity is defined in terms of a gradient, whereas conductance is defined in terms of a potential drop. Thus, the thermal conductivity, λ, is defined as the ratio of the heat flux, say the ground heat flux G, divided by the temperature gradient i.e. $\frac{\partial T}{\partial z}$, whereas thermal conductance is the ratio of heat flux divided by the drop in temperature across some distance. The two terms, conductance and conductivity, are related, however; in this case the distance in question is Δz. In calculating conductance or resistance in the atmosphere, a distance Δz. is implicit as the separation across the specific resistor. In subsequent discussions, the subscript H will signify the resistance for heat flux and also for water vapor flux.

Resistances for water vapor and sensible heat flux are almost <u>identical</u> (within the error of measurement); accordingly, the Bowen ratio can be expressed in terms of the ratio $\frac{\Delta T}{\Delta q}$, where the delta quantities refer to differences in temperature and specific humidity in the vertical. Since this ratio is proportional to the ratio of heat to water vapor flux, a Bowen ratio measurement, constituting a pair of temperature and specific humidity measurements on a vertical mast, can be used, in conjunction with the surface energy balance equation, to determine vertical fluxes of heat and water vapor. Indeed, the Bowen ratio approach is one of the principal means for measuring the fluxes in the field. For elaboration upon this idea please refer to the measurements section in Monteith's book..

Resistances vary considerably with the amount of atmospheric turbulence (which itself depends on the sensible heat flux as well as on the roughness length and the wind speed); resistances to heat, water vapor and momentum tend to be lowest during midday and largest at night. Can you think of a physical explanation for the variation of atmospheric resistance between day and night?

Latitude and Seasonal Variations

We have all observed that both the sun and the air are hottest during the middle of the day. Yet air is not directly heated by the sun. Most of the sun's energy reaches the ground where, in the form of radiant energy, it is either reflected or emitted back to space or, in the form of heat energy, it is either stored (temporarily) in the ground or transferred back to the atmosphere in the form of latent or sensible heat fluxes. Over the ocean the ambient air temperature hardly varies by more than a fraction of a degree centigrade (or if you wish to be truly American Fahrenheit) during the day and may actually be warmer at night in some instances. Let us then investigate this diurnal behavior of a net radiation and surface temperature, as well as the surface energy fluxes.

The Effect of Terrain Slope, Solar Elevation, and Azimuth

Take a flashlight and aim it directly toward the earth. The beam makes a roundish spot on the ground. Rotate the flashlight toward the horizon, so that the spot becomes broader and finally disappears, as the light is made horizontal. Shine that same flashlight, still horizontal, but perpendicular to a wall and the illuminated spot reappears. Rotate the light horizontally so that the angle of the beam with the wall, initially 90°, changes. Once again, the spot becomes oblong and increases in area.

The analog of the flashlight holds for the sun's direct beam, which illuminates a larger area when the sun is lower above the horizon (Figure 1.2). The diffuse (or scattered) part of the beam comes from all angles and the absorbed part, of course, disappears as the beam becomes horizontal. Late (or early) in the day, or even at noon in high latitudes, especially in winter, the sun's beam makes a small angle (the elevation angle) with the horizontal. A sloping surface (such as a hillside) that happens to make a 90° angle with the sun's beam will experience direct sunlight resembling that for solar noon (sun directly above), but the sun must also shine through a greater depth of the atmosphere than for the case of an overhead sun. A nearly vertical wall oriented perpendicularly to the sun near dawn on a winter's day still won't receive the amount of solar radiation equivalent to the tropics; a greater amount of energy would have been scattered and absorbed as the beam passes through the greater atmospheric depth. The point to be made here is that the smaller the sun angle with a surface, the smaller the flux per unit area (the flux being the same but the area larger).

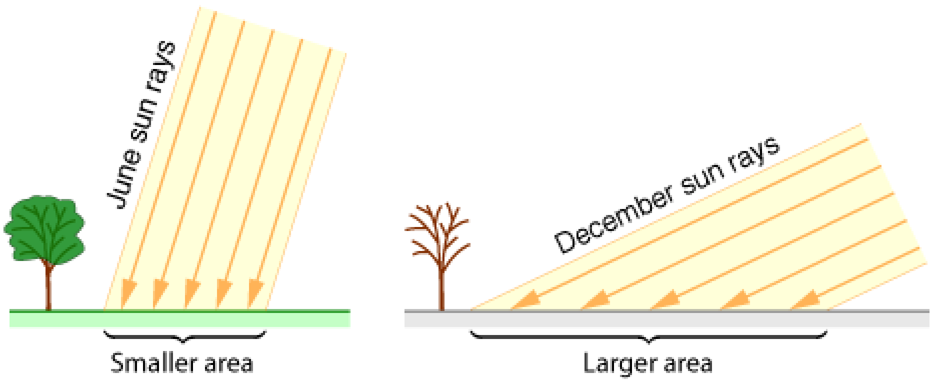


Figure 1.2. The two beams of sunlight above are identical but cover different amounts of earth area due to the beam angle. A perpendicular beam covers the smallest area while lower angles cover a larger area making the energy received per unit area less for lower sun angles. (Image courtesy of Daniel V. Schroeder, http://physics.weber.edu/schroeder/ua/SunAndSeasons.html.)

Since the sun tends to remain in the southern half of the sky at mid-latitudes in the northern hemisphere (except for near dawn and dusk, when it is located in the northeast or northwest), ground slopes facing the south (orientation 0° azimuth in the model) will receive more sunlight than those facing north. Similarly, slopes facing east will receive more sun in the morning than in the afternoon, and vice-versa for the west. You can invent other scenarios. In general, up to a point, the greater the slope the greater the difference from a horizontal ground surface and the more perpendicular the sun's rays become toward that surface (assuming it's tipped toward the sun). (Can you intuit approximately at what elevation angle would a sun-facing slope receive the greatest direct sunlight for a sun 40° above the horizon?)

Clearly, both the 'elevation angle' and the 'azimuth angle' of the terrain are important parameters in determining the amount of solar energy received on a horizontal surface. In your simulations, elevation angle ("slope" in the model) is the angle between an imaginary horizontal surface and the ground terrain indicating how much the terrain is tilted above a horizontal position. The azimuth angle ("aspect" in the model) of this tilted surface is measured from due south with zero degrees of azimuth facing due south. Negative azimuth values are measured counterclockwise from south toward the east, and positive values are measured clockwise from south to the west. 180 degrees of azimuth faces due north. This definition also suits meteorologists who define a north wind as that which blows toward the south (zero degrees). So also does the sun's elevation angle because that affects the amount of solar energy attenuated from the beam.

Simulations

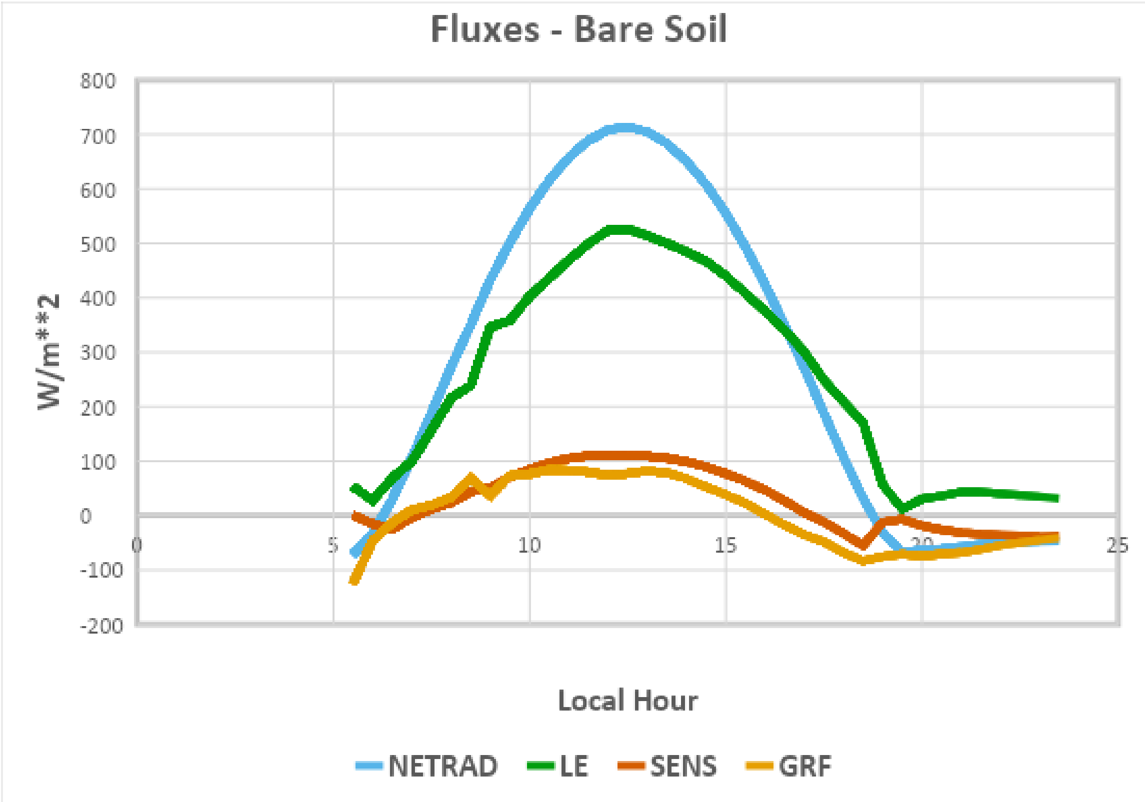
Level 1

Simulation series 1

- **Run simulation using bare soil parameters**

Let us take things one step at a time and consider a mid-latitude site during August at about latitude 40° N (specifically 39.25 N latitude in Kansas). The object is to produce a simulation for the case of bare soil and clear skies, so use the file i_model_bare_soil.json as the input file which sets the vegetation fraction (frveg) to 0.0 and index_soils to 1. Important output parameters to concentrate on are the sensible and latent heat fluxes and the net radiation. We will also look at the surface radiant and air temperatures, which are at 'screen'-level elevation (1.5 m) or at 50 m elevation above the ground. As we have said, the surface is somewhat of an abstraction. However, all elevations are with respect to local ground level. Temperatures, however, are expressed as potential temperatures so that we can compare the values at different elevations. The exception to this rule is the radiant surface temperature and the ground temperatures, which remain as an ordinary temperature. However, differences in potential temperature between the surface and 10 meters above the surface are considered as negligible.

Run a simulation first using conditions provided in the i_model_bare_soil.json input data file. Plot versus time the surface air temperature (say at 50 or 10 m), the sensible and latent heat fluxes, and the net radiation. For example, flux and temperature graphs might look similar to the following:



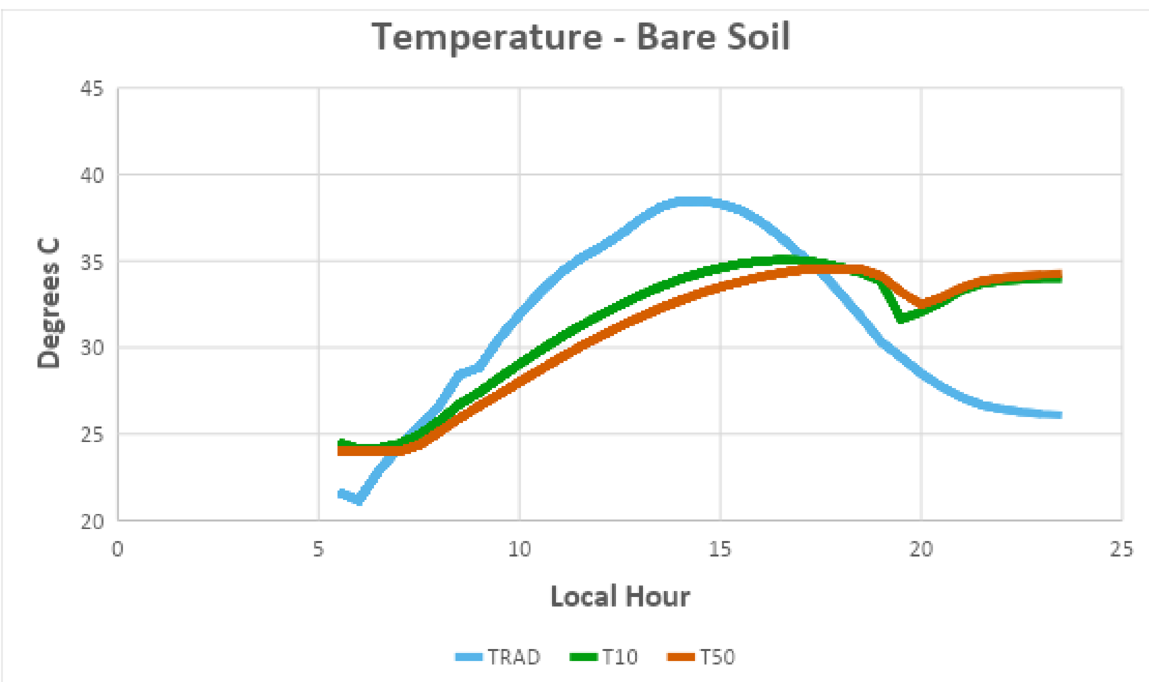


Figure 1.3. Graphs of fluxes (top) and temperatures (bottom) using output from a bare soil run of the Simsphere model using default parameters.

Not surprisingly, the maximum fluxes occur close to local noon, with the latent heat flux slightly lagging the net radiation in the afternoon. Note that the surface air temperature lags still further, reaching a maximum after 1300 h. Note further that the heat flux into the ground (ground flux) reaches a maximum during the morning hours and actually becomes negative (upward) during the late afternoon. Sensible heat flux becomes negative during the evening as latent heat flux almost vanishes.

Questions

1. Recalling our reference to fluxes being driven by changes in a quantity such as temperature, can you provide a physical explanation for the negative sensible heat flux? (Hint: sensible heat flux tends to be directed down the temperature gradient, from higher to lower temperatures.)
2. Why should the air temperature lag the radiant surface temperature? (Hint: recall that the ocean temperatures tend to lag the land temperatures. Could this be due to a difference in heat storage?)
3. Realizing that the heat flux into the ground is also proportional to the vertical gradient of temperature just below the surface, can you explain why the ground flux is a maximum in the morning? What do you suppose is happening to the vertical temperature gradient in the ground with time and why? (Hint: Plot the temperature of the ground (one level below the surface) and the radiant surface temperature.)
4. How much more solar radiation would the south facing side of a gable roof located at 30 degrees latitude receive than the north facing side if the roof slopes 40 degrees? What would the optimum slope be in winter to receive the maximum solar flux?

Simulation series 2

- **Change latitude and/or date**

Now make some additional simulations changing only the latitude (to 20° and then to 60° N) and the date from August to October.

Questions

1. Compare the lower and higher latitude fluxes by plotting both on the same graph. What are the main differences between these various simulations in the fluxes and in the surface temperature and why? (Small differences in the simulations may be puzzling, such as the times at which fluxes maximize. You may discover some interesting features in the fluxes and temperatures that we did not anticipate. Not everything can be explained readily, however.) Hint: Review the section above on Terrain Slope, Solar Elevation, and Azimuth.

Simulation series 3

- **Run simulation with a slope of 20% (11.3°), comparing west-, south-, and east-facing slopes with the default (zero slope).**

Return to the default conditions for simulation #1. Now examine the slope effect with the model by varying the terrain slope and azimuth. We do this by specifying the azimuth and elevation angles for a hypothetical area of ground surface surrounding the point at which the computations apply. The imaginary surface area may tilt upward toward the south (azimuth zero) at a slope of 20% (11.3°). By comparison, it is rare to encounter a road with a slope greater than 15% (8.5°). Compare this result with west-facing (azimuth +90°) and east-facing (azimuth -90°) slopes. Note the diurnal variation in surface temperature, surface energy fluxes and net radiation. Note how the relevant model parameters behave in response to differing slope angles, latitude and time of year. Note the early morning and late afternoon fluxes and temperature and compare with the control run. Do you find anything surprising here?

Questions

1. At what time is the solar flux and the net radiation flux maximized for the different azimuth angles? How do radiation sunset and solar sunset vary with azimuth angle?
2. At what times of day is the solar flux stronger for the north-facing (azimuth 180°) versus south-facing (azimuth 0°) slopes? How about the east-facing versus the west-facing slopes. Can you tell if these differences are reflected in the sensible and latent heat fluxes and the air temperatures?

Level 2

Feel free to experiment with more than just these simulations. Learn to identify the important and relevant differences between the default run and the other simulations and to recognize the significant implications of these results, which may even be counter intuitive.

1. Consider the optimal slope, direction and angle to build a ski resort. What combination of factors would allow the least direct and total solar radiation on the slopes for a given latitude? Assuming that more solar radiation reaching the slope results in more snow blowing costs, would it be cheaper to have a resort in a south-facing slope at 50° latitude versus a west-facing slope at 40° latitude? Are there advantages/disadvantages to having a northeast- versus a northwest-facing slope? You might also want to consider the air temperature differences using a typical winter sounding for each location..
2. You are an architect building a new home in Buffalo, NY. You decide to include some windows on the north side of the roof and you want sunlight to strike these windows as much as possible since they are above the kitchen. If these windows were placed on a 20 degree slope (that of the roof), how much net radiation would strike the windows on this side of the house? Let's say that you also want a fairly steep roof because of the high snowfall typical for Buffalo (approximately 45 degrees north latitude). You could make the angle of the roof greater than 20 degrees (say 50 degrees). What would be a good angle to choose in a new home that might constitute a good compromise to allow the

maximum sunlight on the windows while maintaining a sufficient roof slope to allow snow to slide off in the winter?

Equations

[Eqn 1.1]

$$R_n = L\downarrow - L\uparrow + S\downarrow(1 - \alpha) = L_e E + H + G$$

[Eqn 1.2]

$$\text{Flux of sensible heat proportional to } \frac{\Delta T}{r_H}$$

Terms introduced in this chapter

Terms to look up and remember
Albedo
Bowen Ratio
Conductance
Conductivity
Emissivity
Flux
Gradient
Irradiance
Kirchoff's Law
Latent heat flux
Plank's Law
Radiance
Resistance
Sensible heat flux
Wein's Law

Recommended Reading

- Oke, T., *Boundary Layer Climates*, 1987, Methuen & Co., pp 20-51.
- Sellers, W.D., *Physical Climatology*, 1965, The University of Chicago Press, Chapter 8.
- Monteith, J.L., *Principles of Environmental Physics*, 1975, pp 10-13.

Simsphere Workbook: Chapter 2

Wind Speed in Simsphere

Introduction

A well-known radical, anti-war organization during the 1960's was known as the "Weathermen," so called because it claimed to know which way the wind was blowing; (it evidently didn't know which way the wind was blowing enough to have the foresight to call themselves Weatherpersons.) In this session we will endeavor to become more astute Weatherpersons (whether you like it or not!) by finding out what parameters affect the wind speed and how wind speed influences the surface energy fluxes and the surface temperature. Obviously, the initialized wind speeds themselves govern the wind speeds, although the latter can evolve during the simulation. The purpose of this chapter, therefore, is to explore the effects of wind speed on surface radiant and air temperatures and the surface energy fluxes.

Background

First, ask yourself the following questions: Would you expect the sensible heat flux (and, for that matter, the latent heat flux) to increase or decrease with increasing wind speed? How about the surface temperature? To answer this question, first consider the concept of atmospheric resistance. Away from hard surfaces (where molecular processes dominate the flux of heat across the air interface directly in contact with that surface), the atmospheric resistance is related to the efficiency of turbulent eddies to carry away the heat and water vapor; refer to Eqn. 2.2 and 2.3. Would you expect the turbulence to increase or decrease with increasing wind speed? Would this suggest an increase or decrease in the atmospheric resistance and so, in the sensible heat flux with increasing wind speed? How about latent heat flux? However, wouldn't an increase in wind speed also change the vertical temperature profile and decrease the vertical temperature differences between any pair of levels (ΔT), thereby tending to lower the sensible heat flux? If ΔT decreases, would this occur with a decrease in both the surface temperature and the temperature at some level above the surface, or would just one of the other levels change temperature?

Answers to these questions may not be obvious or even determinable without performing simulations. The basic question we will pose in the scenario is: would you expect the surface temperatures to increase or decrease in response to change in wind speed?

Now you see the problem. You can possibly perceive the answers to these questions, but mathematically (using the resistance formula given by Eqn. 2.2) you are unable to prove your conclusions without further analysis. Now consider the further dilemma: the energy for the sensible heat and for evaporation must come ultimately from the net radiation. If we somehow increase the sensible heat flux by manipulation of the wind speed (or some other parameter), wouldn't we also likely change the latent heat flux in the same sense? If so, where does that extra energy come from, or are both fluxes somehow constrained by a mutual competition for the same available radiant energy (Eqn. 2.1)? (Later chapters will ask how the availability of substrate water influences these parameters.)

Procedure

Let's make some simulations and observe the behavior of the surface energy fluxes, the surface temperature and the Bowen ratio. (Before going on you might wish to consult Oke's book on Boundary Layer Climates, Second Edition, pp 32-42 and Monteith's book Principles of Environmental Physics, Second Edition.) In the simulations that follow, you will be examining the role of the wind speed in modifying the surface fluxes of heat and water vapor as well as the surface and near-surface air temperatures.

As you make the following simulations, you will notice that the surface sensible heat flux and the surface temperature are related. This is not surprising in view of Eqn. 2.2 and its allied resistance laws. We find, however, that the surface sensible heat flux maximizes within an hour or so of midday, but the surface temperature maximizes in the mid-afternoon, as is consistent with our own observations and the diurnal temperature lag described in Chapter 1. What about the vertical temperature gradient when the wind speed increases or decreases? Would the effect of changing the wind speed change the resistance and the vertical temperature (or moisture) gradients so as to partially counteract the other's effect on the fluxes? After all, Eqn. 2.2 tells us that the sensible heat flux depends not on the temperature but on the vertical temperature gradient (as expressed by the difference between the surface temperature and that at the top of the surface layer). We explore this aspect further in Scenario 5.

Surface Wind speed, Temperature, Specific Humidity and other Constituents in the Atmospheric Surface Layer

First, what do we mean by a 'surface layer'? Specialists in the atmospheric boundary layer may define the surface layer as the depth above the surface over which the vertical variations in the fluxes of heat, moisture and momentum do not vary by more than about 10%. Somewhat erroneously, this layer is called the layer of 'constant flux' . This term is not meant to imply that the fluxes do not change with height, but that they change in a small way (small being defined as less than 10% of the value at the surface). The need to define a surface layer is not as arbitrary and whimsical as it may seem. Whether we are talking about the atmospheric surface layer, the airflow over my desk or the flow of blood within the walls of an artery, the variation of fluid speed versus distance in the surface layer has a characteristic logarithmic variation, such that the greatest variation with distance occurring at the surface interface. Viewed graphically, this type of wind profile would show a decay of speed toward the surface. Such a logarithmic relationship results in a straight line when the speed is plotted versus the log of the distance from the surface. Figure 2.1 shows a series of data points for two such wind speed plots, one over a rough surface and one over a smooth surface.

Typically, the depth of the surface layer is about 50 m during the day and perhaps somewhat less at night. Accordingly, in our model we will take the surface layer to be 50 m in depth. Temperature, specific humidity, momentum, carbon dioxide and ozone conform to this logarithmic profile over which the fluxes themselves remain constant. Because a logarithmic quantity can not have zero value, the wind speed variation shown in Figure 2.1 has a non-zero value of height at zero wind speed. The height at which the wind speed vanishes is called the roughness height, which is assigned the symbol z_o. The smaller the slope of the straight line the larger the roughness length and the greater the associated turbulence. Indeed, the slope of the straight line is inversely proportional to a parameter known as the friction velocity, which is a direct measure of the turbulence.

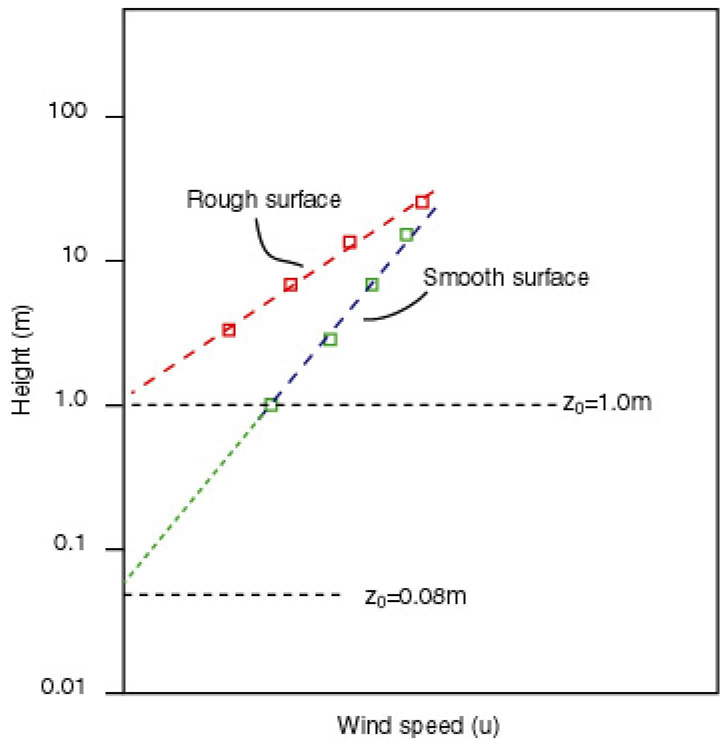


Figure 2.1. Wind Speed vs. Height.

Roughness heights must be assigned in the model. A general rule of thumb, which holds well for uniform vegetated surfaces, is that the roughness height is about one-tenth the average height of the roughness elements. This rule is more problematic for irregularly distributed roughness elements such as in a city. If buildings are typically about three stories high (10 m), a reasonable roughness height could be 1 m. The concept is not very clear when talking about isolated roughness elements such as trees in a field. A single tree (or two trees) 10 m high sitting isolated in a field generally will not influence the roughness height for air flowing over that field. In that case a suitable roughness height might be that appropriate to the shorter vegetation in the field, say 0.1 m. As one adds trees to this picture, the roughness height will gradually become dominated by the trees, so that a forest of trees 10 m high might have a roughness of 1 m. Beyond a certain density, when turbulent eddies in the atmosphere can no longer penetrate to the surface between the trees, the roughness may decrease as the tree canopy becomes denser. In this case the wind profile over a dense canopy of trees may respond as if the actual surface were near tree top level and the roughness height would be appropriate to a smooth layer formed by an ensemble of tree crowns. Thus, care must be taken in assigning a roughness height in a particular problem.

Other variables than wind speed, such as temperature (which is associated with sensible heat flux), specific humidity (which is associated with latent heat flux), carbon dioxide and ozone also obey logarithmic profile laws, such that the change in these variables with height increases in an exponential manner as one approaches the surface. A rule is that when a surface interface acts as a source for the flux the associated variable is a maximum at the surface. When the surface acts as a sink for the flux the associated variable is a minimum at the surface. Thus, in the case of temperature, the surface constitutes a source of sensible heat flux during the day and exhibits a temperature maximum; a temperature minimum occurs at night under conditions of downward sensible heat flux. Momentum flux always has a sink at the surface where the wind speed (at the roughness height) is always zero.

The mathematical surface - that level which the logarithmic functions perceive as the surface -- need not be identical to the physical surface that you and I perceive as the ground. The height of the mathematical surface may lie some distance above the ground surface. This level is referred to as the *displacement plane*. In the case of a dense tree canopy, the wind speed may behave as if the tops of the trees constituted a smooth surface. The actual ground surface beneath the trees might be meters below the mathematically defined displacement height. In our model, all heights are taken with respect to the displacement plane. Although the user must decide the actual height of the displacement plane above the physical surface, we need not specify the displacement height. All heights in the model are defined with respect to the displacement height. Typically, the displacement height is about two-thirds of the average height of the obstacles except when the obstacles are irregularly spaced, close together, or quite far apart. For example, in the case of the dense tree canopy, the displacement plane might be close to tree top level rather than at two-thirds the height of the trees.

Simulations

Level 1

Simulation series 1

- **Change wind speeds at all levels**

Let us first consider the role of the wind speed itself. After running the default case, make another run after decreasing the wind speeds by, say, half **at all levels** in the wind speed profile. Watch and explain what happens to the fluxes and the air temperatures in comparison to the default case. Check the behavior of the ground fluxes and the maximum temperature. Is this behavior consistent with the flux and resistance laws of Eqn. 2.2 and 2.3? Does the net radiation increase or decrease as the wind speed slows down, and what conditions would lead to this result? (For the answer to this, look at the radiant surface temperature and ask yourself what effect that parameter has in the net radiant flux at the surface.)

Look closely at the sensible heat flux, and you should see small fluctuations in this parameter and in the wind speed at 50 m during the evening, which die down when the wind speeds decrease. Note that the evaporation vanishes at night no matter what the wind speed.

- **Geostrophic wind and the balance of forces**

The geostrophic wind speed is simply a measure of the pressure gradient and therefore the pressure gradient force. Since we specify the surface geostrophic wind we are effectively specifying the surface pressure gradient. Normally on the large scale and away from the surface, where friction is very important, the pressure gradient force is closely balanced by the Coriolis force, which is proportional to the wind speed. So when the wind speed is large, the pressure gradient force (geostrophic wind speed) must be large in order to balance it. If balance does not occur, however, there is a net force and a net force requires an acceleration or deceleration of the wind speed.

In the model, we assume that, initially, the Coriolis and pressure gradient forces are balanced (geostrophic balance) at all levels above 550 m from the surface. This assumption is not made at the surface. Instead, we specify the surface pressure gradient force to be constant from 550 m above the surface to the surface. Whereas the two forces are defined initially in balance at 550 m and above, they are not necessarily equal below this level. Why do we not let the two be equal all the way to the ground? The reason is friction, which causes the winds to decrease with decreasing altitude below about 0.5 – 1.0 km above the ground. (In fact, friction indirectly extends its effect through mixing to greater heights.) Since the wind speed decreases fairly rapidly with lowering height near the surface, the pressure gradient force will likely exceed the Coriolis force below 550 m. If a force balance is maintained it must include the frictional force.

Agreed that this is a bit of an artifice. We make this assumption only to start the model in a reasonable manner. What actually happens once the model begins to execute is that an effective force imbalance is transmitted up through the mixing layer via the effects of turbulent mixing associated with the sensible heat flux. This turbulent mixing, manifested by sensible heat fluxes, can vary in time. The imbalance between Coriolis and pressure gradient forces, which were initially equal above 550 m, is created by a kind of virtual friction produced by the turbulent mixing in which faster and slower moving air mix. This mixing is manifested by faster moving air moving to levels with slower moving air and visa versa. This mixing creates a virtual friction force, which may tend to balance, which is to say that the three forces, Coriolis, pressure gradient and friction may be closely in balance. However, close is not exact, and the effect of mixing is inevitably to create a resultant acceleration or deceleration. These accelerations can be large at times, causing the wind to vary rapidly with time. If the imbalance between the forces is excessive, large oscillations in the wind field can develop as the result of the component accelerations of the wind. In reality, such oscillations may be associated with the development of the nocturnal, low-level (1000 - 1500 m) jet. However, if you impose something unreasonable on the wind field, the simulation may decide to self-destruct. It is the model's way of saying 'Ouch p-p-please give me something reasonable!')

Simulation series 2

- **Night time winds**

Re-run the default simulation, reducing the wind speeds again by a factor of two. Look at the night-time winds and temperatures. (If nothing dramatic happens, decrease the wind speeds again uniformly by a factor of two. Observe the fluxes again but pay particular attention to the winds and surface temperatures at night. Do you conclude that a calm night causes more cooling? How does the change in surface air and surface radiant temperatures compare between the default and light wind cases during the day and at night.

You might see the surface air temperature at 1.5 m decrease when the wind speed is decreased. Intuitively, one would expect a surface to cool when the wind speed is increased and visa versa. Blow on your hand and you will see that this is true. So why might the 1.5 m temperature be higher in the higher wind speed case? Does this apply also to the surface radiant temperature? Hint: what has happened to the vertical temperature difference from 1.5 m to the surface when the wind speed is increased or decreased?

Level 2

Simulation series 3

- Sometimes the temperature in the State College, PA 'Barrens', a wilderness region with scrubby vegetation and poor heat conducting soils, cools to sub-freezing temperatures at the surface in August. See if you can simulate, using the State College sounding, a freezing night during summertime by reducing the wind speeds in the initial conditions.

What might you expect under mixed conditions; that of a cloudy night with strong winds or a clear night with calm winds? You can make this simulation by imposing a 'cloud fraction' of say 0.5. If you were a fruit grower, what types of weather conditions could conceivably threaten your crop by night time freezing temperatures?

At nightfall, the virtual frictional force discussed above vanishes with the disappearance of mixing provided by sensible heat flux. What effect might this sudden change in the force balance have on the winds? Do you see any hint that a profound acceleration is occurring in the wind speeds above the lowest 0.5 km? (Hint: look at the evolution of the vertical profiles of the wind in the animation.)

Simulation series 4:

- Make a pair of runs for roughness heights of, say, 0.5 m and 0.1 m and note the differences in the wind speed at 10 m for both situations and at the same time of day.

Suppose that we wanted to know what is the wind speed at face level (1.5 m) for these two simulations, but all we had was the wind speed at 10 m. Plot the 10 m wind speed on a piece of semi log paper in which the x-axis is labelled wind speed and the y-axis (the log axis) is labelled height. Label the height axis at 0.1, 1 and 10 m. Knowing that the wind speed is zero at the roughness height, connect the two points by a straight line from 10 m for each simulation and determine the wind speed at face level for these two roughness cases.

Equations

[Eqn 2.1]

$$R_{net} = S\downarrow(1 - \alpha_0) + L\downarrow - L\uparrow = L_eE + H + G$$

[Eqn 2.2]

H (surface sensible heat flux) is proportional to $\frac{\Delta T}{r_H}$, and

[Eqn 2.3]

r_H is proportional to $\frac{1}{u}$, where u is the wind speed and ΔT is the vertical temperature difference

Terms

Terms to look up and remember
Roughness length
Displacement height or displacement plane
Friction velocity

Simsphere Workbook: Chapter 3

Soil Moisture in Simsphere

Introduction

So far, the simulations are run with the same initial *soil water content*. We are about to explore what happens when we change this very important parameter. First, however, let us start with some elementary definitions of what we mean by soil water content. A familiar term, which is extensively used in the literature, particularly with regard to climate simulations, is the '*soil moisture availability*'. We define this term in two seemingly independent ways; (1) in the soil as the fraction of '*field capacity*', and assign it the symbol M. The subscript "s" refers to a surface layer of some depth measured from the surface (*e.g.*, 10 cm), and the subscript "o" to a skin surface layer. Others may define moisture availability differently, for example as the fraction of 'extractable' soil water content, where some residual amount of soil water (e.g. 0.05 by volume) is considered to be un-extractable or as the fraction of saturation. We also define this term with regard to the evaporation flux. We will return to the difference between these two definitions and to a further discussion of the distinction between field capacity and saturation.

An alternate term used to represent soil moisture is the *volumetric soil water content*, w, which is the volume of soil water content per unit volume of soil. Sometimes the volumetric water content is expressed as a water depth (*e.g.*, 10 cm), whereas that depth only has meaning when the depth of the column is also understood, *e.g.*, 100 cm. Thus, for a unit surface area, the volumetric soil water content is equal to the equivalent depth of water, *e.g.* ten centimeters of water in a column of 1 meter corresponds to a volumetric fraction of 0.10. Think of this as a cylindrical tube one-tenth filled with water. You may also see the term '*gravimetric soil water content*' reported in the literature; this is the mass of water per unit mass of dry soil. The volumetric and gravimetric definitions are equivalent except that the latter is multiplied equal to the former multiplied by the soil density. Various other measures of soil water content exist, such as the soil water tension, *soil matric potential* and the *potential head*, all of which have the units of pressure (Pa or bars) or length. For further details on soil water content units, we refer you to D. Hillel's excellent book, **Introduction to Soil Physics**, pp. 8-12.

Another parameter related to the soil water content is the *hydraulic pressure*, otherwise called the *soil tension or water potential*. Hydraulic pressure is a difficult concept which is more relevant to a later part of this course; for the present, simply regard this parameter, which is customarily assigned the symbol Ψ, as a measure of the soil water content.

Let us first see how soil water content is used in the model. A related expression to w is the *atmospheric moisture availability*, M_a. To explain what we mean by this term, we present Eqn. 3.1, which is simply a variation of the resistance law, as expressed in Eqn. 2.2, but for latent heat flux. Here, the potential drop is the *specific humidity* (q), the vertical difference Δq across an atmospheric layer in the vertical governed by an atmospheric resistance, which is the same as referred to in Eqn. 2.2. (Instead of specific humidity, we could alternately use *mixing ratio* r or *water vapor pressure* carbon dioxide concentration, depending on what flux we are talking about).

For a flux within the atmospheric surface layer, the *layer of 'constant flux'*, we can write the specific humidity difference as that between the surface and the top of the surface layer (or any other pair of levels in that surface layer). This applies so long as the resistance in the formula pertains to that same vertical interval, *e.g.* the resistance of the atmosphere to the vertical transport of sensible or latent heat from point a to point b would be $r_{a,b}$. Now, it's no trick to measure the specific humidity at the top of the surface layer, or indeed, at any level above the ground surface. The problem, however, is to specify (let alone measure) the specific humidity at the bottom of our imaginary vertical column, *i.e.* at the ground surface, which is hypothetically in the air directly touching the surface. Of course we cpu;d try to calculate the flux of water vapor by specifying specific humidity q at two ends of an elevated vertical column, *i.e.* at two levels above the ground, but there are computational problems in doing so, such as having to measure the humidity at the two levels.

Before, proceeding, let us define the atmospheric *surface layer*. Theoreticians like to define the surface layer as a layer of constant flux, which means that the vertical flux is the same at the bottom of the layer as at the top. Now, this is obviously a contradiction in terms, as the warming or cooling of the atmosphere (or the drying or moistening of it) is dependent on a difference in sensible heat flux vertically between the top and bottom of the column. However, if we take that column to be short enough the vertical difference in flux values will be small. What is small? Normally in science, one considers a quantity small if it is an order of magnitude (factor of 10) smaller than the numbers to which that small value is compared. Thus, the surface layer is defined as the layer over which the fluxes do not change in the vertical by more than 10%. Nominally, we take this surface layer to have a depth of 50 m, although in reality that depth will vary with time and likely be smaller at night than during the day. For computational purposes, we will henceforth take the surface layer depth as 50 m.

In considering moisture availability let us first tie the flux of moisture in the air just above the surface to the amount of substrate water and the specific humidity at the ground surface which is at the level in the atmosphere most closely tied to the soil water content. Second, to make the calculation of vapor flux we need to calculate or specify a vertical difference in specific humidity in the surface layer. We could choose one level slightly above the surface and one at the surface. If so, we are faced with the problem that the bottom layer of the column is indeterminate in that we are unsure exactly where the surface lies, and there is no equivalent surface specific humidity corresponding to the surface radiant temperature, which can be more easily measured than specific humidity at the surface. If we choose to measure the specific humidity at the bottom of the column but at a level just above the surface, we would still need to relate the specific humidity at that level to a value at the soil surface. Alternately stated, we would have trouble closing the system mathematically.

Having chosen q_a , the specific humidity at the top of the surface layer, we are still faced with the problem of determining an imaginary value of q_0 (specific humidity at the earth's surface) for the 'bottom' of the layer (Eqn. 3.2). To do this, we resort to an artefact, which is as follows: Let us define a *'saturation' specific humidity* q_S at the surface temperature (T_0); the fractional *relative* *humidity* at the surface h_0 is thus approximately[1] equal to $\frac{q_0}{q_S(T_0)}$ allowing us to rewrite Eqn. 3.2 as Eqn. 3.3. This type of formulation, although used by some who model, still leaves us with a highly variable and difficult to prescribe unknown parameter, that of h_0, which is still dependent on q_0.

We must admit that the problem of linking q_0 or h_0 to the substrate water content has not been rigorously solved, although some expedients for circumventing the whole issue exist. No matter, let us proceed by defining a 'potential evaporation' $L_E E_{not}$. There are various ways to define this quantity. Ours is, simply, the evaporation occurring over a saturated surface of temperature T_0, using Eqn. 3.2, for 100% relative humidity at the surface $h_0 = 1.0$. This is Eqn. 3.4. (Other definitions of potential evaporation are possible). We now define the quantity, M_a, the '*atmospheric moisture availability*', which is simply the ratio of actual evaporation to potential evaporation. Thus we can write Eqn. 3.1 as Eqn. 3.5.

Intuitively, M_a must be related to soil water content. The leap of faith which we require in order to proceed is to set M_a equal to M_0 , where M_0 is the soil water content as a fraction of field capacity in the soil surface layer. Because there is very little else one can do, this equivalence between M_a and M_0 is adopted by many who model the land surface, although some attempt to demonstrate erudition by using a power law rather than a simple ratio for soil water content. There is some justification for such an assumption based on field measurements, but the exact form of the relationship between M_a and M_0 undoubtedly varies with soil type, amongst other things. Nevertheless, the evaporation will vary from zero for $M_a = 0$ (perfectly dry soil) to potential evaporation for perfectly wet soil. These two somewhat related variables both vary between 0 and 1.0, probably in some non-linear fashion with each other. When the soil is absolutely dry both variables will have the value of zero. When at field capacity, both M_a and M_0 will equal 1.0. Simply equating one to the other, as we do in the model, may not be quite correct, but the differences can never be great, as both are constrained to vary in more or less the same way in between the same limits. So let us proclaim the equivalence close enough for mathematically modelling the land surface.

Eqn. 3.5 ties evaporation closely to soil water content. By setting $M_a = 0$ (totally dry soil) the evaporation is completely suppressed. For M_a not equal to zero, however, a decrease in M_a, while certainly causing a decrease in evaporation, also corresponds to an increase in sensible heat flux and, therefore, according to Eqn. 2.1, to a rise in surface temperature. However, a rise in surface temperature corresponds to an increase in the saturation specific humidity at the surface $q_S(T_0)$ and so to an increase in the vertical gradient in q and therefore in potential evaporation. Thus, a compensation (negative feedback) exists which attempts to brake the decrease in evaporation caused by the decrease in soil water content. The result is that a 10% decrease in M_a may correspond *in some instances* to far less than a 10% decrease in evaporation and in other instances to a greater than 10% decrease. Clearly, since a 100% change in M_a causes a 100% change in evaporation, there must be ranges of M_a in which evaporation changes less rapidly with changing M_a, and ranges where evaporation changes more rapidly than an equivalent change in M_a; n'est-ce pas ?

Before proceeding to the simulations, we should more precisely define field capacity and soil saturation. Here is an experiment for you. Take a flower pot, fill it with water until it starts to drip (but be careful here). When the dripping ceases we can say that the water content in the soil is in equilibrium with gravity and is at field capacity. If we were to add water beyond field capacity until all the pores between the soil grains were filled with water the soil would then be at saturation. At this point, the soil would really look like mud. A measure of saturation for a given soil sample is its *porosity* which is the percent of the soil occupied by spaces between grains. A rule of thumb is that field capacity tends to be about 0.75 of saturation, although both field capacity and saturation vary with soil type to the extent that the ratio between them is not really a constant: 0.75. Typically, a soil may have a saturation value of about 0.44 by volume and a field capacity of 0.33 by volume.

Field capacity and other soil parameters are set in the model by the choice of soil type. We will not be concerned with soil properties except for vegetated surfaces, as the bare soil evaporation does not make use of soil properties explicitly. This seemingly oversight on our part is simply a convenience.

Most models of surface energy fluxes use field capacity rather than saturation to define moisture availability. There are two reasons for this choice. One is that saturation in natural soils is fleeting after a rainstorm. Very quickly, the excess water beyond field capacity diffuses downward, or runs off or flows directly to deeper layers via cracks in the substrate, thereby reducing the soil water content to field capacity. The main reason, however, is that the fluxes from saturated soil are virtually identical to that over soil at field capacity, and therefore there is virtually no sensitivity of the surface fluxes to soil water content above field capacity. In some models, the value of zero soil moisture availability is taken at a low value, loosely identified with the so-called '*wilting point*'. We do not adopt this as the lower limit (preferring instead that zero be the lower limit for soil water content) because we believe that evaporation can continue even when the soil is below wilting point. Wilting occurs only for plants and not for soils.

Simulations

Level 1

In this scenario we give you carte blanche to fiddle with the surface moisture availability as you so desire, however you might like to ponder the following in making your simulations. Note that we do not expect you to do everything suggested; simply take a facet and explore it as you develop your understanding of the concept of moisture availability!

Simulation series 1

- **Vary Surface Moisture Availability**

In these series of simulations we would like to see how the fluxes and surface air and surface radiant temperatures fluctuate when soil water content is changed. We are interested in the relative differences between the individual simulations. Two values of moisture availability, one for the soil surface (M_o) and one for the *root zone* (M_r), are required to make a simulation. Until now both have been set, the former at a value of 0.5 (half of field capacity) and the latter at 0.75.

After making a run with the standard parameters used in chapter 1, reduce M_o (parameter "F" in the model input file) to 0.1, and then increase it to 0.9.

Look at the changes in sensible and latent heat fluxes, as well as the surface temperature and surface moisture availability itself.

Explore the sensitivity of evaporation and of surface temperature T_0 to changes in M_o. (We define '*sensitivity*' of evaporation to changes in M_o as $\frac{\Delta L_e E}{\Delta M_o}$.)

Questions:

1. Based on these two simulations, in what part of that range of Mo is the sensitivity of evaporation and of surface temperature to soil moisture the greatest? Where are those sensitivities the smallest? Are those sensitivities constant throughout the day?
2. How does the *Bowen Ratio* change with changing M_o ? (Recall that the Bowen Ratio is the ratio of sensible to latent heat flux at the surface.)
3. How does M_o itself vary throughout the day and why should it vary in time?
4. How does *ground flux* v(the flux from the surface up or down through the layer of soil just below the surface) vary with changing M_o? It should be evident to you by now that the moisture availability profoundly affects the Bowen Ratio and the individual fluxes.

Simulation series 2

- **Change the windspeed and moisture availability**

Let's see how surface fluxes are affected by a changing M_o at different wind speeds. In other words, we want to see how the sensitivities of the fluxes are determined in the above simulation change with wind speed? In so doing, lets examine the variation of soil water content itself with time during the day.

Vary the wind speed just as you did in Simulation 1 of Chapter 2, but perform the simulations with $M_o = 0.1$. Repeat these simulations using a $M_o = 0.9$. To keep things straight, it may help to make a table or chart listing the varying parameters of these runs and the features seen in the output.

Questions:

1. You may have noted from the previous scenario that wind speed affects the Bowen ratio. What we would like to investigate is whether that change in Bowen

ratio due to changing wind speed is likely to be in the same sense for all values of M_o. Alternately stated, does a change in surface energy fluxes brought about by a change in wind speed equally affect sensible and latent heat fluxes at all values of M_o?
2. We know that decreasing the wind speed causes the atmospheric resistance to go up and the fluxes to go down. (You can see this effect in reverse by blowing on wet skin or on hot skin caused by a burn.) Does this occur in the atmosphere? Obviously, were the soil absolutely dry, the change in surface latent heat flux due to a change in wind speed would be zero, whereas this would not be the case were the soil wet. The goal here is to assess at which range of soil water content the wind speed plays the largest role in determining fluxes.
3. What happens with the surface moisture availability versus time during the day?

Simulation series 3

- **Change the root zone moisture availability**

Root zone moisture availability will be very important when plants are introduced but will have a small effect in the case of bare soil.

Change the root zone moisture availability M_r (parameter "FSUB" in the model input file) to 0.1 and then increase this parameter to 0.9.

Questions:

1. Is there any effect on the fluxes or temperatures? (Recall that we are dealing only with bare soil; there is no vegetation.)
2. What happens to the surface soil water (water content in the top few millimeters of soil represented by M_o (SMA in the output)) during the day? Why does it decrease? Would it decrease if the two moisture availability values (root zone and surface) were set equal initially?
3. Why does M_o decrease during the day even when the root zone soil water content is high? What happens to M_o at night? Can you explain why the latter tends to increase at night when the root zone soil water content is high?

The above are all questions you should be able to answer without much knowledge of soil physics. The key here is that two factors affect the soil water content in our simple two-layer soil scheme, evaporation and diffusion of soil water vertically within the substrate. Water is lost from the surface layer during evaporation (but not directly from the root zone soil water content) but it can also flow down the gradient of soil water content, even at night, within the soil. Later we will see that transpiration, the evaporation from leaves, is taken by the roots from water in the root zone layer.

Finally, a suggestion: Don't perform more simulations than you are able to comprehend at this moment. If the results seem bewildering, stop and think about them. Don't worry about covering all bases.

Level 2

As shown in these simulations, soil water content is a very crucial parameter in simulating surface energy fluxes and temperatures. However, obtaining an accurate estimation of soil water content is very difficult to obtain either directly or empirically to insert into a model. On the other hand, it is relatively easy to specify parameters such as season, latitude, slope, azimuth, and wind speed in these models, as these can be accurately measured.

Qualitatively, compare the effects of varying soil water content with those of varying latitude, slope, azimuth, wind speed, and time of the year on surface fluxes and temperatures.

Look back at the results from previous simulations in Chapters 1 and 2. Try ranking these variables in order from most important to least important in determining the surface energy balance. Note that the ranking may vary depending on the range you specify. For example, time of year may not matter between May and September, but it could be crucial between September and December.

This should provide you with a good understanding of how the factors we have looked at so far may affect future simulations, *all other things being equal*. It is important to understand the uncertainty of various inputs into a model along with their potential influence on the output. In this case, soil water content is a relatively uncertain parameter, yet its potential to influence the simulation is very high.

To demonstrate this, the following simulations examine the effects of varying the surface moisture availability (F in the model) and the root zone moisture availability (FSUB in the model) on several lawn-watering scenarios: 1) A dry root zone with a heavily watered surface at daybreak, 2) A dry root zone and dry surface (a non-watering case), 3) Repeat #1 but for a cloudy day, 4) Repeat #1 but for nighttime hours only. For each case, observe how the surface (SMA) and root zone (RZMA) moisture availability change during the run:

- Perform a simulation with standard conditions with the surface moisture availability at field capacity (M_o = 1.0) and the root zone moisture availability at a low value, say 0.1, representative of a heavily watered surface with a very thirsty root zone layer. This will simulate the watering case. Start at the standard early morning time and run until 5 AM the following day (set timend equal to 2900 in the model input file which is 2900-2400=500 AM of the next day).
- Perform the simulation with a low value of soil water content at the surface (M_o= 0.1) also starting at the early morning time. This simulates the non-watering case.
- Run a third simulation with a cloudy sky (set cloud_flag to “true” and cld_fract equal to 7 tenths in the model input file) for the watered case.
- Finally, start the simulation at 5 PM (strtim=1700 in the model input file) and let it run until 5 AM the next morning for the watered case.

Questions:

1. In the first three simulations, can you assess qualitatively how much of the surface water reaches the root zone and how much is "lost" to evaporation by the end of the day.
2. What is the difference in the root zone moisture availability and surface moisture availability at 3 AM (27 hours) in the first three cases?
3. If the day were cloudy, would it matter so much if the lawn were watered in the morning?
4. What is the difference in water efficiency (root zone storage versus water lost to evaporation) after 12 hours of simulation for the case when the lawn is watered in the morning versus when it is watered in the evening?
5. What do the experiments suggest regarding the efficacy of watering at the start of the day, the end of the day, or whether watering is needed at all?

Equations

[Eqn 3.1]

$$L_eE = \rho L_e \frac{\Delta q}{r_v}$$

[Eqn 3.2]

$$L_eE = \rho L_e \left[\frac{q_0 - q_a}{r_{oa}}\right]$$

[Eqn 3.3]

$$L_eE = \rho L_e \frac{h_o q_s(T_o) - q_a}{r_{oa}}$$

[Eqn 3.4]

$$L_eE_{pot} = \rho L_e \frac{q_s(T_o) - q_a}{r_{oa}}$$

[Eqn 3.5]

$$L_eE = \frac{M_a\left(q_s(T_o) - q_a\right)}{r_{o,a}};\ M_a = \frac{L_eE}{L_eE_{pot}}$$

Terms

Terms to look up and remember
gravimetric soil water content
mixing ratio
potential evaporation
saturation vapor pressure
sensitivity curve
soil porosity
soil water potential; hydraulic potential; soil tension

soil water saturation
transpiration
vapor pressure
volumetric soil water content
wilting point

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Simsphere Workbook: Chapter 4

Thermal Inertia in Simsphere

Introduction

We all know, or think we know, what inertia is. If something or somebody is reluctant to move in the face of an outside force, we call that inertia. If a body is heated externally, *e.g.*, by the sun, but does not change its temperature very rapidly, that body can be said to have a large thermal inertia. In the absence of all surface fluxes except the sensible heat flux into the atmosphere (H) and the heat flux into the soil (G), the *amplitude of the daily temperature wave* (the maximum and minimum temperatures reached in a 24-hour period) is a crude measure of the thermal inertia. The greater and more rapid the change in daily temperatures, the smaller the thermal inertia of the surface.

An important factor in the ability of an object to change temperature is called *heat capacity*. The quantity of heat required to raise the temperature of a system by a certain amount is a measure of the heat capacity of a system, commonly expressed in units of Joules per kg per degree K. The higher an objects' heat capacity, the longer it will take to increase its temperature given a certain amount of applied heat. The change in temperature of a mass due to an input of Q Joules of heat per kg of mass per second is expressed in Eqn. 4.1, where C' is the heat capacity and DT a temperature change.

Heat capacity is not the whole story behind thermal inertia, however, because it does not tell us how the temperature will rise if heat is applied to a boundary of a system, such as the ground surface for example, because the amount of heat conducted into a material depends on its *thermal conductivity*. If the sun heats a square of ground surface, the rate of heat conducted into the ground depends upon the thermal conductivity and the temperature gradient of the soil just beneath the surface. This is expressed by Eqn. 4.2, where λ' is the soil thermal conductivity, Δz a distance interval in the ground and ΔT a temperature change.

Thermal inertia also differs from heat capacity in the following way: If a body is large and the heat applied externally on the surface, the change in temperature of the material depends not only on the heat capacity and conductivity, but also on the rate at which heat can be diffused from the surface to the interior or from the interior to the surface. Therefore, thermal inertia is dependent on any two of the following three parameters: heat capacity, thermal conductivity and *thermal diffusivity*.

Clearly, how much the substrate temperature rises depends on how rapidly the heat diffuses in the ground, as well as on how rapidly the heat is conducted through the surface into the substrate. The ability of a medium to diffuse heat is given by its thermal diffusivity, κ', which is defined for the soil as $\kappa' = \dfrac{\lambda}{\rho' C'}$ where λ ' is the soil conductivity, C is the heat capacity, and ρ ' is the soil density.

Because the vertical gradient of the heat flux G in a one-dimensional vertical column of soil is proportional to the rate of change with time in heat content Q (in Joules per kg per second) of the material, we can combine Eqn. 4.1 and 4.2 to yield the familiar Fickian diffusion equation, (Eqn. 4.3) which is one of the most useful, if not overworked, equations in science and engineering. In our model, it serves as the basis for computing the rate of change of temperature in the soil. (You might want to consult the passages recommended in one of the four references listed below for a slightly more detailed discussion of the diffusion equation.)

Soils have a wide range of thermal diffusivities, thermal conductivities, and densities. Remember, however, that two of the following three parameters, λ ' and κ' and C', are independent thermal parameters, meaning that any one of these can be figured out from the other two. For convenience, we limit ourselves to a discussion of just thermal diffusivity and thermal conductivity with regards to their impact on thermal inertia.

Theoretical studies show that if one applies a periodic heating at the surface of a semi-infinite medium such as a column of soil, the rate of heat flux into the substrate is proportional to a quantity $\lambda' \kappa'^{-\frac{1}{2}}$, which is also equal to $\dfrac{1}{\sqrt{\lambda' C' \rho'}}$ [Eqn. 4.4]; (also see p. 139 of Sellers's book). This quantity is called the *thermal inertia* **P**, although it has various other names, such as the *conductive capacity* or *thermal property*. A small thermal inertia results in a small influx of heat, G, into the substrate and a larger surface temperature rise. Since the night-time cooling at the surface is partly dependent on the supply of heat from below, a smaller thermal inertia can also result in a greater night-time cooling due to lack of heat storage in deeper layers of the soil.

Thermal inertia is expressed in the model in cgs units. These units are $\mathrm{cal\ cm^{-2}\ K^{-1}\ s^{-\frac{1}{2}}}$. Typical values for most natural soils and urban materials range from 0.02 to 0.06 $\mathrm{cal\ cm^{-2}\ K^{-1}\ s^{-\frac{1}{2}}}$. This is a somewhat ungainly quantity, so an artefact called the Thermal Inertia Unit, (or TIU) is employed. Let us examine the derivation of TIU. First let us arbitrarily take the square root daily period $\omega = \dfrac{2\pi}{86400\ \mathrm{seconds}}$ and divide it into P. The purpose in using a totally irrelevant parameter, ω , the earth's rotation rate, is simply to remove the square root from the time units in the expression for P. The resultant unit is $\mathrm{cal\ cm^{-2}\ s^{-1}}$. Further dividing by 4.184 Joules per calorie and multiplying by 10 000 square cm per square meter yields a much simpler and more familiar set of units in SI notation, 357 W m^{-2} K^{-1} per $\mathrm{cal\ cm^{-2}\ K^{-1}\ s^{-\frac{1}{2}}}$. So 35.7 TIU is equal to 0.1 $\mathrm{cal\ cm^{-2}\ K^{-1}\ s^{-\frac{1}{2}}}$, which is a fairly large value, and 1 TIU is equal to about 0.003

$cal\ cm^{-2}\ K^{-1}\ s^{-\frac{1}{2}}$, which is exceedingly small. Because many old tables express thermal inertia in cgs, we will use this system for expressing thermal inertia in the model, and simply forget how awful the units are! The user, however, can input the thermal inertia in TIU. For most natural surfaces, the value of TIU is between 4 and 30. In real soils, TIU also varies with soil water content. For dry soil, the TIU is about 6 units and for soil at field capacity it is about 28 TIU. Although the model will not permit the user to exceed reasonable bounds for TIU, we recommend that values not be set below 6 or above about 30.

Applications of the Thermal Inertia concept

A theoretical implication of the diffusion equation is that, given an external periodic heating of the surface, the amplitude of surface temperature (twice the difference between the maximum and minimum surface temperature) is related to the type of surface composition. This idea first appealed (during the 1970's) to certain geologists, who were looking for ways to classify land surface properties using the surface infrared temperature, which can be measured from a satellite. (Note that the **Manual of Remote Sensing** lists the thermal inertia for different types of soils and materials.) These geologists attempted to use the measurements of surface temperature to determine the mineral properties of the soil, first on the moon and later on earth. They were later to abandon the idea because they discovered that surface moisture and vegetation scrambled up the measurements in a way that quite limited their usefulness.

A background on the development of the thermal inertia concept appears on the first three pages of Price's (1982) article, listed below. Price attempts to go beyond an earlier concept called apparent thermal inertia, which was just a simple measure of the diurnal surface temperature amplitude. He introduces the concept of a *diurnal heat capacity* D, for which he derives a complex expression in terms of the diurnal surface temperature amplitude. So constituted, Price's analytical expression for D (which is P scaled by the earth's rotation rate) contains terms that relate the temperature wave to other parameters, such as soil moisture, wind speed and roughness length. Price's articles followed the launching of a satellite during the late 1970s, called the Heat Capacity Mapping Mission (HCMM). The HCMM was supposed to provide a measure of the surface thermal inertia pattern, but its full promise was never fully achieved, partly because of the short duration of the satellite's life span (1 yr.) and partly because the surface, with its vegetation cover, turned out to be even more complex than most people had anticipated. Like the Geologists, the Meteorologists found themselves in trouble in underestimating the complexity of the earth's surface! It remained for the 'third wave', the plant scientists and agronomists, to hurl themselves into the breach. However, things got somewhat better at this point because of the reduced importance of P in the vegetation problem. Indeed, when we come to vegetation, the effect of P diminishes with increasing vegetation amount.

Yet the impression that the radiant surface temperature contained valuable information pertaining to the land surface properties persisted. In Price's thermal inertia model the amplitude of the daily temperature wave is a measure of the inertia of the atmosphere to solar heating, the larger the amplitude the smaller the inertia and vice versa. Since the largest fluctuation in surface temperature due to solar heating is at the ground, the surface temperature will contain the greatest information about the surface properties. The maximum fluctuation in surface temperature is approximately between 2 AM and 2 PM, approximately the times of some satellite overpasses. Carlson et al. (1981) were able to derive values of thermal inertia for the land surface using satellite measurements of radiant surface temperatures at about 2 AM and 2 PM from the Heat Capacity Mapping Mission Satellite data.

Thermal Inertia in Simsphere

The Simsphere model is interactive and does not assume a sinusoidal variation in solar forcing. It is nevertheless possible to demonstrate the theoretical result from analytical but simpler models that thermal inertia is somewhat more fundamental to the diurnal temperature wave than thermal conductivity and diffusivity are by themselves. By that, we mean that if thermal conductivity and diffusivity are varied independently but keeping the thermal inertia constant the simulated temperature wave is very nearly unchanged, so long as extreme values of any of these parameters are avoided. We also allow P to depend on soil water content according to a formula that seemed to allow the best results. Specifying a zero thermal inertia allows the model to determine its own value by this formula. The user may override the default treatment of P by inserting something other than a value of zero for thermal inertia.

The Simsphere model uses both a soil thermal conductivity and a thermal diffusivity, (which are assumed to have the same values throughout), but it obtains these parameters from the specification of thermal inertia alone. The justification for using one parameter to generate two is as follows: If one plots λ' versus κ' for a wide variety of natural soils and materials (e.g., as taken from the **Manual of Remote Sensing**), one finds that the two parameters are highly correlated. Accordingly, we have used this correlative relationship to compute a thermal conductivity and diffusivity, given a thermal inertia supplied by the user. This statistical device somewhat reduces the necessity to guess at values which are scarcely known.

Precipitable Water

Another important parameter that profoundly affects night time cooling is the *down welling long wave (thermal) radiation* [Eqn. 2.1]. This is expressed by the formula, given by Eqn. 4.5, which relates the down welling thermal radiation to the *emissivity* of the atmosphere ϵ_a (nominally at a temperature $\overline{T}$, which is representative of a layer not far above the surface, *e.g.*, 50 or 150 m). Recall that the emissivity is the ratio of the actual thermal radiation to the blackbody radiation temperature, $\overline{T}$. Since the principal absorber in the atmosphere is not the air but the water vapor (and, to a lesser extent, carbon dioxide), the emissivity of the atmosphere (ϵ_a) is made a function of the precipitable water (W) in the atmosphere. The latter is the equivalent depth of liquid water in a column of air if all the vapor were condensed out as rainfall. Typically, the precipitable water amount is between 1 and 3 cm.

The simple use of precipitable water in the formula for down welling thermal radiation expresses the relationship between the amount of *absorber* (water vapor) and emissivity; alternately stated, the more water vapor in the atmosphere the closer the atmosphere resembles a black body, ε_a = 1.0. Since most of the water vapor in the atmosphere is close to the surface (i.e. the lowest 2 - 3 km), the effective radiating temperature will be that in the lower atmosphere. Increasing the precipitable water increases the downward long wave flux $L\downarrow$ and therefore retards the nocturnal cooling.

To this point, we are still talking of clear skies. The atmosphere is simulated as a blackbody radiator by arbitrarily increasing the precipitable water such that the emissivity of the atmosphere approaches 1.0. However, the condition that ε_a = 1.0 can also occur, in effect, with the presence of dense, low cloud cover. A cloud layer effectively makes the atmosphere a blackbody radiator at the temperature of the cloud base. If low enough, that emitting surface may have a temperature not far from that of the ground surface, in which case the net upward minus downward long wave radiation may be very close to zero. Of course, to simulate clouds during the day we would also need to reduce the incoming solar flux by a factor appropriate to whatever type and amount.

Simulations

Level 1

In this chapter, we will begin looking quite extensively at the behavior of night time temperatures and fluxes. Remember to extend the 'end time of the run' out to 5 AM in order to simulate the entire night time (24 hour) period.

Simulation series 1

- **Vary only the thermal inertia**

Soil type and condition is a large determinate of the thermal inertia of a surface. The Barrens, near State College, PA, boasts one of the largest amplitude temperature waves in the area, particularly on winter nights. This is primarily due to the sandy soil type of the Barrens, which typically has values near 6 TIU. Wet, loamy soils may have values about 18 -24 TIU. Many types of urban materials have values of P ranging from 6 to 24 TIU, but the old adage that urban materials are better conductors than rural soils is not supported by measurements shown in various tables (*e.g.* the **Manual of Remote Sensing**). (A note for the future: the type of soil one chooses is critical for the plant simulations. This choice depends upon the soil type. It is best to make sure in later simulations that the value of P, if chosen by the user, conforms to the type of soil selected for the plant modules.)

Thus, if a non-zero value of P is chosen, the thermal conductivity and diffusivity of the soil are calculated internally in the model, and is independent of the soil moisture. A choice of zero, however, means that P is dependent upon M_o, and is calculated accordingly. However, making P dependent on M limits our ability to investigate the variation in P over urban surfaces, where the material more than the substrate water content determines the value of P.

The simulations to this point have been made using a standard value of thermal inertia equal to 13 TIU. Make a default run, followed by some simulations with varying thermal inertia (parameter "TP" in the model input file). You should try and keep the values between 6 and 30 TIU, as the model may explode if given numbers out of this range. Concentrate on the surface flux output, including the ground flux, as well as the surface temperatures.

Questions

1. What do you observe about the daily temperature curve as P is varied?
2. Is the effect of changing P on surface air or radiant temperature more dramatic at night than during the day?
3. Examine the differences in the Bowen ratio throughout the day. Which of the surface fluxes appears most affected by a change in thermal inertia?
4. Try a very low value of P (= 5 TIU), simulating the Barrens environment. Why does the model yield such low night time temperatures under these circumstances? What is missing in the model that is present in the real world, and would control the rapid temperature decrease?

Simulation series 2

- **Vary the thermal inertia and the surface moisture availability**

Make simulations using a thermal inertia of 6, 12, and 18 TIU under different surface moisture environments (M_o = 0.1, 0.5, and 0.9). Again, look at the surface fluxes and temperatures in your output.

Questions

1. How does P affect the Bowen ratio for differing values of M_o?
2. Does the thermal inertia change the sensible heat flux more for dry or wet surfaces? Pay particular attention to the ground flux.
3. How is the ground flux affected by an increase in P and why?
4. How does a changing P affect the moisture availability itself?
5. Why does the daytime surface temperature response to P differ from the night time response so significantly?

Simulation series 3

- **Vary the precipitable water**

Vary the precipitable water in the sounding (and, in effect, the atmospheric emissivity) by increasing or decreasing the dew point depressions through the depth of the sounding. Note how these changes affect the nighttime temperature. Run the default case, then perform a simulation with a large value of ε_a (*e.g.*, 0.9) by increasing the precipitable water (reduce the sounding dew point depressions by, say, 5 C). Next, try lowering the precipitable water (increase the sounding dew point depressions by, say 5 C) and concentrate once again on the daily temperature cycle and surface fluxes.

Alternately, you can vary the cloud cover, if you wish. Note that this change will also affect incoming solar flux during the day, as mentioned. Make the cloud cover parameter ("CLOUD_FLAG" and "CLD_FRACT" in the model input file) equal to some fraction, e.g., 10 tenths (100%). Unless you have time or inclination, do not expend more effort on the precipitable water/cloud issue by performing more than one or two simulations.

Questions

1. What similarity to thermal inertia does precipitable water have insofar as the daily temperature wave is concerned?
2. At what range of W are the most drastic temperature variations seen (low or high values)?
3. Does the effect of a low W counteract the effect of a high P at night (and vice-versa)? In other words, does precipitable water exert a greater control on the daily temperature wave than thermal inertia does for a given fraction of its likely range?
4. Given that clear, calm nights are also those in which the air is dry (low precipitable water), what would you conclude would happen to the night time temperatures on clear, calm nights. To check this out further than you did in Chapter 2, run a simulation exactly as you did in that chapter but with a low value of TIU.
5. Given your simulation in question 4, can you produce a sub-freezing temperature for the summertime, mid latitude sounding at night, such as occasionally happens in the State College barrens in summer?

Level 2

Up to this point, we have been using the standard vertical profile of winds, temperature, and moisture for the summertime mid latitude sounding. Now, think about a location such as Florida compared to a desert region. Florida is characterized by high humidity throughout the year due the fact that it is surrounded by warm water, and its daily temperature wave does not vary more than 20°F on most occasions. The desert Southwest, however, is known for its large amplitude temperature wave, reaching high temperatures well over 100 F and then plummeting down to the 40s at night. It might be easy to explain this away due to the longwave cooling at night and low precipitable water of the desert, but that is not the whole story. Think about what other factors make the desert SW so hot during the day, yet so cool at night.

1. Run simulations using a desert sounding and compare it to that of a Florida-type sounding. Consider all the factors we have discussed through Chapter 4, including M_o, surface fluxes, wind speed, thermal inertia, and precipitable water. Also, hypothesize on how other variables, those that we have not discussed yet, could play a role, such as soil type, atmospheric humidity, and vegetation. Also, think about the region you live in; what physical parameters that we have discussed most greatly influence your overall climate?
2. Let us say I own a very large field of very short vegetation (and therefore I can, for the sake of these simulations, ignore vegetation for the present) and I wish to prevent the

night time temperatures from reaching freezing. If I wet the surface layer of the ground with water so as to make Mo equal to 1.0 will I get a higher night time temperature than if I leave the surface dry (M_o = 0.1). In approaching this problem you will want to make the thermal inertial large for the wet case and small for the dry case. Explore the possibilities of wetting the surface with both a mid-latitude and a tropical sounding. Never mind that the tropical sound may not exhibit low night time temperatures, as it pertains to a summertime situation; just examine the behavior of the night time temperatures and compare the various cases you have chosen. Or, you can introduce your own sounding appropriate of a December sounding over northern Florida during December.

Equations

[Eqn 4.1]

$$\rho' C' \frac{\Delta T}{\Delta t} = \frac{\Delta G}{\Delta z} = Q$$

[Eqn 4.2]

$$\lambda' \frac{\Delta T}{\Delta z} = G$$

[Eqn 4.3]

$$\kappa' \frac{\partial^2 T}{\partial z^2} = \frac{\partial T}{\partial t}$$

[Eqn 4.4]

$$\frac{\lambda'}{\sqrt{\kappa}} = \frac{1}{\sqrt{\lambda C' \rho'}} \propto P$$

[Eqn 4.5]

$$L\downarrow = \varepsilon(W)\,\sigma T_a^4$$

Terms

Terms to look up and remember
apparent thermal inertia
heat capacity
preciptible water
thermal conductivity
thermal diffusivity
thermal inertia

Simsphere Workbook: Chapter 5

Atmospheric Profile Background

Please review the introductory notes on the 'Structure of the Model' pertaining to the Planetary Boundary Layer before proceeding.

Introduction

We are all familiar with the idea that temperature decreases with height; if you do not believe this, stick your head out of an airplane, or, better yet, look at the 200 mb (40,000 ft level) temperatures on the Penn State Weather Station map wall. If we lift a parcel, or 'blob', of air by a dry adiabatic path (without any addition or subtraction of heat) the air parcel expands and the temperature decreases by an amount approximately equal to 10º C per kilometer. This is called the *dry adiabatic lapse rate* of temperature, and pertains to situations where parcels of air are lifted or lowered in the atmosphere; (A sinking parcel of air would see its temperature increase by 10º C per kilometer). Customarily, lapse rate is defined mathematically as $-\frac{\partial T}{\partial z}$ where the minus sign is simply present to make the quantity usually positive.

The atmosphere as a whole, however, changes its vertical temperature profile as different air streams reach varying levels of the atmosphere. The change in temperature with height of the atmosphere itself is called the *environmental lapse rate*. Under normal circumstances, temperature decreases with height, sometimes by more than 10º C per kilometer, and sometimes by less. On average, temperature decreases at a rate of about 5º C per kilometer. An environmental lapse rate of 10º C per kilometer (dry adiabatic) in a column of air will result from completely mixing parcels of air in that column dry adiabatically. Turbulent heating and mixing of an air column will result in the formation of uniform heat content, a uniform distribution of passive constituents, such as water vapor and dust, and is called the *mixing layer* or sometimes just the *mixed layer*. When temperature increases with height, we say that there is a temperature inversion. This typically occurs near dawn as the air nearest the Earth has cooled off more rapidly than the air above it due to radiational cooling.

Now, if we return to the classic conduction laws and our discussion of resistances, we note that heat is conducted down a temperature gradient. Since we are talking about sensible heat, the appropriate gradient for the conduction of sensible heat is not the temperature but the potential temperature. The *potential temperature* is simply a temperature normalized for *adiabatic compression or expansion*, *i.e.* it is defined as the temperature an air parcel would be if it were brought down or up dry adiabatically to some reference level (1000 Mb). Therefore, if the measured temperature at 2000 meters above the 1000 Mb level was 12º C, the potential temperature of that level would turn out to be 32º C, using the dry adiabatic lapse rate for lowering a parcel of air by 2 km to 1000 Mb. In meteorology, the reference level is chosen as 1000 Mb; in the model, the reference level for all calculations is the ground surface, but potential temperature is still accounted for in the conventional manner in the output. (The difference between actual temperature and potential temperature is unimportant over the depth of the surface layer (50 m), so we will henceforth simply refer to the temperature when discussing vertical derivatives in the surface layer.) Please note that surface radiant temperature is the actual temperature.

In a *statically stable atmosphere*, one where vertical air motions are suppressed, potential temperature increases with height. As we shall see, the atmosphere is almost everywhere statically stable, except over shallow layers, such as those in immediate contact with the ground (the surface layer). The mixing layer (typical depth 1-3 km) grows with time in the model, and takes sensible heat flux at the ground surface and distributes it within a layer of constant potential temperature above the surface layer. (A minor limitation of the model is that it automatically mixes potential temperature evenly in the mixing layer. We could have computed potential temperature in the mixing layer differently, allowing it to evolve more naturally, but we didn't feel at the time that it was worth the added computational effort.)

The environmental lapse rate of temperature is directly tied to the atmospheric static stability. When the environmental lapse rate is exactly dry adiabatic (10º C per kilometer), there is zero variation of potential temperature with height and we say that the atmosphere is in *neutral equilibrium*. Under these conditions, a parcel of air forced upwards (downwards) will stay where it is once moved, and not tend to sink or rise after released because it will have cooled (warmed) at exactly the same rate as the environment. (here, we are taking the liberty to ignore small effects due to the differences between parcel and environmental water vapor. Strictly speaking, buoyancy is evaluated with respect to air density differences rather than air temperature differences.) When the environmental lapse rate of potential temperature increases with height, the atmosphere is *statically stable*. The more stable the environment, the more the atmosphere resists convective overturning by air motions. If a parcel of air were brought upwards (downwards) in a statically stable environment, it would tend to sink (rise) back to its original level once released because it will have cooled (warmed) more rapidly than the environment and will be like a stone in a glass of water (cooler = more dense). If the environmental lapse rate of potential temperature decreases with height, the atmosphere is *statically unstable* with respect to dry adiabatic processes. Under these circumstances, a parcel of air brought upwards (downwards) would tend to keep rising (sinking) once released because it will have cooled (warmed) less rapidly than the environment and will be more buoyant than the surrounding environment (like a hot air balloon). A statically unstable environment promotes the formation of convective clouds and thunderstorms. As stated, however, the environment is almost always statically stable on the large scale.

Of less importance, but worth noting, is that the model also computes a downward flux of sensible heat at the top of the mixing layer; this flux is entrained from above because the mixing layer rises into air with a higher potential temperature (under typical, statically stable conditions). The top of the mixing layer is crowned with a small inversion, called a *mixing inversion*. Both momentum and specific humidity are mixed by diffusion using mixing coefficients that depend on the magnitude of the sensible heat flux and the wind speed. As the mixing layer grows with time, drier air from above the mixing layer (the usual case) and faster moving air from above (also the usual case) are mixed down into the mixing layer within which both specific humidity and momentum tend to be fairly evenly mixed.

In examining the vertical profile of temperature (environmental lapse rate) as the surface layer evolves, we notice that the surface temperature is warmer than the air temperature at 50 m during the day, and is cooler at night. At night, the ground cools with time, as does the atmosphere near the ground, including the bottom part of the day's mixing layer. This leads to the formation of an inversion near the ground at night. This also makes sense in view of our conduction laws, because the sensible heat flux is upward during the day (down the potential temperature gradient) and downward at night.

However, consider the following paradox: the difference between the surface temperature and that at the top of the surface layer (50 m) is about as large during the day as at night (except that the signs are reversed -- increasing temperature at night, decreasing temperature during the day), but the night time sensible heat flux is rather minuscule. One might wonder why the magnitude of the nocturnal sensible heat flux is so trivial if the vertical gradient of potential temperature is the same as during the day (when the sensible heat flux is quite significant) as at night?

Simulations

Level 1

Simulation series 1

Change the static stability

Perform a default simulation, then one where the temperatures of the lowest four layers of the atmosphere are required to become more stable. Create a slight inversion in these layers, for example, keep 24º C at the surface, then increase the three layers above the surface to 25, 26 and 27º C (the second, third and fourth values of “TS” in the model input file). Graph the surface temperature (TRAD in the output file) and the 50 m temperature (T50 in the output file) as well as the surface fluxes and compare the two runs.

Questions

1. Because we have made the atmosphere more statically stable, would you expect more or less vertical air motions?
2. What would this imply about the transfer of sensible heat upwards through the atmosphere?
3. Do your temperature results at the surface and 50 meters support this idea? What effect might a more stable lower atmosphere have on the distribution of pollutants and water vapor (look at the vertical profile output or the specific humidity near the surface and at 50 m for each run)?

Simulation series 2

Vary the surface moisture availability

Redo your moisture availability simulations using a low (0.1) and a high moisture availability (0.9). Watch the growth of the mixing layer as reflected in the vertical profile of wind speed, potential temperature and specific humidity.

Questions

1. Can you identify the top of the mixing layer?
2. Which simulation creates a larger mixing layer and why?
3. At what time does the mixing layer appear to cease growing?
4. Which combination of static stability and moisture availability would create the shallowest mixing layer?
5. Why does the specific humidity not appear perfectly mixed in either case, especially near the top of the mixing layer?
6. Why don't the winds become perfectly mixed?

Level 2

Simulation series 3

The Los Angeles (L.A.) area is known for its smoggy and hazy conditions where a large amount of pollution and water vapor is trapped near the ground. Due to a stable atmosphere, the mixing layer of the L.A. basin is very shallow. An obvious solution to their problem would be to reduce the release of harmful pollutants into the atmosphere. Simulate what effect a reduction in pollutants would have on the mixing layer depth by treating pollutants as water vapor in the model. (Water vapor and pollutants tend to absorb solar radiation.) In other words, create a very stable sounding with ample water vapor (representing both water vapor and pollutants). Then, reduce the water vapor considerably and view the effects on the mixing layer depth and temperature profile. Try the same thing by including a small cloud factor, say 0.1.

Questions

1. Have the high levels of air pollution contributed to the shallow mixing layer and hazy conditions of Los Angeles? In other words, will L.A. still be prone to hazy conditions (due only to water vapor) once pollution is minimized, or will the reduction of pollution help 'stir' the lower atmosphere? Compare your L.A. simulations with a less stable atmosphere representative of say, Pittsburgh.
2. Do you think the reputation of Los Angeles as heavy air polluters may be exaggerated, and that they are simply victims of their own static stability?

Simulation series 4

Using the same conditions for L.A. in the first simulation, observe what happens to the lapse rate at night and in the first part of the morning the next day; (you can observe the latter by letting the model run through the first few hours after dawn the next day). Find the bottom of the daytime mixed layer and note the top as well.

Questions

1. What happens to the bottom of the daytime mixed layer with time at night?
2. What is responsible for the erosion of this layer?
3. At what time in the morning will the surface air begin to mix through the nocturnal inversion to reach the previous day's mixed layer?

Terms

Terms to look up and remember
buoyancy
dry adiabatic
inversion
lapse rate
mixing inversion
potential temperature
static stability: neutral; stable; unstable

References

- Stull, R. B., 1988, An Introduction to Boundary Layer Meteorology, pp 105-109.

Simsphere Workbook: Chapter 6

Simsphere Vegetation

Introduction

Now that we move into the second part of the course, we will study the role of vegetation in the modification of the surface microclimate. It is important that you understand and relate your readings and previous work to the present conditions, those that involve the addition of a layer of vegetation at the soil surface. At this point, we would like to introduce the following word of caution: don't prejudge what vegetation will or will not do to the surface energy balance. Depending on the state of the soil moisture, the state of stress within the plant (we will later define 'stress'), and the state of the atmosphere (winds, temperature and humidity), plants may evaporate water ('*transpire*') more or less rapidly than bare soil, even if the soil water content and atmospheric conditions are identical for both vegetated and bare soil surfaces.

The concept of soil water content, especially for plants, becomes an increasingly more complex idea as one examines its various aspects. Take, for example, the concept of moisture availability. To what extent does moisture availability control surface fluxes when applied in a model for a vegetated surface? This is just one example, but there are many more, and you will need to question their significance within the context of vegetation. We will, as before, begin very simply by considering just a few important parameters, but we will soon gain ground (excuse the pun) and add several levels of abstraction that may be quite new to many of you.

A Little Plant Physiology

As scandalous as it may seem, some plant scientists believe that plants do not have a mind of their own, at least when it comes to the exchange of water vapor with the atmosphere. Idso (1983) even had the nerve to suggest that plants act as passive wicks and simply transfer water through their straw-like little bodies between the soil and the air, in response to atmospheric demand and availability of water in the soil. These poor veggy creatures, says Idso, have no say whatsoever in the face of atmospheric 'extortion'. If the air is dry, the sun strong and the wind speed large, the plants must cough up more water vapor or face the consequences. Or must they? Can plants do anything about their environment? Can they form a union and resist outside demands on their well-being.

The problem lies in the fact that plants do not benefit by losing water to the atmosphere. The water evaporated from their leaves (*transpiration*) is parted with reluctance. (Plants can store a certain amount of water in their roots, stems and leaves - for a 'rainy' (or, rather, a dry) day (so to speak !). (Note that transpiration plus evaporation from the surface of the earth together is called *evapotranspiration*.) Plants must nevertheless transpire for two reasons. First, they need to obtain sustenance to live by bringing up nutrients in the form of sap from the roots; they must also reproduce, and in order to do so the plants must absorb carbon atoms in the form of ambient carbon dioxide vapor (CO_2) through many small pores, called *stomates*. (Unlike you and me, plants take food from both ends!) As carbon dioxide enters the leaf, and nutrients and water enter the roots from the soil, water vapor escapes by the same route to the outside, after first being evaporated from liquid water in the *substomatal cavities* -- the intercellular air spaces (Figure 7.1 shows a schematic illustration of the leaf structure see Jones, 1983; p 110). Water vapor then escapes through the surface of the leaf to the surrounding airstream. Carbon dioxide enters by the same path, but in the opposite direction and with nearly identical aerodynamic parameters (*resistances*). There are thousands of stomates on a typical leaf, but an individual one is so small that it can not be seen with the naked eye. A small fraction of the water in the intercellular leaf mass avoids the main route through the stomates and is conducted through the outer waxy surface of the leaf, called the *cuticle*, to the surrounding air layer.

The second reason for the plant's parsimonious behavior with regard to water vapor is that *photosynthesis* is limited by sunlight. The more sunlight the larger the *assimilation rate* of carbon by the plant. However, under strong sunlight the leaf temperature can rise to dangerous levels, subjecting the plant to heat prostration, to which it responds by ceasing to make new tissue, losing its leaves or simply dying. In order to regulate its temperature, the plant must evaporate water from its leaves, thus keeping the plant cool (evaporation is a cooling process).

The ratio of the carbon assimilation rate (FCO_2) to the transpiration rate (*e.g.* per unit leaf area T_r) is a measure of the *water use efficiency* of a plant. For the plant, its game is to maximize the water use efficiency by holding on to as much water as it can while still absorbing a maximum amount of CO_2. Some plant scientists believe that each plant has evolved an optimum efficiency ratio for its particular set of conditions. Lower the efficiency and the amount of biomass produced by the plant decreases.

We provide much of the information above as an introduction to how plants survive, operate, and affect the atmosphere. We will get into more detail on these processes in later chapters, after we learn how vegetation, in general, affects the surface energy budget differently than a bare soil surface does.

Simulations

Level 1

The object of this scenario is to run the model with a standard set of vegetation parameters (as in scenario 1) and compare the results with the bare soil case which you will need to rerun. The reason for rerunning the bare soil case is to compare it with simulations made for a realistic crop canopy for which we can reasonably estimate the parameters. Do not become overwhelmed with the vast amount of detail and options that face you in the vegetation scenario. Take each step, one at a time and do not become lost changing a vast number of parameters. As we emphasized in the introduction to this chapter, do not expect that the results for the vegetation simulation will differ greatly from those for bare soil, or that the differences are specifically due to vegetation. Since things will get a bit complicated, let us just examine the latent heat, sensible heat and ground fluxes and the surface temperature for this chapter. One word of caution: the moisture parameter that governs the plant hydraulics is the root zone moisture availability, which is set independently of the surface moisture availability (although the two soil layers can exchange water according to the vertical gradient of soil water). If you set the root zone moisture availability at too small a value, you will kill the plant. The model reacts to this senseless act of butchery by yielding bizarre results or simply dying! Do not be alarmed if the values bounce around a bit during the first hour of the simulation. Like you and me, the plants take a bit of time to get going in the morning. Note also that the vegetation module applies only during conditions of positive net radiation; at night we treat the surface as if the vegetation were absent.

Simulation series 1

- **Include vegetation**

Run a standard simulation for the bare soil case, that is the simulation you started with in the first chapter. Then, introduce a mature corn canopy to the model by assigning the following parameters:

- o leaf area index (LAI) of 3.0 (XLAI=3.0),
- o fractional vegetation cover set at 100% (FRVEG=100.0),
- o surface roughness of 20 cm (0.2 meters) (ZO=0.2),
- o vegetation height of 2 meters (VEGHEIGHT=2.0),
- o vegetation type **corn (INDEX_VEGGIES=4)**.

As before, use a *surface* moisture availability of 0.5 (F=0.5). However, assign the same value of 0.5 for the **deep layer** moisture availability (FSUB=0.5, and be careful if you ever choose to decrease this parameter below 0.3 - 0.4 when vegetation is specified). Make a simulation under these conditions and compare your results to the standard case you obtained for the non-vegetated conditions with the same soil water content, concentrating solely on the surface fluxes, including the ground flux, and the surface temperatures.

The above parameters are representative of conditions in a mature corn field. Consult the 'user's guide' for more details on these parameters as well as the relationship between LAI and fractional vegetation cover.

Questions

1. How has this corn field affected the surface energy budget and the surface temperature?
2. Are the changes as you would expect?
3. Why does there appear to be a lag in the maximum ground flux for the vegetated run?

Simulation series 2

- **Vary the surface and root zone moisture availability**

Make a table with bare soil and vegetation cases in 2 rows, and low soil moisture and high soil moisture in 2 columns. Record in each cell the 10 meter temperature, sensible heat flux and latent heat flux at, say, 1200 hours for each model run case:

- o For bare soil, low soil moisture, use FRVEG=0.0, F=0.1, FSUB=0.1
- o For bare soil, high soil moisture, use FRVEG=0.0, F=0.9, FSUB=0.9
- o For vegetated, low soil moisture, use FRVEG=100.0, F=0.1, FSUB=0.1
- o For vegetated, high soil moisture, use FRVEG=100.0, F=0.9, FSUB=0.9

Questions

1. What are the sensitivities of the surface fluxes and temperatures to increasing soil water content in vegetation versus bare soil simulations? How do the surface and root zone soil moisture availability values vary between the bare soil and vegetation runs, and why?
2. Why is the 10 meter temperature significantly warmer for the bare soil, low soil moisture case than for the other cases?
3. Is surface moisture availability a more important parameter in determining the surface energy budget for bare soil or for vegetation?

Simulation series 3

- **Vary the leaf area index (LAI)**

Restore the moisture availability to their original values of 0.5. Now, vary the LAI from 3.0 to 7.0 and then from 3.0 to 1.0. (It is not a wise idea to let LAI become less than 1.0, as this begins to pose a contradiction in terms.)

Questions

1. Which component of the surface energy balance (which flux) does a change in the thickness (LAI) of the vegetation canopy affect most?
2. Can you tell if the overall changes in the fluxes and temperatures due to the varying LAI are more or less than those seen in simulation 1?
3. At what values of LAI do the results begin to show an insensitivity to changing LAI?

Simulation series 4

- **Vary the thermal inertia and/or moisture availability**

Try varying the thermal inertia from 13 to 6 TIU or the moisture availability from 0.5 to 0.9 at differing levels of LAI (1,3, or 7). Focus on the magnitude of the changes in fluxes due to these parameters being varied.

Questions

1. What can you conclude about the importance of soil surface characteristics (e.g., thermal inertia or M at the surface) for the fluxes in the cases of dense and sparse vegetation?
2. Can you tell if these two parameters are more important for bare soil or for vegetated conditions?

3. What about for sparse vs. dense vegetation? Is M0 more important than thermal inertia overall?

Simulation series 5

- **Vary the fractional vegetation cover**

Perform a vegetation simulation identical to that of simulation #1. Next, lower the fractional vegetation cover from 100% to 50% and compare it with the original run. This indicates that half the ground is covered by corn with an LAI of 3.0, while the other half of the ground consists of bare soil.

Bear in mind that the model output represents the combined, average fluxes from the vegetated and bare soil areas.

Questions

1. What major changes do you see in the flux and temperature output?
2. Does fractional vegetation cover appear to be a more or less important parameter than the LAI?
3. How do the surface fluxes and temperature vary with changing vegetation cover?
4. At what point does a variation in fractional vegetation cover become relatively unimportant, or at least much less sensitive than in changing from no vegetation to 20% vegetation, say?

Simulation series 6

- **Vary the surface roughness**

Finally, if you feel like playing still more with the model, vary surface roughness (z_o) and see how things change. The present value corresponds to a full, mature corn canopy with a roughness of 20 cm (about 0.1 the height of the corn). After running the standard case, change the roughness to 10 cm and then 30 cm. Again, we design these simulations to give a feel for the impact certain parameters have on the atmosphere. Some parameters, as we have seen, have a much greater influence than others do, and it is important to know which ones if we are to push ahead and utilize this model to its fullest potential. One thing to remember about roughness length is that it always appears in the mathematical formulation in a logarithm function.

Questions

1. Do you see much change in the energy budget or temperatures?

If you wanted to be a little more realistic in this scenario, a decrease in the roughness might also correspond to a decrease in the height of the corn canopy, size of the leaves, and the LAI. We again iterate that you confine yourself to changing just one parameter. Changing everything at once and running the model may give you the impression of driving a souped up car, but the experience may also be 'dangerous', not the least danger of which is being unable to interpret the results!! In any case the suggestions for additional simulations made in these last two or three paragraphs are only to provoke conjecture as to how the various fluxes, temperatures and winds might respond to changes in external parameters. There is obviously not enough time to do everything.

Level 2

Construct a graph in which the vertical axis is labelled fractional vegetation cover and the horizontal axis is labelled surface radiant temperature. With $M_o = 0$ vary the fractional vegetation cover from zero to 1.0. (Remember that zero fractional vegetation cover renders LAI meaningless.) Plot this line on the graph. Then do the same for $M_o = 0.2, 0.4, 0.6, 0.8, 1.0$. Plot these lines. Now you have an overview of the sensitivity of the surface radiant temperature over the full range of M_o and fractional vegetation cover. You can do the same for evapotranspiration.

Questions

1. What sort of shape does the overall plot make? Contrast the behavior of the M0 isopleths in your graph versus the evapotranspiration isopleths.
2. Given a fractional vegetation cover of 0.2, what is the warmest that a surface radiant temperature can become when M0 goes to zero? Why doesn't the evapotranspiration go to zero at this point?
3. In what part of the figure would one no longer see meaningfully different values of M0, given a surface radiant temperature and a fractional vegetation cover measurement?
4. Draw an arrow to represent the change in M0, surface radiant temperature and evapotranspiration to represent a land surface parcel undergoing development over time from a corn field to a housing development.
5. After reflecting on question 4, draw another arrow to represent the changes that would occur if you were a landscape architect and could plant some grass and tree cover on your development.

Terms

Terms to look up and remember
Assimilation Rate
Cuticle
Photosynthesis
Stomate
Substomatal cavities
Transpiration
Water Use Efficiency

References

- Idso, S., 1983, Stomatal regulation of evaporation from well-watered plant canopies: a new synthesis, **Ag. Met.**, 29, 213 - 217.
- Jones, H. G., 1983, **Plants and Microclimate**, Cambridge University Press, 323 pp.
- Figure 1 Pathways for water loss from one surface of a leaf. (Adapted from Jones, 1983; p. 110)
- Oke, T. R., 1987, Boundary Layer Climates, Second Edition, pp 38- 2; 59-76.

Simsphere Workbook: Chapter 7

Plant Microclimate

Introduction

Let us now take a closer look at plants and at our plant model. Recall that the model treats the vegetation layer as a 'big leaf'. In the big leaf model, we imagine that the flux relationships and ambient conditions pertain to a single big leaf. This is identical to saying that the energy balance for a single leaf pertains to all of the leaves, each of which is identical to all the others; however we do not quite go so far as to make the entire vegetation canopy a single uniform leaf. Our big leaf is porous, having interleaf air spaces which envelop a canopy microclimate that is different from that in the atmosphere above the canopy.

Below the leaf canopy and its air spaces lies the bare soil. Imposing a layer of vegetation between atmosphere and soil results in some significant differences between the bare soil and vegetation cases. First, ground fluxes are exchanged between the soil and interleaf air spaces, rather than directly between soil and the air above the canopy. Second, radiative fluxes are distributed between bare soil and vegetation according to the density of vegetation and, third, the plants extract water not from the soil surface but from a deeper *root zone*. This structure allows for a decoupling between the atmosphere above the big leaf, the canopy microclimate, and the soil surface. Thus, vapor flux from the leaves change the local specific humidity inside the canopy, whereas over bare soil the water vapor introduced at the surface is mixed to higher levels. Therefore, the water vapor in a plant canopy is restrained from mixing due to the lower wind speeds within the interleaf air spaces. Whether a group of plants can completely 'take control' of their environment or not is a question that has not been fully explored, but we will attempt to do so in a later scenario.

An important question that we will pose in later chapters is how the plant microclimate differs from that for the atmosphere over bare soil. In so doing, we will investigate to what extent plants can control their microclimate, including the very important fluxes of sensible and latent heat. In this chapter, we first examine the vapor and sensible heat flux pathways between the big leaf (and the soil) and the air layer above the canopy, and then discuss the mechanisms by which the leaf regulates transpiration.

Latent heat fluxes from the leaf

Let us consider some elementary relationships governing the fluxes of heat and water vapor between the plant and the environment. Recall from the last scenario and the attendant figure, that the flux of water vapor from a plant (or latent heat flux) is called the *transpiration*. The transpiration flux must pass through several resistors between the interior of the leaf and the airstream above the plant canopy. First is the resistance imposed by the leaf skin, then the air resistance in the leaf boundary layer, and finally the air resistance between the interleaf air spaces and the atmosphere above the plant canopy.

Now, look at the whole pathway for water vapor between the leaf and the air layer above the plants. Can the plant really influence these resistances or is it subject to the whims of the environment? After all, any number of factors play a role in the regulation of these resistances. Some of these factors originate in the environment and some in the plant. Like reflexes in a human being (perspiring, blinking, etc.), our own responses, especially those on which we depend for survival, are regulated without conscious intervention. We believe that plants can exert a reflexive control under some circumstances (recall our reference to Idso (1983) in the last scenario). They can regulate the size of the openings in their leaf through which the vapor passes to the outside of the leaf, but perhaps no more than we can regulate the size of the pupils in our eyes. These openings, shown in Figure 7.1, are the stomates. Plants regulate their *stomatal aperture*, and, in turn, their *stomatal resistance* in such a way as to limit or augment the flow of water to the outside.

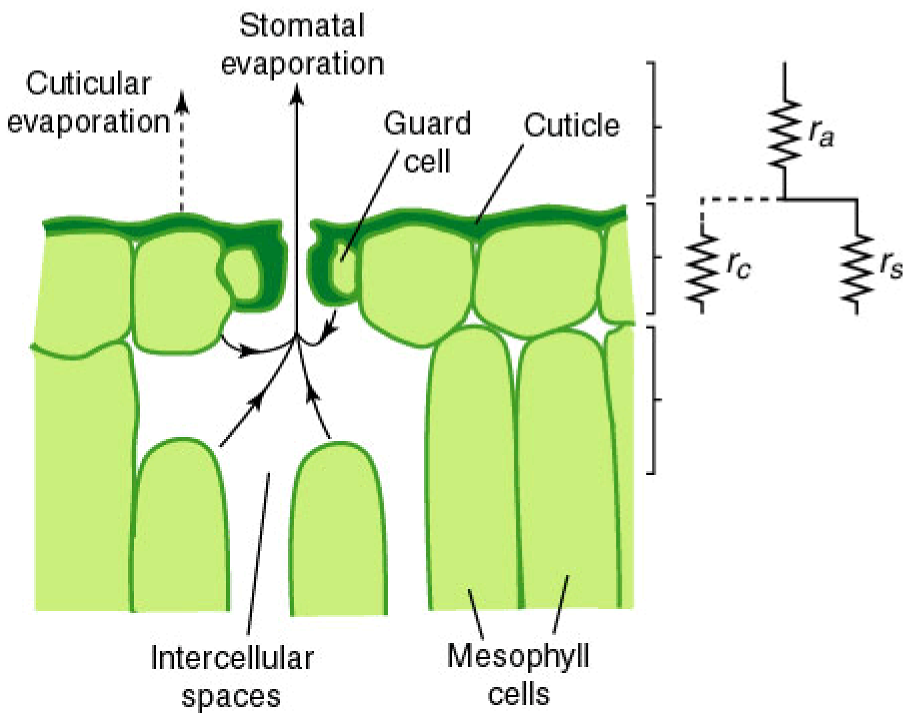

Figure 7.1 Pathways for water loss from one surface of a leaf, showing the boundary layer (r_a), cuticular (r_c) and stomatal (r_s) resistances. The leaf resistance is in parallel sums of two resistances, one representing r_s and the other plus r_c (Freely adapted from Jones, 1983; p. 110)

Consider the transpiration flux (T) per leaf area (A) in resistance notation, of the form given in Eqn. 3.1, referring only to the series resistance of stomata and leaf boundary layer. This formulation is shown in Eqn. 7.1 for the case of a leaf situated inside a canopy. Note that the potential drop in specific humidity occurs between saturation at the leaf temperature T_l, to that in the *interleaf air spaces* where the specific humidity is q_{af}. The ability of the stomates to impede the flow of water vapor from the interior of the leaf to the atmosphere can be represented as a resistance r_s, which is called the *stomatal resistance*. (The inverse of stomatal resistance, a term favored by many plant scientists, is the *stomatal conductance* g_s.) Stomatal aperture is controlled by cells on the leaf *epidermis* surrounding the stomatal cavity; these are called *guard cells*. Beneath the epidermis lie the *mesophyll* cells, the factories responsible for photosynthesis. Some aspects of this structure are shown in Figure 7.1, which is taken from a book by Jones (1983). The subsurface (mesophyll) cells are surrounded by *intercellular air spaces*, the substomatal cavities within which water vapor prepares to exit through the stomates and carbon dioxide molecules prepare to undergo photosynthesis. These substomatal cavities are busy two-way streets with carbon dioxide molecules and water vapor molecules hurrying in different directions, along with unwanted constituents from the outside, such as sulphur dioxide and ozone. Water passes from the inside of the leaf through parallel channels: the stomates (within which the flow is governed by stomatal resistance) and the cuticle (within which the flow is governed by the *cuticular resistance* r_{cut} but expressed as r_c by Jones in Figure 7.1). The cuticle is a hard, waxy surface on the leaf, through which atmospheric constituents can pass only with great difficulty. Since the cuticular resistance is usually an order of magnitude larger than the stomatal resistance, the principle channel for water vapor is through the stomates and we will henceforth assume that the *individual leaf stomatal resistance* r_l is approximately that of the stomates. When considering a leaf canopy, the individual leaf resistance r_l is scaled by the leaf area index (LAI) to produce the canopy stomatal resistance r_s which is lower than r_l because the individual leaf resistances function in parallel. Our model accounts for both r_s and r_{cut} but the latter exerts almost no effect on the results and its exact or even approximate values are only guesses. Thus, for the present we can consider the transpiration stream from a leaf to the air to pass through two series resistors before it can escape to the interleaf air spaces: The *stomatal resistance* and the *leaf boundary layer resistance* r_{af}. These two resistors are represented in the denominator of the flux equation (Eqn. 7.1).

Now, r_{af} is governed by the mechanics of laminar fluid flow (of air) over an object, in this case the leaf. In accordance with the theory of heat transfer, raf decreases as the wind speed increases and decreases with decreasing leaf size (Campbell, 1977; pp 66-67). Thus the leaf boundary layer resistance will tend to be small in strong winds when the leaf size is small. (In addition to wind speed, you can also change leaf size in the initial conditions of the model.)

Ignoring the cuticular resistance, the total resistance to water vapor from the interior of the leaf to the surrounding air spaces is $r_s + r_{af}$. Naturally, leaves have two sides, so Figure 7.2 (taken from Monteith 1990; p 188; see also his figure 9.6) shows the two parallel branches from both sides of the leaf; the fluxes from both sides of the leaf join in parallel and make their way to the top of the canopy. The top side of the leaf is called the *adaxial* side and the bottom side the *abaxial* side. Some leaves transpire through only a single side, having no stomates on the other face. (Here, we will stop short of describing the further adventures of the transpiration stream in moving from the air in the interleaf air space surrounding the leaf boundary layer through the atmospheric surface layer above the plant canopy.)

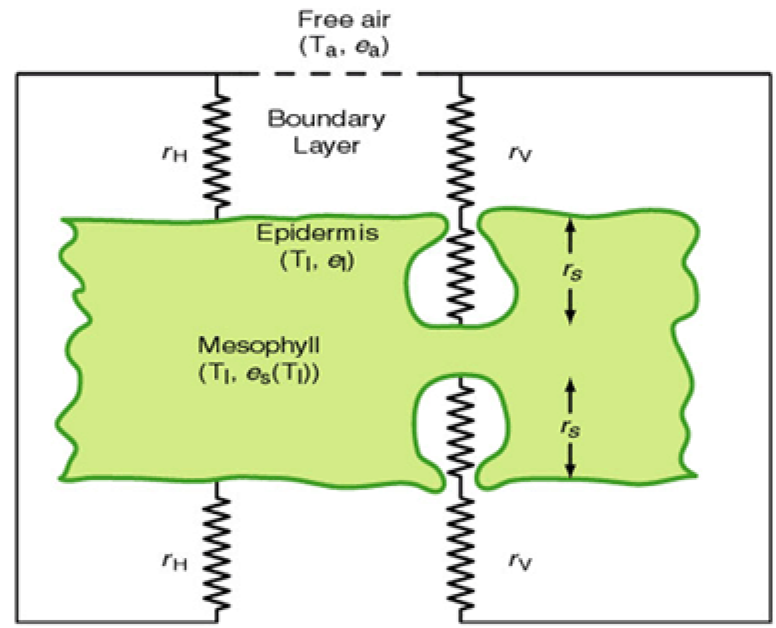

Figure 7.2 Electrical analog for transpiration from a leaf (E; right side) and a leaf heat balance (C; left side). Note that the symbols T_o and e_o stand for leaf temperature and specific humidity. Boundary layer resistances for heat and water vapor (r_h) are considered to be identical. r_s represents the stomatal resistance for this single stomate pair on upfacing and downfacing sides of the leaf. Note also that T_1 and e_1 refer to temperature and vapor pressure at the outside of the leaf boundary layer.(Freely adapted from Monteith, 1975; p 180.)

Figure 7.3 (Monteith, 1990; p 246) shows a many tiered canopy model in which the fluxes from each leaf layer join the main stream of fluxes. The flux from each leaf level moves from the leaf to the surrounding air space (through stomatal and leaf boundary layer resistances) and thence to the top of the canopy. The lowest segment of the vapor flux is the evaporation from the bare soil (L_eE) at the base of the canopy. Evaporation from the surface layer of bare soil is governed by the same constraints discussed in the second scenario, (*e.g.*, the moisture availability). However, the sunlight and wind speed are limited by the presence of the vegetation and the resistance between the surface and the interleaf air spaces is increased due to a reduced wind speed inside the canopy. The sum of the transpiration and the surface evaporation is called the *evapotranspiration*.

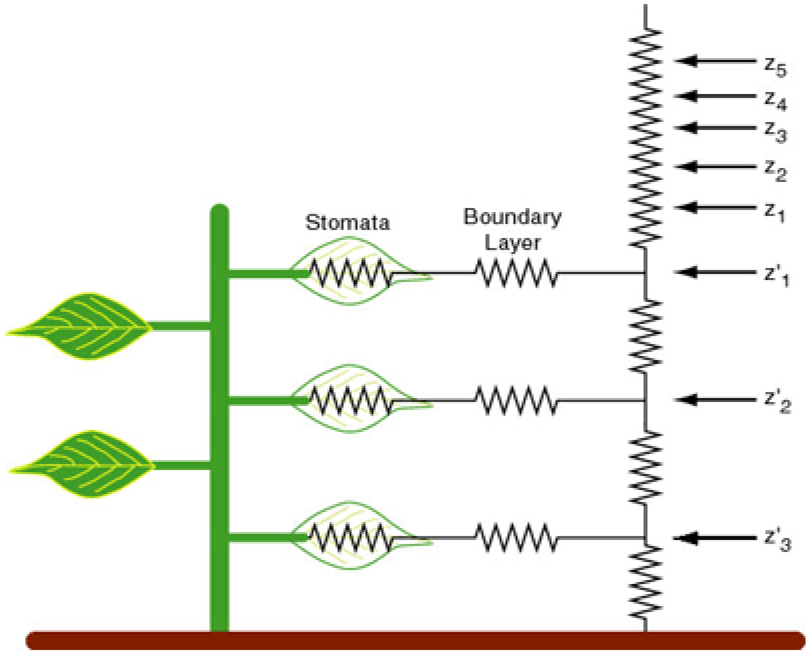

Figure 7.3 Resistance model for a plant in a stand of vegetation. Serrated segments refer to resistors, labeled accordingly. Our model is represented by just one layer (rather than three layers) of vegetation. (Freely adapted from Monteith, 1990; p 246.)

Such multi-layered models of plant vapor and heat fluxes, as depicted in Figure 7.3, can be quite elaborate, if not cumbersome and unnecessary. (You may wish to note that some models employ two vegetation layers. Most of the fluxes occur in the upper third of the canopy in the sunlit leaves. Since our model employs a single layer of vegetation, there is only one pair of air resistors to consider within the vegetation canopy: one between the leaf surface and the surrounding air and another between the ground surface and the interleaf air spaces. After that, the ground fluxes and leaf fluxes join and move above the canopy through air resistance at the top and above the canopy, called r_a; this resistance, like all resistances presented thus far, decreases with increasing wind speed.

Sensible heat flux

Unlike vapor flux, the sensible heat flux, H, from the leaves does not originate from inside the leaf, but from the surface of the leaves and so does not need to pass through the stomates. Resistance equations governing sensible heat flux in our model are identical to that in Eqn. 7.1, but without the stomatal or cuticle resistance components. Note that in Figure 7.2 we depict the sensible heat flux H on the left side and the latent heat flux on the right side of the figure. Sensible heat flux also originates at the bare soil surface beneath the canopy, and flows upward through the air resistance between soil and interleaf air spaces.

The two components of evaporated water flux (transpiration and bare soil evaporation) and the sum of the vegetation and bare soil sensible heat flux are limited by the total net radiation over the canopy. In the model, heat is not stored in the vegetation, but substrate sensible heat flux is exchanged between the bare soil surface and the substrate. We also assume that resistances for heat, water vapor and momentum are equal inside the plant canopy. Two vegetation parameters must be understood in more detail before implementing various plant canopies in our model, however: The leaf area index and the stomatal resistance.

Leaf area index

One of the most complex concepts in this type of model is to calculate the total fluxes over a horizontal surface area. Fluxes are exchanged between the surface area of the leaves and the environment, as well as between the soil surface and the environment. The leaf surface area is not the same as the horizontal surface area, however. To equate fluxes over leaf surfaces to those over a horizontal surface, one must know the total surface area of the leaves, among other parameters. The concept of a surface is easier to imagine for a horizontal ground surface than for a multi-faceted leaf canopy. A useful parameter in determining the net canopy flux over a horizontal surface area is the ratio of total one-sided leaf area to a unit of underlying horizontal surface area. This ratio is called the ***l**eaf **a**rea **i**ndex* or *LAI*. The LAI tends to range from a value of zero (bare soil) to about 7 for a dense canopy. Values less than about 3 pertain to canopies in which bare soil patches become visible with decreasing LAI. At small LAI (*e.g.*, less than values close to 1) the 'big leaf' model is no longer appropriate. Instead, more accurately, the plant canopy is depicted as clumps of vegetation with a specific LAI covering a fraction of the surface (Fr) and separated by areas of bare soil. Let's ignore this complication for the present and avoid low values of LAI. Both the LAI and the fractional vegetation cover (Fr) is specified in the model.

You might ask, 'why is just the one side of a leaf and not both sides considered in the LAI'. The answer is that many leaves have stomates on one side only, or at least they undergo most of the transpiration and sensible heat flux on one side. (This is not always the case and the bookkeeping of all the fluxes is definitely a source of confusion, even for the modeller.)

Leaf area index is used by plant scientists in many ways. It is a measure of the amount of *green leaf biomass* in a vegetation canopy. In the model, leaf area index is used to partition the solar and long wave radiation between the underlying soil and the leaves, the larger the LAI the smaller the amount of sunlight reaching the underlying soil surface. (This partition function also depends on solar zenith angle, the lower the sun the less underlying bare soil 'visible' to the sun.) Leaf area index is also used to scale the leaf fluxes to a horizontal surface for a plant canopy. As we have said, this scaling problem is presently a very difficult one to deal with; it will be discussed briefly later on.

Stomatal resistance

Stomatal resistance is a measure of the aperture size of the stomates. As such, the stomatal resistance governs the flow of water vapor through the stomates. Since there are thousands of stomates on a leaf, the individual resistance for all the stomates are added together in parallel (the inverse of the sum of the inverse resistances for each stomate) to equal the average stomatal resistance for the leaf.

For most plants (the so-called *C3 and C4 plants*, but not for *CAM plants* which transpire at night), stomatal resistance is thought to be a function of several variables as expressed in Eqn. 7.2:

- The amount of absorbed sunlight, more precisely the amount of ***photosynthetically active radiation (PAR)***.
- The substrate and/or **leaf water status**,
- The **vapor pressure deficit** for the leaf (the difference in vapor pressure between the inside and the outside of the leaf)
- The ambient carbon dioxide concentration
- The leaf temperature.

Let us treat each of the above factors one at a time.

Solar flux

Stomatal resistance decreases with increasing solar flux. Since the PAR is roughly one-half the total solar flux, stomatal resistance will tend to decrease with time during the morning and increase with time in the afternoon. Various studies show that the sensitivity of stomatal resistance to light intensity is greatest at low light intensity and effectively vanishes as the light becomes stronger. Alternately stated, the leaf is no longer light sensitive above some threshold light intensity. This threshold is typically about 50-200 Wm-2 for the total solar flux, which is roughly the amount received at midday beneath a mostly overcast sky at mid latitudes in summer. Thus, the curve of stomatal resistance versus time tends to reach a minimum during the middle part of the day, but it is relatively flat for several hours around noon when the sun is shining.

Leaf or root zone water status

Stomatal resistance tends to increase as water becomes more limited. This seems reasonable since the plant is trying to protect itself from water loss. In this context, the plant really does exercise its judgment, but it does so in such an obscure fashion that plant scientists are pulling each other's beards arguing about how the plant does it! The prevailing opinion as of 1990 was that the water content in the root zone is the important governing parameter. In the words of Jones (1983), 'There is now strong circumstantial evidence that the plant growth regulator (a hormone produced by the plant) *abscisic acid* or ABA is involved in mediating most responses to water stress.' By this statement we mean that abscisic acid is passed from the root zone to the leaves to stimulate stomatal opening and closing. In short plants, this hormone is transferred in the sap in a matter of minutes or at most an hour or so between the roots and the leaves. In trees, this process takes hours.

All the votes have not yet been counted in the contentious issue of whether leaves or roots control the stomatal opening. Before about 1985, most scientists were willing to believe that the control of stomatal resistance lay in the leaf water status. Later, some scientists became convinced that the stomates perceive water deprivation only through messages from the roots in the form of ABA. Others felt and continued to feel that the leaf water status controls the stomatal opening, at least on a short term, since a transient and partial stomatal closure is often reported to occur near midday, (*midday stomatal closure*) even when the soil moisture content is not changing very rapidly with time. Recently there has been an attempt by a plant scientist named Tardieu to wed these two ideas, root zone control by ABA and leaf control by sensing the leaf water content.

For now, we emphasize the leaf water status control, although it seems probable that both root and leaf have a say in the matter. When you run the simulations with this mechanism of leaf control in effect, you will find out how easy it is to surmise that the roots control the stomatal resistance, as the root zone soil water content still plays a role in the stomatal resistance, though by indirect means even though no direct relationship exists in the model. A secondary option (called the Deardorff formulation) allows the stomatal resistance to vary with soil water content. You can choose the Deardorff formulation in the model for calculating stomatal resistance in the model, if you like. But keep in mind, in choosing the formulation you also choose sides in this controversy.

Vapor pressure deficit

Equally understood, is how vapor pressure deficit (VPD) affects the stomatal resistance. Since vapor pressure (e) is closely proportional to specific humidity, let us write Eqn. 7.1 in terms of vapor pressure decrease from the leaf $e_s(T_l)$ to that in the interleaf air space (e_{af}). This difference we will define as the *vapor pressure deficit*, although the latter can be defined and is defined elsewhere in terms of other pairs of vapor pressures (e.g., between the interior of the canopy and the atmosphere above). All that is really accepted by plant scientists is that the stomatal resistance tends to increase with increasing vapor pressure deficit, $e_s(T_l) - e_{af}$. How the plant accomplishes this adjustment remains a mystery. The sensitivity of stomatal resistance to VPD is highly variable from one type of plant to another, but it is especially significant in beans and trees.

Again, it would seem that the plant has a say in how much water is lost, since r_l increases in response to an increasing VPD, which constitutes the numerator of Eqn. 7.3. Note that the numerator, however, represents the demand of the atmosphere for evaporation, whereas the VPD effect on r_l, which operates through the denominator, represents a compensation for the demand. Since both the numerator and denominator of Eqn. 7.3 increase with increasing VPD, the net effect may be that transpiration does not increase *monotonically* (one increasing as the other increases) with the numerator in this equation. It is possible for the vapor flux to increase or decrease or first increase and subsequently decrease with increasing VPD. The latter behaviour is sometimes referred to as the '**feed forward**' process, so named by Farquhar (1978). Later, we will address the question of feedback between vapor pressure deficit and stomatal resistance.

Some plants are much more sensitive to VPD than other plants. Soybeans, for example, have a large sensitivity of stomatal resistance to VPD. Corn, on the other hand, has a relatively small sensitivity to VPD. For corn, we should not expect to find a braking effect on r_l due to increasing

VPD. Thus, we have made the sensitivity of stomatal resistance in corn to VPD equal to zero. (You will later have the option in the simulations to choose soybeans, which would have a significant sensitivity to VPD.)

Carbon dioxide concentration and leaf temperature

The remaining two factors in Eqn. 7.2, carbon dioxide concentration and leaf temperature, are relatively unimportant for normal atmospheric fluctuations. Carbon dioxide concentration is increasing with time over many decades and may double before the end of the 21st century. Currently (in 2020), the CO2 concentration in the atmosphere is over 410 parts per million by volume (ppmv), compared with 360 ppmv during the 1990s and about 300 ppmv at the beginning of the 20th century. While we may want to simulate both the CO_2 fluxes and the effect of increased ambient carbon dioxide concentration on stomatal resistance later in the course, the present level of carbon dioxide concentration can be treated as a constant, about 410 parts per million by volume (ppmv). That value varies by less than 10% above the canopy during a season.

Finally, stomatal resistance is not insensitive to leaf temperature especially at the boundaries of tolerance, *e.g.*, 5 and 35 C, beyond which the plant tends to shut down. Some measurements suggest that, up to a point, an increase in temperature favors an increase in photosynthesis. We will assume that the plant is growing within the bounds of temperature tolerance and ignore the effect of temperature within the tolerance range, as the effect is not well studied.

There are about as many stomatal resistance formulas as individuals who have tried to model the stomatal resistance. Jarvis (1976) suggests one such formulation. The Jarvis equation, expressed by Eqn. 7.2, simply equates the stomatal resistance to the product of various functions. Let us consider just three of them in Eqn. 7.2: one representing solar flux f(S), one representing leaf potential $f(\Psi_l)$ or soil water content f(w) (soil water potential is an option in some models), and one representing the effect of VPD. We favor the Jarvis formulation using the leaf water status factor $f(\Psi_l)$; the Deardorff equation uses f(w). These functions vary from 1.0 to infinity. Although a full discussion of stomatal resistance lies outside the scope of this course, the relationship between r_s and VPD, as we formulate it, is somewhat more complex (and indirect) than indicated by Eqn. 7.2.

The equation also contains a factor r_{smin} , which is the stomatal resistance when the leaf or soil is at full water capacity, at full sunlight and when the vapor pressure deficit effect is negligible. This type of parameter is called the *minimum stomatal resistance*, although it is somewhat of an elusive parameter, even in our model. Minimum stomatal resistance is a somewhat ambiguous term, as discussed in the literature and sometimes without much thought. We might think of this parameter loosely as the value of r_s on the sunlit side of a leaf at noon on a bright sunny day when the plant is engorged with water ($f(\Psi_l)$ and f(S) = 1.0) (this still begs the issue of the vapor pressure deficit function f(VPD)). Typical values for minimum stomatal resistance vary considerably; *e.g.*, from about 25-50 sm^{-1} for crops such as corn and soybeans to 200-300 sm^{-1} for many types of trees. Korner (1979) furnishes a huge table of stomatal conductances for different plants; recall that a conductance is simply the inverse of resistance. Some people refer to the minimum stomatal resistance as that minimum value near midday; the latter is a convenient quantity for comparing day-to-day fluctuations in stomatal resistance. It is not clear if

Korner's numbers represent some kind of theoretical minimum stomatal resistance or simply a daily average under unspecified conditions.

Simulations

Level 1

Let us look at the variation of stomatal resistance for various simulations.

Simulation series 1

- **Observe the sensitivity of stomatal resistance to solar radiation**

First, plot r_s (stomatal resistance, STMR in the model output) as a function of time for one day. Use the default corn canopy parameters from Chapter 6 with a LAI of 7. This will give you a sense of how r_s varies with solar intensity throughout the day. To see the solar effect more clearly, you can simulate a cloudy day by simply changing the cloud factor.

Questions

1. Equate a low stomatal resistance with a large stomatal opening. Are the stomates opened up longer on sunny or cloudy days?
2. Does the amount of solar radiation on the leaves affect the nighttime stomatal resistance?
3. What is the daily minimum stomatal resistance?
4. How does stomatal resistance change under cloudy conditions?

Simulation series 2

- **Observe the sensitivity of stomatal resistance to root zone soil moisture**

Change the vegetation parameterization to Deardorff (set "stmtype" to "D" in the model input file) and assign a wilting point value of 0.12 ("WILT" in the model input file) instead of 0.08. (The official wilting point for the soil is about 0.12 soil water content by volume, equivalent to a moisture availability of about 0.4; field capacity is 0.34 by volume.) The vegetation fraction should still be set to 100 percent. Make simulations with 3 different values of the root zone moisture availability M_r (FSUB in the model input file) equal to the standard value of 0.50 used in the first chapter, and also 0.9 and 0.4. This will show you how r_s varies with root zone soil water content. Try decreasing M_o (F in the model input file) well below 0.4 for the soil surface while keeping the root zone (M_r) moist. Note that the deep layer M_r influences the transpiration only via the stomatal resistance function and therefore is not equivalent to the ratio of evaporation to potential evaporation, as in the bare soil case.

Questions

1. How does the minimum stomatal resistance change with varying root zone moisture? Is it as you would have guessed?

2. Does the surface layer moisture have a limiting effect on the stomatal resistance at all? Remember, you have just done simulations using the idea that the root zone moisture controls the stomatal openings (you took sides!), and other (Carlson/Lynn) parameterizations have not seen a change in stomatal resistance due to varying moisture availability in the soil.

Simulation series 3

Returning to the standard (Carlson/Lynn) stomatal resistance option (set stmtype to ”L”) and keeping M_r above 0.4, lower the LAI from 7 down to 5, and then to 3 or 2. Next, change the minimum stomatal resistance from 50 to 100 s m^{-1}. For each of these cases, plot the stomatal resistance versus the vapor pressure deficit.

Questions

1. What happens to the stomatal resistance and to the fluxes as the canopy becomes thinner?
2. Can you explain why the plant is conserving water under these conditions?
3. Does the vapor pressure deficit relationship make sense in light of the material presented in this chapter?

Level 2

Equation 7.2 tells us that the stomatal resistance, and therefore the latent heat flux from a vegetation canopy, depends on up to 5 physical variables. In level 1, we looked at how r_s changes due to each variable independently. We know, however, that as one of these variables changes, another can change as a result and counteract the effect of the original variable on the stomate. For example, if the solar radiation is increased, the stomatal resistance will lower, and more transpiration will take place. However, as more water vapor evaporates into the canopy, the VPD will lower as well, tending to increase the stomatal resistance and decrease the transpiration rate. This type of negative feedback can be seen in a variety of ways between these 5 variables, in particular the first three in the above list. Think about the interaction between these variables, and the overall effect on stomatal resistance.

Questions

1. Can you tell which of these 3 variables has the largest overall control on transpiration and should be given the most consideration?
2. As you vary the root zone soil water content M_r does the stomatal resistance change? If so, how does this occur when the model makes no explicit relationship between stomatal resistance and soil water content?
3. Vary the wind speed by doubling values near the surface. Does this affect stomatal resistance? If so, would we be justified in adding another component to the stomatal resistance equation, that of wind speed?

4. Try fooling around with the leaf water status (Carlson/Lynn) parameterization. Does that theory place more control in the hands of the leaf water status than the Deardorff places on the root zone moisture? (Be careful as you lower the value of M_r; avoid reducing it below a value of 0.4.)
5. Try to get an overall feeling how each of these variables impact stomatal resistance, and then relate that to how they impact surface fluxes and eventually temperatures. In other words, don't lose sight of the big picture just because we are looking at microscopic features on leaves!

Definitions

(r_s) = Bulk stomatal resistance for the leaf canopy
(g_s) = Stomatal conductance for the leaf

Equations

[Eqn 7.1]

$$T_r = \rho Le \frac{q_s(T)_1 - q_{af}}{r_1 + r_{af}} \quad r_1 \cong r_s$$

[Eqn 7.2]

$$r_s = r_{smin} f(S) \cdot \{f(\Psi_1) \text{ or } f(w)\} f(VPD) \cdot f(CO_2) \cdot f(T_1)$$

[Eqn 7.3]

$$T_r \propto \frac{e_s(T_1) - e_{af}}{r_1 + r_{af}}$$

Terms introduced in this chapter

Terms to look up and remember
abaxial surface
abscisic acid ABA
adaxial surface
biomass
cuticular resistance
epidermis
evapotranspiration
feed forward process
guard cells
intercellular air spaces
interleaf air spaces
LAI
leaf area index
leaf boundary layer resistance
leaf resistance

leaf water potential
leaf water status
mesophyll
midday stomatal closure
minimum stomatal resistance
PAR
photosynthetically active radiation
root zone
stomatal resistance
substomatal cavities
vapor pressure deficit
vapor pressure deficit
VPD

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Simsphere Workbook: Chapter 8

The Vertical Structure of the Plant Canopy

Introduction

Now that we are somewhat familiar with the concept of a vegetation canopy, let us drop into that canopy to examine its microclimate. In so doing, we shall introduce two new ideas: The first, which is non-controversial, is that the structure and density of the plant canopy can influence the plant microclimate. The second, not accepted by all scientists, is that an ensemble of plants can compensate for increased atmospheric demand for water vapor by modifying its canopy environment. In the next two chapters, we shall examine some possible **negative feedback** mechanisms that allow the plant to compensate for excessive atmospheric demand or even for a change in its own physiological makeup. We shall begin with the vertical structure of the temperature, wind, and water vapor content inside the plant canopy and try to understand what factors are important in affecting the canopy climate.

Before exploring the consequences for the internal feedback mechanisms in the plant microenvironment, we would like to present an abridged discourse on the subject of photosynthesis, courtesy of an old biology textbook (Hardin, 1966; pp 231-232). We feel that some background information on this plant process is useful, if not essential, for understanding this and subsequent scenarios. Any decent biology book should provide the same information.

Background

Photosynthesis

Before the early eighteenth century, scientists believed that plants received all their nourishment from the soil. Later, people understood that their nourishment was derived from the atmosphere and that light was an integral part in the process. A Swiss minister named Senebier showed late in the 18th century that CO_2 + H_2O + light energy yields O_2 + plant material. Thus, light and water are necessary to convert carbon dioxide into sustenance for the plant.

Later, investigators realized that the material being formed in the plant consisted of carbohydrates, whose chemical formulae contains combinations of carbon and hydrogen atoms. The production of carbohydrates, however, occurs only within the green areas of the leaves, in cellular bodies called *chloroplasts*. The chloroplasts contain a chemical compound called *chlorophyll*, which exists in a variety of similar forms. Chlorophyll is required if photosynthesis is to take place.

If a chlorophyll molecule is exposed to light, excited electrons are captured and used to synthesize molecules, such as *adenosine tri phosphate (ATP)*, which is used 'to fuel' chemical reactions in the plant. (Incidentally, ATP is also used in human cells and is an essential, though limited, fuel burned within the muscles that enables athletes to achieve high performance.) A second series of reaction involves the splitting of water molecules, a process aided by the presence of the chlorophyll where a compound nicotinamide adenine dinucleotide (NADPH) is the end product. These reactions are part of the *light reactions* of photosynthesis. Rather than become ensconced in details you should simply consider the role of these compounds as sources of energy along the pathways in the synthesis of carbohydrates.

The photosynthetic process is not fully realized until the amount of organic carbon in the cell is increased. This is done by coupling CO_2 (inorganic carbon) with organic compounds, reactions called *dark reactions* (because they do not require sunlight), which produce a variety of constituents in the plant cell (essentially carbohydrates). Photosynthesis follows three pathways: C- 3, C- 4 and CAM (crassulacean acid metabolism).

Although plants generally intake carbon dioxide and exhale oxygen, some reactions liberate carbon dioxide, and it is passed back into the atmosphere. This exhalation is called **respiration**; respiration that occurs without the aid of sunlight is called *dark respiration*. Generally, the amount of carbon dioxide respired to the atmosphere is small compared to that taken up by the plant. However, dark respiration can occur at night, resulting in a rise in ambient carbon dioxide concentration in the plant canopy. An even more important source of carbon dioxide in the plant canopy is produced as a by-product when dead plant material decays. Since the latter type of reaction can also occur at night, the night time concentrations of carbon dioxide inside the canopy sometimes rise to levels above normal daytime concentrations, as shown for the CO_2 night time profile in Figure 8.1. We will have more to say about this in a later scenario.

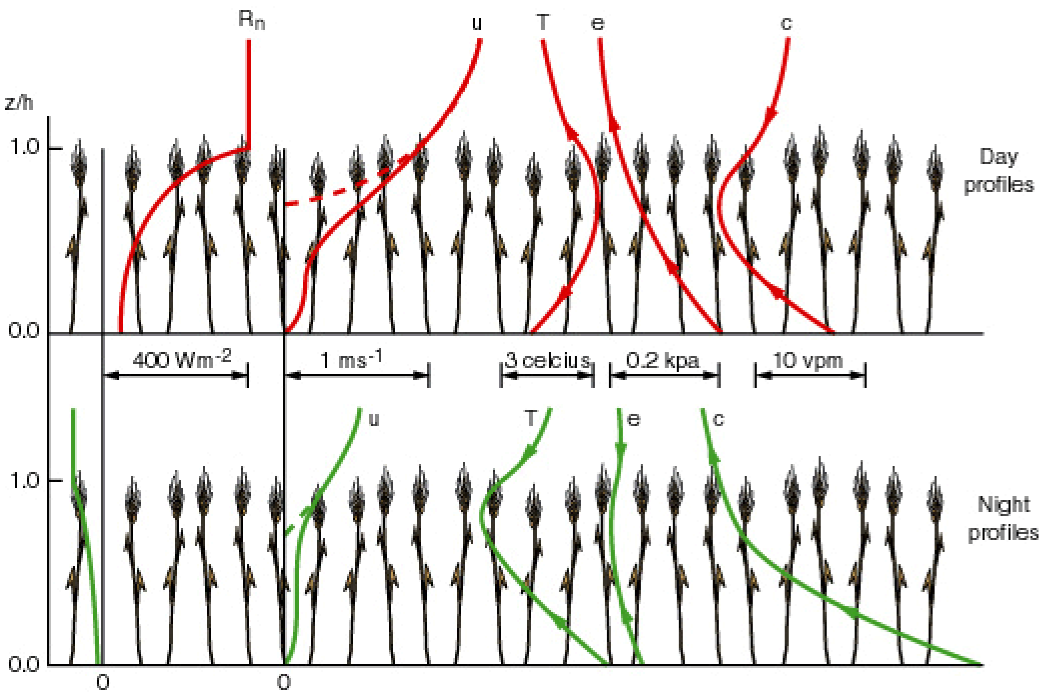

Figure 8.1. Idealized profiles of net radiation (R_n), windspeed (u), air temperature (T), vapor pressure (e) and CO_2 concentration (c) in a field crop growing to a height h plotted as a function of z/h, where z is the height above the surface and h is the height of the canopy. The pecked wind profiles represent an extrapolation of the logarithmic relations between u and (z-d) above the canopy, where d is the displacement depth; (Freely adapted from Monteith, 1990; p. 232). The top part of the figure pertains to typical daytime profiles, the bottom part to night profiles.

Vertical structure of temperature, wind speed and humidity in a plant canopy

It doesn't take a meteorologist to realize that wind speed increases with height and that temperature decreases with height away from a surface heated by the sun, whereas at night, the temperature increases with height away from a surface through radiative cooling. However, let's consider what happens inside a plant canopy during the day and at night.

Windspeed

If we were to plot time-averaged wind speed versus the log of height, the profile would tend to be a straight line. In linear space, this distribution of wind speed is shown by the dashed lines on the wind speed profiles in Figure 8.1. The height at which the time-averaged wind speed extrapolates to a zero wind speed is called the *roughness height*, a parameter you may have already encountered. The higher the vegetation, the larger the roughness height. In general, agronomists take the roughness height to be about 0.1 the height of the vegetation, for a uniformly distributed vegetation canopy. Agronomists also refer to a *displacement height* (h), which is the height above the ground surface that the logarithmic wind profile 'sees' as the ground surface; this is typically about two-thirds the height of a uniform vegetation canopy. Therefore, to calculate at what level the roughness height of a 1 meter canopy is, you must first account for what the atmosphere sees as the ground surface, which would be 2/3 of 1 meter, for a displacement height of .667 meters. Then, the roughness height, which is .10 of 1 meter, would correspond to .10 meters above the .667 level, or .777 meters above the actual ground surface. Keep in mind that the actual roughness height is still .10 meters, but with respect to what the atmosphere sees as the ground, which is now .667 meters. (Note: although we talk of roughness and displacement heights in the context of vegetation, these parameters are generally applicable to any type of surface and the roughness height requires specification in the model.)

You must take care in interpreting roughness height. First, you should understand that the parameter, while related to the physical surface, is only a constant of integration in the log profile law. Second, it is possible for two or more roughness lengths to exist at one surface location, but each roughness length may apply to a different scale. Third, the roughness of a sparse crop canopy can be greater than that of a dense canopy because the latter tends to appear as a smoother surface to the ambient wind. Thus, a very dense canopy of soybeans may appear to the atmosphere as a fairly smooth surface, whose roughness elements consist only of the leaves protruding above a relatively uniform surface, whereas a canopy of sagebrush with the same height as the soybeans may appear as a much rougher surface to the atmosphere.

Except for the concept of the *friction velocity*, the logarithmic profile laws, such as those used in this model, do not apply immediately above or inside a vegetation canopy. Therefore, we need not worry about displacement height, except for the fact that Monteith (1990; p. 232) refers to it in his discussion of Figure 8.1; taken from his book. In actuality, the wind speed does not quite go to zero inside the canopy; instead it reaches a relatively small value in the upper third of the vegetation canopy and then decreases very slowly with decreasing height; this is illustrated by the solid curves in the wind speed profiles in Figure 8.1. The value of the wind speed in the slowly decreasing regime is nevertheless a function of the so-called friction velocity, which is a function of the wind speed above the canopy (at the top of the surface layer) and the roughness height.

Temperature

During the day, temperature is maximized at the earth's surface while during the night it is minimized. This is also true for vegetation canopies, although the maximum temperature tends to occur near the top of the vegetation, as shown in Figure 8.1. This is due to shading of the underlying soil, as the soil surface temperature tends to be cooler than the canopy temperature (or that just above the canopy) during the day.

Up to this point, you have plotted the 'surface radiant' and 'surface air temperatures' in your scenarios. Surface radiant temperature is relatively easy to imagine for a smooth, bare soil surface, but for vegetation, the surface temperature becomes an abstraction. As we define it, it is the equivalent temperature of the ground, *i.e.*, the temperature corresponding to the amount of upwelling long wave (thermal) radiation from the surface. Another important temperature that the model can calculate is the leaf temperature (T_l), which is that appropriate to a sunlit leaf near the top of the canopy. The interleaf air temperature (T_{af}) is calculated as a function of this temperature, the air temperature above the canopy, the ground temperature, as well as being a function of various air resistances. We can, therefore, take account of six temperatures in the vegetation scenarios: the interleaf air space temperature (T_{af}), the ground surface temperature (T_g), the temperature just above the canopy (i.e. at screen height; T_a), the radiometric surface temperature (T_o), the leaf temperature (Tl), and the temperature at the top of the surface layer (50 m). (Let us not confuse these temperatures with potential temperature.)

Simulations

Level 1

Simulation series 1

Vary the relative humidity near the surface

a) Run the default vegetation case for the corn canopy (LAI = 3, vegetation height = 2 m, fractional vegetation cover FRVEG = 100%, roughness height ZO = .20 m), but this time note the distribution of wind speed, temperature and specific humidity (vapor pressure) inside and above the canopy. The values inside the canopy (the interleaf air spaces) are WFOL, TFOL and SHFOL in the output file. The values at anemometer level (10 m) are W10, T10 and SH10, and the values at the top of the surface layer (50 m) are W50, T50 and SH50. In addition, note the vapor pressure deficit (VPDEF) in the canopy.

b) Now, change the relative humidity in the initial temperature-dew-point sounding. Lower the relative humidity (increase the vapor pressure deficit) in the lower layers by increasing the **dew point depression** (DEP in the model input file) by some factor (say 2.0 C) in the lowest three levels. Then increase the relative humidity (decrease the vapor pressure deficit) by increasing the dewpoint depression from its original value in the lowest three levels. Compare the same temperature, moisture, and wind speeds as above. Note how the specific humidity and the vapor pressure deficit change at 50 meters and in the canopy with this change in external relative humidity. Note that the formula (Eqn 3.1 for the calculation of latent heat flux) suggests that doubling the numerator (proportional to the temperature-dewpoint difference and ultimately to the VPD) should cause evapotranspiration to increase by a factor of two. Here we ask ourselves the question: What would the plant do in order to avoid the increased demand on its precious moisture?

Questions

1. Are the results at all levels what you expected due to an increased atmospheric demand for water vapor?
2. How does the change in evapotranspiration compare to the change in the VPD? Hint: Evapotranspiration (the sum of the transpiration and surface evaporation, see chapter 7) is the latent heat flux LE scaled by the latent heat of evaporation.
3. In this circumstance, does it appear that the corn canopy has decoupled wholly or partially from the atmosphere above?

Simulation series 2

Vary the resistances inside and outside the canopy

Keep the same graphs up from simulation 1 to compare to those generated here.

a) Using the same dew point depressions as simulation 1, reduce the wind speeds by a factor of two at all levels in the sounding. This effectively increases surface layer resistances. As you have already discovered in an earlier scenario for bare soil, decreasing surface roughness also causes an increase in the surface layer resistances. (Alternately, you can produce the same effect by decreasing the surface roughness (ZO) by, say, a factor of 10 or 20; if you wish to do this instead of changing the wind feel free to do so). This will increase all resistances external to the plant, including the leaf boundary layer resistance (r_{af}) and the resistance between the canopy and the surface layer.

b) Next, reduce the leaf width (WIDTH in the model input file) by a factor of four from the default 0.12 to 0.03 m; this will decrease the leaf boundary layer resistance by about a factor of two but not affect the other resistances to heat and water vapor. What we have effectively done here is increased the resistances outside the plant canopy while lowering those within. According to McNaughton and Jarvis, we are decreasing the ratio of leaf stomatal to boundary layer resistance as well as decreasing the ratio of canopy to surface layer resistance. Think about what has happened to the amount of mixing within the canopy compared to above.

Now, compare the relevant parameters, especially humidity, and how they change compared to simulation 1 above. See if the changes in canopy temperature and humidity are as extreme as when you reduced the relative humidity in the higher wind regime and with the larger leaf size (the base case).

Questions

1. Do you find any evidence of buffering? (By this,we mean to ask whether changes imposed in the environment seem to be attenuated within the plant canopy.)
2. How might the leaf size affect the energy budget of the plants? Would it make sense for different plants to have different energy consumption strategies based on leaf size?
3. What happens to evapotranspiration over a plant canopy on a windy day versus a day with light winds?

Level 2

a) Find optimal conditions for canopy decoupling

Play around with the wind speed (or roughness length), leaf size, and initial relative humidity on the sounding to see whether you can find optimum buffering, *i.e.*, a situation in which a change in ambient relative humidity is not reflected (or weakly reflected) by changes in the canopy relative humidity due to internal feedback. If you do find it, reduce the LAI and note how rapidly the effect diminishes with decreasing canopy density. Try to get a feel for how dense our corn canopy must be to effectively decouple from the atmosphere above under ideal conditions. We are not sure whether such a buffering (as predicted by Jarvis and McNaughton, 1986 and later by McNaughton and Jarvis, 1991) can actually be found in the model, or whether it even occurs in nature, but we are very curious to learn under which conditions auto-humidification can be induced. Good hunting!

b) Vary the type and characteristics of the vegetation canopy

So far, we have been examining variations in our default corn canopy. Now, try changing the type and characteristics of the vegetation canopy itself. For example, change the crop type, canopy height (and roughness length), fractional vegetation cover, etc. See if the 'optimal conditions' for decoupling you found in (a) also work for other types of vegetation, or are there new optimal conditions for each variation. Be creative but only change one variable at a time to avoid confusion. See if there is also an optimal canopy height, crop type, etc. for which the canopy decouples (is buffered from atmospheric or other external changes) most readily. You may also want to view the effects of changing these parameters on the surface radiant temperature, because they all have a direct impact on the amount of sunlight that can penetrate to the soil surface.

Definitions

CO_2 + H_2O + light energy yields O_2 + plant material
(ATP) = Adenosine triphosphate
(NADPH) = Nicotinamide adenine dinucleotide
CAM = Crassulacean acid metabolism
(h) = Displacement height
(T_l) = Leaf temperature
(T_{af}) = Interleaf air temperature
(T_g) = Ground surface temperature
(T_a) = Temperature just above the canopy, or at screen height
(T_o) = Radiometric surface temperature
(PS) = Shelter factor
(VPD) = Vapor pressure deficit
(r_s) = Stomatal resistance
(r_a) = Boundary layer resistance
(r_{af}) = Leaf boundary layer resistance
(R_n) = Net radiation
(u) = Windspeed
(T) = Air temperature
(e) = Vapor pressure
(c) = CO_2 concentration

Terms introduced in this chapter

Terms to look up and remember
feedback
chlorophyll
chloroplast
ATP
NADPH
C- 3, C- 4 and CAM
respiration
dark respiration
light reaction
dark reactions
friction velocity
displacement height
auto humidification (our term)
dewpoint depression

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Appendix

The clear message from the above arguments is that the importance of stomatal control of transpiration depends on the amount of environmental feedback. At the leaf and canopy scales, the gain of the negative feedback loops describing temperature and humidity gradients through the leaf boundary layers and atmospheric surface layer depends on the ratios of leaf stomatal to boundary-layer conductances and the canopy to surface-layer conductances. When these ratios are large, we need only crude models of stomatal responses, but when they are small, we need much better ones. How the stomata respond to leaf-surface saturation deficit will not matter much for a canopy of small aerodynamic roughness and large canopy conductance because the saturation deficit close to the leaves will not change much and such stomatal responses as do occur will have only a small effect on transpiration anyway.

At regional scale, the force of these arguments increases, and even the leaf and surface-layer aerodynamic conductances have little effect when conductances are large (resistances small) because then the dominant negative feedback is by way of processes in the whole planetary boundary layer. A corollary is that there is no practical justification for using any canopy model more complex than the 'big- leaf' model at this scale for areas dominated by highly productive agricultural crops. This is doubly true given the difficulties of average canopy conductance across a mosaic of different fields.

Simsphere Workbook: Chapter 9

Vapor Pressure Feedback Effects On Plants

Introduction

As in real life, things get more and more complex with time. Up until now, we have allowed the stomatal resistance to vary only with solar intensity and leaf water potential (an indirect function of soil water potential). So far, however, the leaf water potential effect would have been weak because we have not tried to stress the plant. In addition, we have not allowed the stomatal resistance to vary with vapor pressure deficit, although such is the case for many plants. This is reasonable, so far, in that our previous chapters have treated the case for corn which we know exhibits a relatively weak relationship between stomatal resistance and vapor pressure deficit. In this chapter, we will allow stomatal resistance to vary with vapor pressure deficit *(VPD)*, although in so doing, we could end up creating a somewhat anomalous corn plant in the computer (sort of a cross between corn and soybeans). What happens inside the vegetation canopy when we make stomatal resistance a function of VPD? Here is a philosophical question: Why do some plants make use of the VPD *feedforward* effect and others not as much? In exploring this issue we dig deeper into the can of worms labelled 'crop modelling'.

Background

Before starting this discussion let us define what we mean by vapor pressure deficit. VPD can have many definitions, the vapor pressure difference between the vapor pressure value at air temperature saturation (say at screen (1.3 m) level) and that of the air at that level. It could also refer to the difference between the value at the saturation temperature of the canopy or the leaves and that in the atmosphere, either above or inside the plant canopy. We will take VPD generally to mean the difference between the saturation vapor pressure at the leaf surface and that in the immediate surroundings of the leaf, say just beyond the leaf boundary layer.

As stated in Chapter 8, many plants (such as soybeans and various species of trees) tend to increase their stomatal resistance in response to an increase in vapor pressure deficit (Jones, 1983; pp. 120-128). Figure 9.1 shows that variations in stomatal conductance as a function of vapor pressure deficit differ from one type of plant to another; (note that 1 kPa is equal to 10 mb). Clearly, plants that have the capability of increasing their stomatal resistance with increasing VPD enjoy an advantage in dry climates or in climates that fluctuate - would you agree?

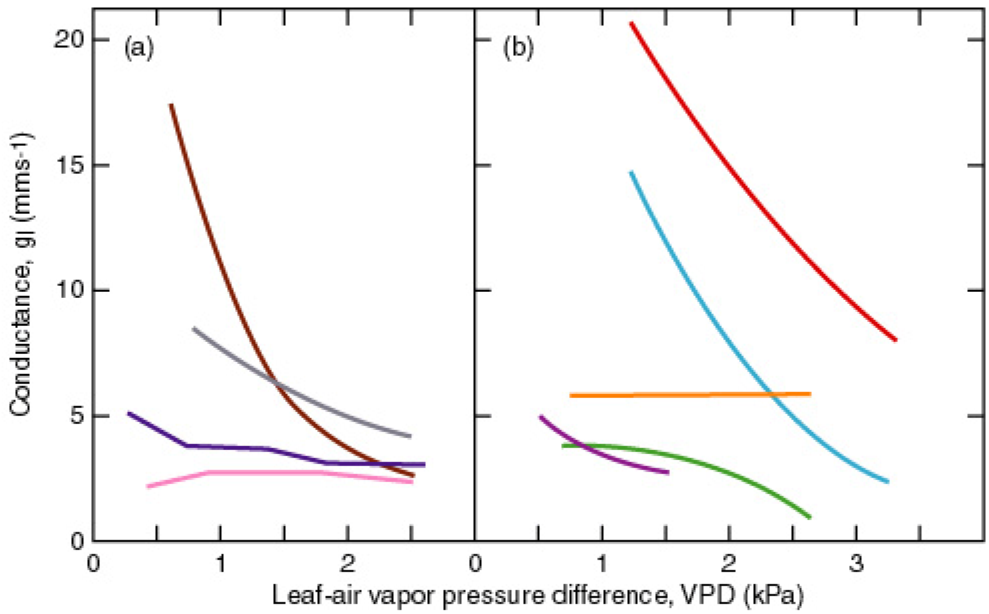

Figure 9.1. Examples of stomatal humidity responses (VPD) for different types of plants. (Freely adapted from Jones, 1983; page 120).

Consider the typical evolution of vapor pressure deficit during the day: As leaves become warmer in the sunlight, the saturation vapor pressure of the leaves increases. Because of the exponential increase in saturation vapor pressure with increasing temperature, vapor pressure in the sub-stomatal cavities is also exponential with increasing temperature. Unless the transpiration can somehow sustain an increase in interleaf vapor pressure equal to that inside the leaf, the vapor pressure deficit between leaf and interleaf air spaces will increase until sometime in the afternoon, when the leaf temperature begins to decrease. Obviously, the lower the leaf temperature, the less will be the increase in VPD and the less the water loss. The plant has a problem here; how to keep its leaf temperature low enough not to cause excessive water loss by having a large vapor pressure deficit. If it tries to reduce transpiration by increasing its stomatal resistance, its leaf temperature will increase and so will the VPD.

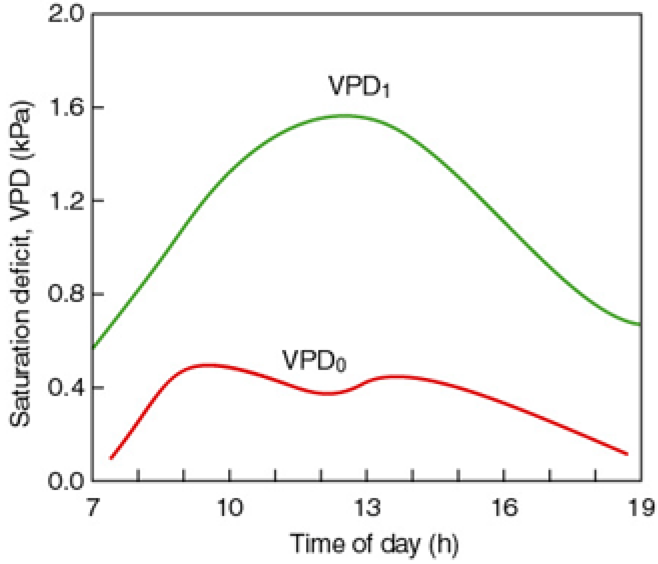

Figure 9.2. The time course of saturation deficit at the leaf surface ($VPDO_0$ - 1 kPA= 10 mb) and above the canopy (VPD_1) for a sugar cane crop in Hawaii (Freely adapted from McNaughton and Jarvis, 1991; page 283).

Allowing stomatal resistance to increase with increasing VPD will retard the typical decrease in stomatal resistance during the first part of the morning (in response to increasing sunlight) and even cause it to increase temporarily for a period during the middle portion of the day, resulting in a minor midday maximum (Figure 9.2, McNaughton and Jarvis 1991; their figure 9). Recent experimental evidence suggests that photosynthesis increases with increasing leaf temperature up to some value, above which the leaf begins to shut down its stomates. This suggests that stomatal resistance tends to decrease up to this temperature because it is in the plant's interest to maximize its growth. Now, we have said that increasing the resistance (or decreasing the conductance) of the stomates retards the transpiration, increases the sensible heat flux, the leaf temperature and, accordingly, the VPD, thereby retarding any decrease in transpiration. This situation constitutes a positive feedback gain between leaf temperature, VPD and stomatal resistance, as discussed in an earlier handout on feedback processes. This type of response is shown in Figure 9.3 (Korner 1985, his figure 3).

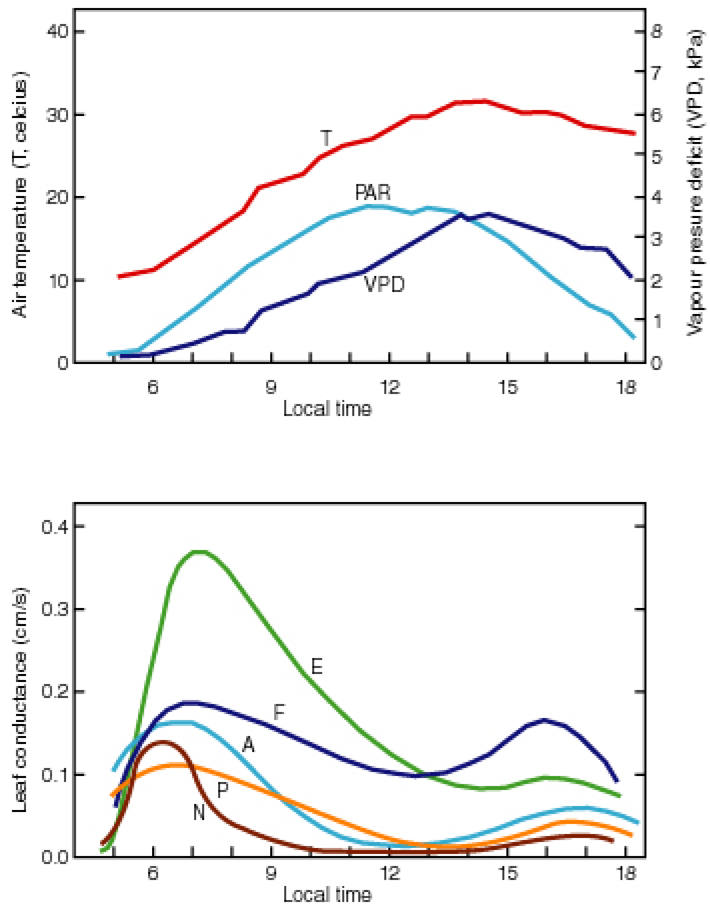

Figure 9.3. (top) Variation of air temperature (T), photosynthetically active solar radiation PAR (PAR) and vapor pressure deficit VPD (VPD) and leaf conductance versus local time and (bottom) for 5 different species of plants for a very hot day in summer with high vapor pressure deficits. Note that all species show a rapid decline in leaf conductance without recovery after about 7 - 8 o'clock in the morning. (Freely adapted from Korner, 1985; Figure 3)

This type of feedback, called feedforward (see previous scenario for introduction of the term), is discussed by Jones (1983; pp 125 - 128). Feedforward is distinguished from ordinary feedback between VPD and stomatal resistance in that a change in stomatal resistance (or conductance) affects the heat fluxes and therefore the transpiration. Ordinary feedback between VPD and transpiration causes an increase in transpiration with increasing VPD. If the increase in VPD also causes an increase in stomatal resistance (the exact mechanism for this is still being debated), but not enough to reduce the transpiration, this constitutes negative feedback. In feedforward, an increase in VPD causes an increase in stomatal resistance, and a decrease in transpiration. In other words, the process of transpiration increasing during drier conditions is reversed so that lower humidity will now result in less transpiration as well. (Sometimes in human activities, such as in competitive foot or bicycle racing, excessive loss of water may cause the sweat mechanism to go haywire and cease, resulting in a dangerous overheating of the body.)

Feedforward is therefore a more direct and efficient mechanism (having one fewer step in the process) for retarding transpiration. Unlike ordinary feedback, it can actually lead to a reduction in transpiration with increasing VPD. This is illustrated in Figure 9.4 (from Jones, 1983; his figure 6.14). Note that the rapid decrease in stomatal conductance with increasing vapor pressure deficit for the apple tree (Figure 9.1; M7) causes a reversal in the slope of the transpiration curve above VPD = 0.9 kPa (9 mb) in Figure 9.3.

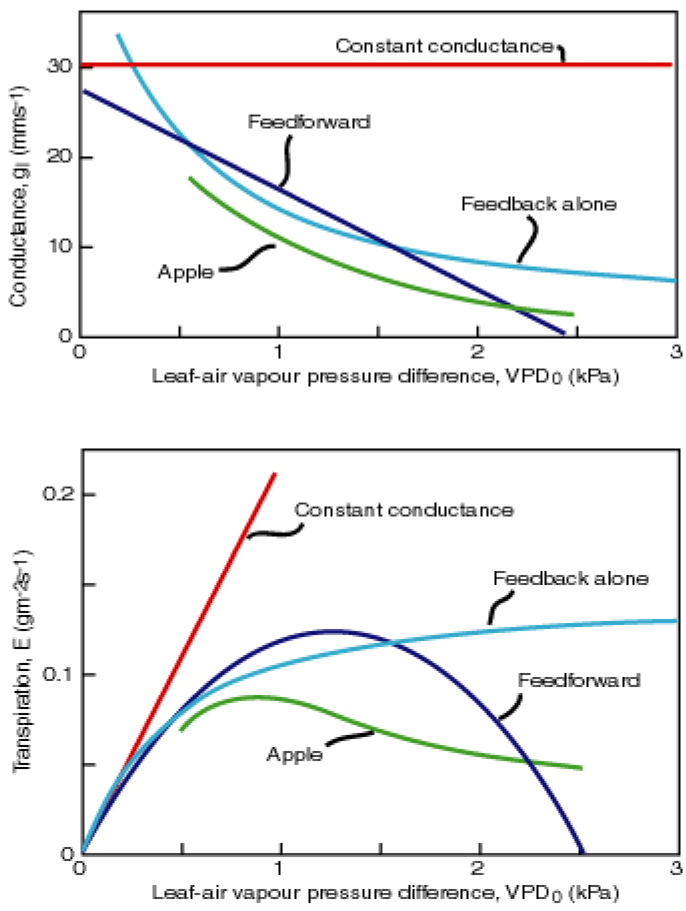

Figure 9.4. Types of stomatal humidity response and the consequent relation between transpiration (E) and VPD at the leaf-air interface (VPDo), between stomatal conductance gl and VPDo and for an apple tree. (Freely adapted from Jones, 1983; p 127).

The exact mechanism for the VPD effect on stomatal resistance is still being debated, but it is thought to be associated with cuticular (or *peristomatal*) transpiration. By that, we mean that the plant sensors of VPD may be the epidermal cells of the leaf cuticle which can sense the dryness of the air in contact with those cells. Here, an increase of dryness of the atmosphere will be felt as a slight desiccation by the epidermal cells, which trigger a response to that dryness in their buddies, the guard cells. Much of this still rests in the realm of speculation, however. Still, many plant scientists believe that, although independent of stomatal resistance, cuticular transpiration triggers a response in the stomatal resistance.

Canopy control of VPD; a reprise

Since either type of positive feedback would result in a decrease in carbon dioxide intake — or at least a reduction from the optimum possible intake — it seems reasonable to suppose that the self-protective aspects of feedback or feedforward, while helping the plant to conserve its water, would have detrimental aspects, not the least of which is a retardation in plant growth.

Let us therefore return once again to the idea of *auto-humidification*. Auto-humidification is a word we have coined in order to suggest that the plant canopy tends to retain a higher humidity in the canopy (much as a crowded theatre might be perceived as humid) than in the atmosphere above the canopy. The canopy thus maintains a higher humidity as a buffer against a detrimental decrease in relative humidity in the atmosphere above the plant canopy, as might occur during a hot, sunny day. Auto-humidification, therefore, would seem to provide a more efficient optimization of carbon dioxide intake in the face of a dry environment. One is further prompted to speculate that any such feedback mechanism must be beneficial to the plant in its normal (evolutionary) habitat but might have a detrimental effect in habitats for which the particular plant has not evolved. We began to explore in the last scenario the idea that the ambient relative humidity above the canopy (e.g. at 50 m) may not depend on the nature of the underlying plant canopy and, conversely, the internal (interleaf) relative humidity may vary somewhat independently of the surface layer parameters at 50 m, depending on LAI and leaf size, as well as on the ambient wind speed and vapor pressure. As stated by McNaughton and Jarvis (1991; see appendix previous scenario), the coupling or lack of coupling between canopy and atmosphere also depends on stomatal resistance, particularly when the latter depends on vapor pressure deficit (our emphasis in this chapter).

Idso (1987) has questioned the validity of stomatal resistance measurements, specifically the relationship between stomatal resistance and vapor pressure (measured with an instrument called the *porometer*, one of which our department of meteorology proudly owns !). He and his co-workers demonstrate that stomatal resistance, as measured inside the canopy, varied little despite large changes in saturation deficit at 1 m above the canopy. Idso interpreted this to mean that the relationship derived from the leaves in the porometer was not valid at the canopy scale where leaves are naturally exposed to the atmosphere. Idso contends that the problem arises because one does not take into consideration, when measuring stomatal resistance, the net effect of all leaves in the canopy, not just the sunlit leaves near the top.

Others, such as McNaughton and Jarvis (1991) argue that the reason for the lack of sensitivity of the stomatal resistance and the surface energy fluxes to changing VPD above the canopy is because VPD varies much less within the canopy. The reduction in VPD variation within the canopy is illustrated in Figure 9.2. Which view is correct depends, in the words of McNaughton and Jarvis (1991), 'upon whether the gains in the individual feedback pathways that operate at canopy level are sufficient to stabilize transpiration, making it largely insensitive to changes in (stomatal conductance) and on how the experiments were conducted'.

This contention underscores how difficult it is to interpret micrometeorological measurements in plant canopies. If Jarvis and McNaughton are correct, what is the use of the great quantities of data showing a variation between stomatal resistance and some vapor pressure deficit measured in the atmosphere just above the canopy? One of the problems in relating measurements to model results is the inevitable averaging of various properties over a scale of the canopy, *i.e.* to approximate a big leaf. The problem becomes even more acute in extending the canopy-scale measurements to the regional scale because now the canopy, like the individual plant in the canopy, is embedded in an even larger scale with its interlocking feedback mechanisms between individual fields. At the regional scale, the individual plant is now two scales down, rather than one scale down as if you go from the canopy down to the plant. Fortunately, we will stop at the scale of the canopy, and leave inter-canopy interactions for other heads to fret over.

Simulations

Level 1

We don't expect all of the previously discussed sophisticated and partly unverified ideas to be validated by the simulations. However, the model does contain a multitude of feedback mechanisms that are understood through simulations. The best thing to do is to observe these mechanisms by changing one parameter at a time. The idea in the level 1 simulations is to vary only VPD (at the leaf surface) and observe how the transpiration changes. In particular, we would like to observe the variation of stomatal resistance with time (as in the previous scenario) and also observe the effect on transpiration (using latent heat flux, LE, as an approximation to transpiration) as VPD increases beyond some threshold when we allow r_s to depend on VPD. By allowing a large value of LAI = 7, we can assume that the evapotranspiration (the output moisture flux) is very nearly equal to the transpiration. To do this without changing other parameters is tricky, but we can certainly attempt to do this in a model. The simplest way to change VPD in this model is to vary the atmospheric humidity.

Simulation series 1

Vary the vapor pressure deficit

First, rerun the base case scenario for corn with LAI = 7. Record the VPD, stomatal resistance, and transpiration (approximately latent heat flux when LAI is large) at noon or some reference time of your own choosing near midday. Save the output from each run since we will be looking at other reference times below. Then, run a couple additional cases, each time with a differing temperature-dewpoint spread in the lowest three or four levels of the temperature and moisture sounding (e.g., try decreasing the low level dewpoints by 2 or 4 C below the default values) and record your results. This is similar to what you have already done in the previous simulation except that we now want to generate a range of VPD values at your reference time. (Don't forget that corn is a good example of a plant with a very weak VPD-stomatal resistance coupling. The linkage between these two parameters is controlled by the parameter 'beta' in the veglut.dat file. For corn, the value is set to zero to indicate a very weak linkage.)

Next, we want to observe how the transpiration changes over the course of the day for the default and dry cases. Using the output from the above experiments, record the VPD, stomatal resistance and transpiration (using LE as a surrogate) at 2 and 4 hours before and after your midday case for the default VPD and also one of the drier VPD cases you ran above. Then plot the transpiration flux and stomatal resistance as a function of the time for your 5 selected hours, plotting the default and dry cases separately.

Questions

1. Note that the 'transpiration plateau' near midday is caused by the sudden increase in stomatal resistance. How does this plateau and stomatal resistance bulge change with increasing VPD?
2. Why do you suppose that stomatal resistance increases with increasing VPD even though no direct connection is made between these two parameters in the model? (To understand this fully it may be necessary to go on to chapter 10 first.)
3. What do you suppose would happen to assimilation (carbon dioxide intake) during the middle of the day in cases where you see the transpiration plateau?

More on the transpiration plateau in Chapter 11.

Simulation series 2

Make the beta parameter non-zero and vary the VPD

The beta parameter is the link between vapor pressure deficit and stomatal resistance. The larger the value of beta, the more strong the relationship between VPD and stomatal resistance. Values of beta of about 0.05 mb bar^{-1} are beginning to be quite hefty; beta probably does not exceed 0.10 in nature. It may be small (less than 0.02) in some plants such as corn. A zero value should be in place for the control run. (Again, consider why stomatal resistance increases with increasing VPD in simulation 1 where beta = 0; chapter 10 may shed some light on this if you are stuck.)

Change the value of beta in the model. In the 'simsphere/data/veglut.dat' file, change the fifth column (fourth number) for 'Corn' from zero to a medium-sized value of beta (0.05-0.10) and repeat the simulations with the same atmospheric temperature-dewpoint spreads (the default case and one dry case) that you just ran above in simulation 1 where beta was equal to zero. (Note that setting beta to 0.05-0.10 is artificially high for corn and would be more representative of a plant such as soybeans; we only make it non-zero here to observe its effect on the model). Again, record the value of VPD, stomatal resistance, and transpiration flux at your five selected times and plot them on a new graph as you did in simulation 1. You should now have two tables and two graphs from your simulations.

Questions

1. How does the midday stomatal resistance change with increasing initial VPD for the non-zero beta runs?
2. Do you see evidence of feedback here, where an increase in VPD and stomatal resistance acts to reduce the maximum transpiration flux as compared to simulation 1?
3. What is the effect of increasing the strength of the VPD-stomatal resistance link (by increasing beta) on the transpiration and the sensible heat flux (refer to the sensible heat flux values in the output from your model runs)?

Simulation series 3

Try and find examples of feedforward

Theoretically, as the VPD and corresponding stomatal resistance keep increasing, the transpiration should reach a level where it starts to decrease (feedforward) as VPD is increased. Although this effect may be elusive, try increasing beta and the VPD in the atmosphere and plot the noontime VPD versus noontime stomatal transpiration for the series of runs to see if a reduction in transpiration is found in the model. If this seems too tedious, put it aside and return to it again after having done the simulations for chapter 10.

Level 2

We will now combine the ideas discussed in the last two chapters. Try to find out what the effect of auto humidification, discovered in Chapter 8, would be on the relationships between VPD, stomatal resistance, and transpiration discussed here. Make as many simulations as necessary, using the parameters that worked for auto humidification previously and applying them here. Vary the atmospheric humidity as you did in previous simulations and note how the in-canopy VPD varies in response.

Questions

1. Does the relationship between VPD and stomatal resistance break down if the canopy is buffered from the atmosphere above?
2. If so, how does this affect the transpiration?
3. Do you see any signs of feedback or feedforward in a canopy that undergoes auto humidification?

Keep in mind that now you are altering numerous parameters, some of which have feedbacks on each other. Therefore, the results might not be as clear cut as they seem. Try and keep an open mind and assess all the changes you make equally. Good luck!

Definitions

VPD = Vapor pressure deficit
1 kPa is equal to 10 mb

Conductance = The inverse of resistance, 1/r
Peristomatal = Cuticular
Porometer = Instrument which measures relationship between stomatal resistance and vapor pressure

Terms

Terms to look up and remember
Peristomatal
Porometer

References

Korner, C., 1985, Humidity responses in forest trees: precautions in thermal scanning surveys, **Arch. Met. Geoph. Biocl., Ser. B**, 36, 83 - 98.

McNaughton, K. and P. G. Jarvis, 1991, Effects of spatial scale on stomatal control of transpiration, **Ag. and Forest Meteor**., 54, 279-302.

Idso, S.B., 1987, An apparent discrepancy between porometry and thermometry relative to the dependence of plant stomatal conductance on air vapor pressure deficit, **Agr. and Forest Meteor**, 40, 105 - 106.

Simsphere Workbook: Chapter 10

Water Potential in Simsphere

Introduction

We have seen that in a dense vegetation canopy almost all the latent heat flux comes from the plants. Just about all of that comes through the leaves and almost all of that through the stomates. Look around you and you will see that the leaves on most plants reside in the upper third of the height of the plant. Of course, most of the sunlight is trapped in the upper third of the canopy, so, if you were a plant, why bother putting leaves any further down? You might well ask: why bother to grow so high in the first place, if most of the leaves are confined to a relatively shallow layer near the top? The answer has to do with keeping up with the Jones', so to speak. If everyone is growing just a little higher than you are in order to steal your sunlight, then you too must grow up higher to stay in competition for the precious life-giving radiation. The bucolic picture of nature as benevolent and peaceful is misleading. Wall Street has nothing on nature; the competition amongst plants in the wild is fierce and deadly, if not a bit slow and silent. You might say that it is a jungle out there!

Yet plants do arrive at compromises, at least within a species. Trees of a single species growing in a particular environment may eventually settle on a height, which is more or less the same for all mature trees in that lot. New trees will have to struggle for themselves (or die) but the trees eventually stop growing beyond a certain height. It takes work to make a big tree, even if the tall timber has its advantages. The point is, that water and therefore sap must be pumped (or rather pulled) from the root zone to the top of the canopy before it can be evaporated or used to make more tree or leaves. Somehow, the leaves need to know how much water is available in the root zone and pump it up to the leaves about the same rate as is being transpired[1]. The formalism for expressing how water rises from roots to leaves and how the leaf water status may affect the stomatal resistance constitutes the focus of this scenario.

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1. This is only true if the plant itself does not use its stored water for transpiration. Plants have the capacity for storing water in their leaves, stems and roots and they make this water available as needed. Our model can accommodate a water flux to and from storage. In general, this flux is relatively small, however.

Water Potential

Water potential is one of the most difficult concepts to understand. One reason for the confusion with regard to water potential is that it has various definitions depending on what water potential one is referring to. In addition, the use of the concept of water potential in many different disciplines has generated a babble of terminology and some erroneous definitions. Water potential describes the water status[2] of the soil and cells within leaves, although the processes involved are vastly different. Nevertheless, water potential is a useful concept because it describes the flow of liquid water through soils and through cell walls in the plant, - from higher to lower water potential across a resistance. This is the essence of the so-called 'K-theory', which forms the basis for our resistance laws referred to in these scenarios. Thus, it is customary to express the flux of liquid water F_w as a resistance law, similar to Eqn. 3.1, in which a resistance (Z) exists across a potential drop.

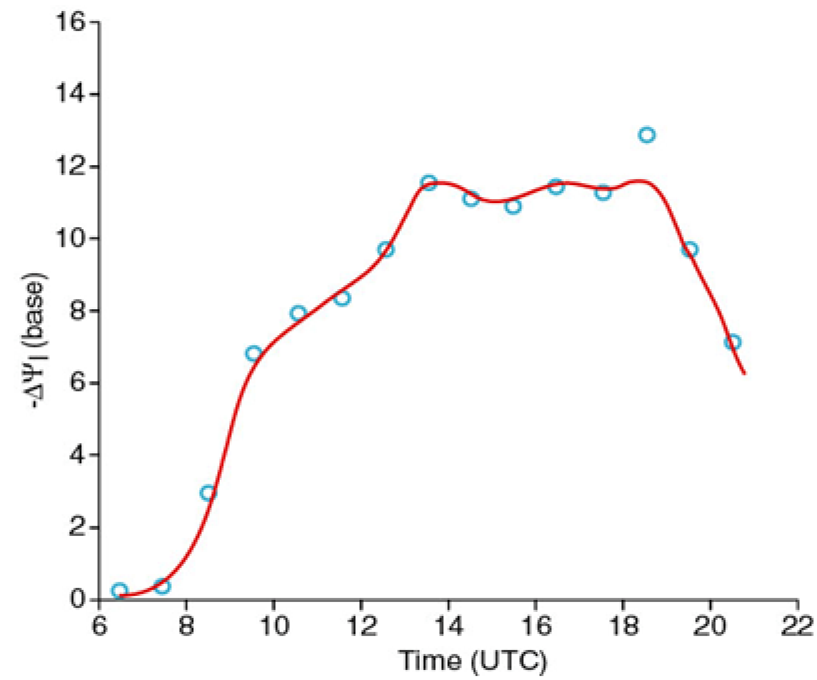

Figure 10.1. Measured differences in leaf water potential from the pre-dawn (base) potential value plotted versus Greenwich (approximately local) time. Dots represent individual HAPEX measurements (made near Lubbon, France) composited for the month of June 1986. Circled crosses represent hourly averages. The actual leaf potential was the plotted value plus the base (soil) water potential. The average value of the base potential was about -1.5 bar.

The units of potential are **bar**[3] (or MegaPascals (MPa), where 1 MPa equals 10 bar). We will make the units of flux in Watts per square metre in order to equate the liquid water flux with the energy fluxes; (Henceforth, in talking about water (as opposed to vapor) flux, we will use the symbol Z, which has the units of bar $(W\ m^{-2})^{-1}$, for resistance and the symbol Ψ for potential.) The area unit is understood to be that of the big leaf. The area in question is the leaf area. The symbol Z is used instead of the resistance r because the units are no longer expressed as $s\ m^{-1}$. We still must convert that to a flux per unit surface area. This type of resistance law is expressed by Eqn. 10.2.

Water potential is generally a negative number, with zero water potential being the value for pure water in equilibrium with its surroundings. Having the units of pressure, water potential also corresponds to an energy per unit volume. As such, we may think of this energy as that required to extract a molecule of water from somewhere, say from a soil particle or from the crevices between them.

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2. The extent to which the water can flow or be extracted compared with a base state i.e. that of pure water.

3. We are still not sure how the plural works with this unit. Is it bar or bars?

Soil Water Potential

As there are many interfaces and pores in soil, the soil water potential may be appreciably less than zero, except for soil at field capacity, when Ψ_{soil} (which we will call Ψ_g) is typically about -0.1 bar. Field capacity is typically about 0.35 by volume for clay and loamy soils and about 0.10 by volume for sand. Check out Hillel (1982; pp 64 - 73) for a further discussion of soil water potential. Since soil water may contain solutes, the osmotic pressure component π is typically between -0.1 and -2 bar. We will ignore this osmotic component of the soil.

Water potential is widely used in soil science and hydrology. It is often referred to as the 'suction' (although we would like to see someone try to create a suction of -15 bar! If you think that you can do it, we'll furnish the straw). Sometimes water potential is called **the pressure** (or **hydraulic**) **head** (h), which is often expressed in metres rather than in pressure units (the conversion from one to the other is water density times gravitational constant times h, is equal to pressure) as shown in Eqn 10.1. Note that h comprises two components: the soil water potential (often referred to as the matric potential, which has no equivalent in the plant) and a gravitational component H. You may see this water potential cited as the **Total Hydraulic Potential**.

As the amount of soil water decreases, the air-water surface retreats into the crevices between the particles and the soil water potential, sometimes referred to as the **soil water tension**, becomes more negative. Can you physically say why this is? Hint: The meteorologists among you may like to consider the so-called Kelvin equation you may have already seen in the cloud physics part of physical meteorology and make a comparison with the equation given on page 70 in Hillel.

Soil water tension is measured by an instrument called a **tensiometer** (Hillel, 1982; pp 81 - 83). Various parametric equations exist relating soil water potential to soil water content for different types of soils (Cosby *et al.*, 1984). You will often see the term **permanent wilting point** used to mean the equivalent soil water content when the volumetric water content is so low that many agricultural crops will not recover. Wilting point is more symbolic than a measurable quantity; it is traditionally assigned the value of -15 bar, Typical values of soil water content at wilting are 12 percent by volume for clay or loamy soils and 3 percent for sand. Of course, the misnomer is that soils do not wilt; plants wilt. To shed some light on this misuse of the term, we will later relate the wilting point to the leaf (rather than the soil) water potential.

Root, Stem and Leaf Water Potentials

As plants modify the **osmotic potential** across their roots water is drawn into the plant and then moves up the stem against gravity. Since the roots remove water from their immediate surroundings, the water content adjacent to the roots is apt to be lower than that at the same level in the root zone but away from the roots. This results in a decrease in soil water potential from the soil around the roots to the surface of the roots.(This is an important point which we will come back to later on.) Water will also flow upward or downward, but in the former case, it can not do so unless the drop in water potential with height is greater than the increase in gravimetric potential with height. Since the change with height in gravimetric potential is generally small (about 0.1 bar per metre), we will henceforth neglect its effect in the resistance equations. The resistance to water flow through the ground is a function of soil water content, and various parameterizations exist relating soil water content to hydraulic conductivity (Cosby *et al.*, 1984).

To enter the roots, the water must flow down a potential gradient, which the plant conveniently arranges by changing the **osmotic pressure** in the roots. Plant roots tend to be very fibrous and fuzzy, thereby having a large surface area (sometimes several hundred square metres!) over which water and nutrients within the water can enter the plant. The resistance to the flow of water down this potential is a function of root mass, among other things, the larger the root mass the smaller the resistance. In the model, this resistance between surrounding soil and the interior of the root surface depends on the soil conductivity, which is, as we have already said, a strong function of soil water content and on the root mass; modelled crudely as a function of plant height. (We assume that the bigger the plant the greater the root mass.)

Once inside the roots, the water flows up through the stem to the leaves through tiny, tube-like channels called **xylem**. This water motion also takes place across another drop in water potential. This potential drop is related to capillary action and to gradients of osmotic pressure within the plant. Fundamentally, it occurs because of the high surface tension of water. Briefly stated, when one molecule of water moves, for whatever reason, others are dragged along with it. For water to be transported to the leaves, the leaf water potential (specifically the **mesophyll water potential** Ψ_m see figure 7.1 in scenario 6) must decrease below that of the soil water potential Ψ_g. This relationship between the flux of water through the plant to the leaves (F_w) is expressed by Eqn. 10.3.

The meaning of leaf water potential becomes clearer when one watches a measurement. Leaf water potential, more precisely, the mesophyll potential, was, until fairly recently, customarily measured with a pressure chamber device, sometimes called a 'pressure bomb'. A piece of a leaf is inserted in a chamber, which is allowed to undergo an increase in pressure. When the pressure reaches the value of the leaf water potential, water is driven from the intercellular spaces of the leaf and beads of it begin to form on the outside of the leaf. A pressure gage, attached to the housing, is used to record the value of pressure at beading. The device is a bit crude and bulky, and the uncertainty in making a field measurement is about 1 bar. It is also a tedious measurement to make. Happily for the field worker, the device has recently been replaced by a more sophisticated apparatus that does not require the death of a leaf in order to make the measurement. (The significance of this uncertainty is elaborated upon in a later scenario)

There is also a further drop in water potential between the mesophyll in the leaf and the epidermis, specifically the guard cells. This potential drop is not associated directly with the rate of water flux in the plant, but is correlated with the rate of vapor flux through the leaf cuticle and with the vapor pressure deficit between the stomatal cavities and the leaf boundary layer. The **epidermal leaf water potential** Ψ_e may constitute a link between the guard cells and the mesophyll and hence the water potentials throughout the plant. Epidermal leaf water potential is very difficult to measure directly without sophisticated equipment. Later, we will touch on its possible significance.

The dynamics of soil, root, plant and leaf are tied together by allowing the transpiration, (Equation 7.3) and the flow of water from soil to leaf (Equation 10.3) to be equated, a situation one might call 'steady state'. That is to say that the flow of water, either as liquid in the plant or as transpiration, is constant along the entire route from soil to leaf to air. In so doing, we have equation 12.3 in which soil water content, acting through the soil conductivity and the root hydraulic potential, influencing the stomatal resistance of the leaves. Thus, the theory incorporates both root and shoot sensitivities. Of course, the plant is not actually in exact steady state. During the day the transpiration may exceed the flow of water from the roots, resulting in a loss of water from the leaves and stem. At night, however, the transpiration is negligible, whereas water still flows upward from the roots, resulting in a storage of water in the leaves and stems. In fact, stems are observed to shrink slightly during the days in response to water losses. Simsphere does have a storage (or capacitance) module, but the user is strongly advised not to employ this component without a deep understanding of the capacitance physics described by Lynn, B.H. and T.N. Carlson, 1990, Agricultural and Forest Meteorology, Vol. 52.

Simulations

Level 1

Simulation series 1

- Run the base case scenario for vegetation but examine the leaf and soil water potentials during the day. Do this again for a lower value of root zone moisture availability (*e.g.* 0.45).

Simulation series 2

- Rerun the base case but increase the xylem resistance. In the simsphere/data/veglut.dat file, change the tenth number (ZP in the model code) for 'Corn' from 0.05 (the current default value) to 0.10 bar (W m-2)-1. Before performing this simulation, ask yourself what will happen to the leaf water potential and the stomatal resistance now that you have increased the denominator in Eqn. 10.3, assuming that this equation and Eqn. 7.3 must both be satisfied. See if your intuition holds up. What changes are occurring in transpiration and in leaf water potential during the mid morning and near mid day ? (If nothing strikes your attention, try this last simulation with a root zone moisture availability of 0.45, recalling what happened when you reduced moisture availability to 0.45 in a previous vegetation scenario.) Do you see a pattern that resembles day 6 in Figure 10.2?

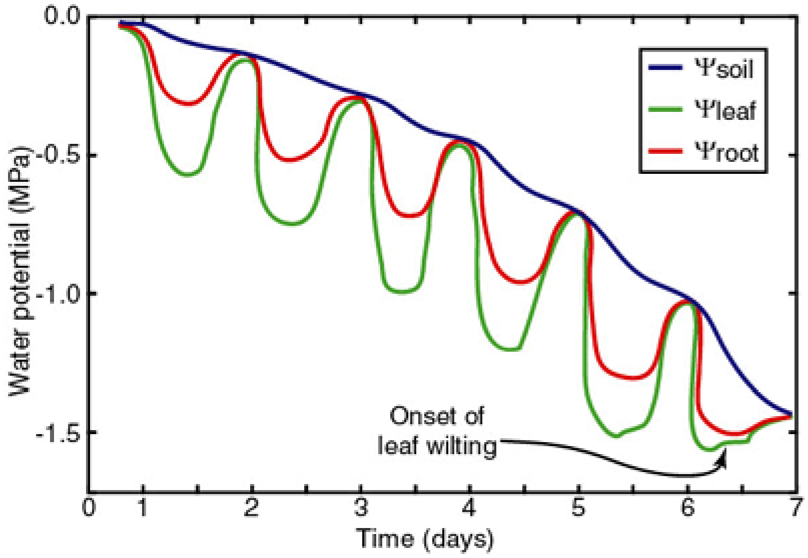

Figure 10.2. Schematic representation of daily changes in the water potential in the soil, root and leaf of a plant in an initially wet soil that dries out over a one week period. Shown are curves for the soil water potential, root xylem water potential and leaf (mesophyll) water potential, as adapted by Noble (1983; his figure 9.13) freely adapted from an article by Slatyer (1967, p 276).

Equations

[Eqn 10.1]
$h = \Psi_g + H$

[Eqn 10.2]
$$F_w = \frac{\Delta\Psi}{Z}$$ (General Form)

[Eqn 10.3]
$$F_w = \frac{\Psi_g - \Psi_1}{Z_g + Z_p} = T_r$$ (Steady State)

Terms

Terms to look up and remember
bar (scientific definition please)
pressure head
hydraulic head
soil water tension
tensiometer
permanent wilting point
xylem
mesophyll water potential
soil water potential
epidermal leaf water potential
osmotic potential
osmotic pressure

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Simsphere Workbook: Chapter 11

More About the Root Zone and Water Potentials

Introduction

Next to incident flux of *photosynthetically active (solar) radiation (PAR)*, soil water content is the most important variable affecting stomatal resistance. Measurements show that as the soil dries out the stomatal resistance tends to increase, slowly at first and then much more rapidly. However, this is about as far as agreement goes in the great 'root-shoot' controversy. Until a few years ago, most texts on stomatal resistance, such as the one by Jones (1983), discussed stomatal resistance in terms of the leaf water status, as represented by the leaf (mesophyll) water potential. The force of this outpouring of research carried over into meteorology in those few atmospheric models whose surface parameterization make explicit reference to plants, such as the BATS and SiB models. (These models, which attempt to describe actual physical processes occurring within plants, are *mechanistic*, as opposed to those that are satisfied only with bulk formulae in which the various coefficients may be determined by curve fitting data. Naturally, there are degrees of mechanistic modelling; one person's mechanism is another's empiricism.)

Anyway, something happened in plant research around 1985. Some plant scientists found that they could experimentally induce stomatal changes by altering the soil water content only, keeping the leaf water potential constant, or by treating the roots with the plant hormone ABA. It is difficult to assess the upheaval in thinking that occurred over 15 years ago, because the adherents of the root-control school have been so aggressive and vocal that a large minority, if not a majority, of the scientific community has been silenced. To speak contrary to the new viewpoint, *i.e.*, to advocate shoot control, is almost equivalent to supporting a politically incorrect point of view, such as that the place of women is only in the kitchen! Like the advocates of this latter and, we might add, regressive social idea, there are still a large number of individuals who secretly harbor the reactionary viewpoint of shoot control, hoping that it will someday make its return. (As of 2018, this contention may be resolved.) Alternately stated, there are still those of us who continue to believe that leaves—specifically leaf water status—exert a profound influence on stomatal resistance. In addition, some scientists have looked at flux measurements made in the field, and find in these measurements strong evidence of a shoot mechanism at work. (We refer to the associated phenomenon, to be discussed below, as the *transpiration plateau*.)

In this scenario, we will investigate the behavior of stomatal resistance with changing soil water content, but we will use a parameterization that involves linking stomatal resistance directly to leaf water potential. It's not that we want to discredit soil water content, but to show how it is possible to calculate stomatal resistance as a function of both soil water content and leaf water status (leaf water potential). You should keep an open mind as you ponder this scenario.

Root zone water content and stomatal resistance

Messages that inhibit stomatal opening are sent from the roots to the leaves in response to the perception of water stress in the root zone. These regulators, which have the general name of *cytokinins*, such as ABA (*abscisic acid*), are synthesized in the roots and carried through the xylem to the stomates. The response time for this communication to take place is a few minutes to perhaps hours for a very large plant.

Virtually all the existing models that relate stomatal resistance directly to soil water content employ some function of either soil water content or soil water potential. The functional form of these equations allows stomatal resistance to increase as the soil water content diminishes. At first, there is a very small change in stomatal resistance with decreasing soil water content below field capacity. Then, at low soil water content, the rate of increase of stomatal resistance with decreasing soil water content becomes very large. This behavior is shown in Figure 11.1, which is taken from Turner (1991; his figure 4). Note that there is almost no variation in stomatal conductance between about 40 and 100 % extractable water in the soil. Apparently, stomatal resistance is very weakly related to soil water content until the latter reaches some value well below saturation. This type of threshold behavior in plants occurs in other contexts, for example in relation to the solar flux (PAR) and to soil hydraulic resistance as a function of soil water content

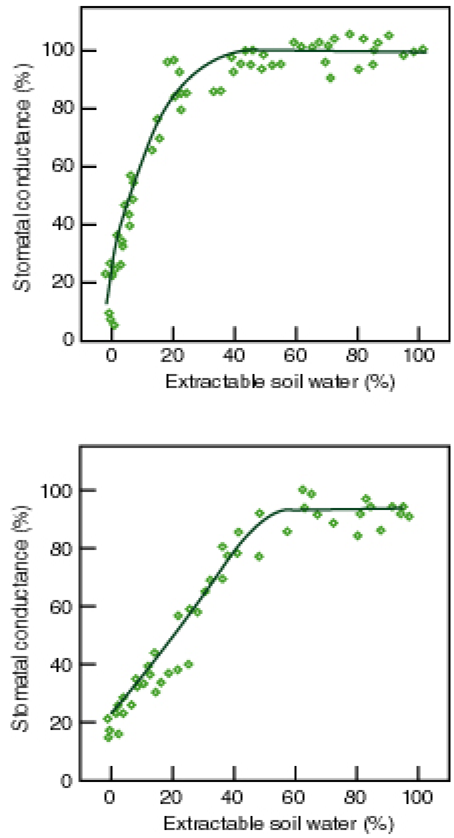

Figure 11.1. Relationship between stomatal conductance and extractable soil water content in two types of plants (Freely adapted from Turner, 1991)

Let us agree to refer to *plant water stress* as occurring when the stomatal resistance is increased (from some reference state) due to a reduction in water content in the soil or leaf. (Stress on the plant may arise from other factors, such as exposure to ozone, as discussed in Chapter 13.) An increase in stomatal resistance that causes an effective shutdown in transpiration defines the so-called soil wilting point. As we have begun to see, however, internal interaction within the system of different factors can mask the effect of a change in just one component. Conversely, the effect of soil water content on stomatal resistance can be indirect, the result of changes in other factors, such as vapor pressure deficit or solar intensity.

If we let you, you could exercise the formulations that link directly the root zone water content and stomatal resistance via the 'Deardorff' equation, which is an option in the model. For the present, however, let us move on to the leaf water potential response because we believe it to be more elegant than the Deardorff equation.

Leaf Water Potential and Stomatal Resistance

Let us return once more to Figure 10.2, which shows a schematic change in leaf water potential during a week of drying. Although the difference between leaf and soil water potentials appear almost constant for the first few days, the leaf water potential appears to crumple against an unseen barrier at about -15 bar. Now look at Figure 11.2, taken from Nizinski and Saugier (1989; their figure 2). They show that the measured leaf water potential of the forest (heavy solid dots) are approximately constant along a value of stomatal conductance of about 0.2 mm s-1 (stomatal resistance of 50 s cm-1) until the leaf water potential reaches a threshold of about -15 bar, below which the stomatal resistance increases rapidly with a further decrease in leaf water potential. This is what we examined in the previous chapter. Ultimately, at a leaf water potential of about -25 bar, the leaf water potential has exceeded the constant value of the cuticular resistance (small solid dots).

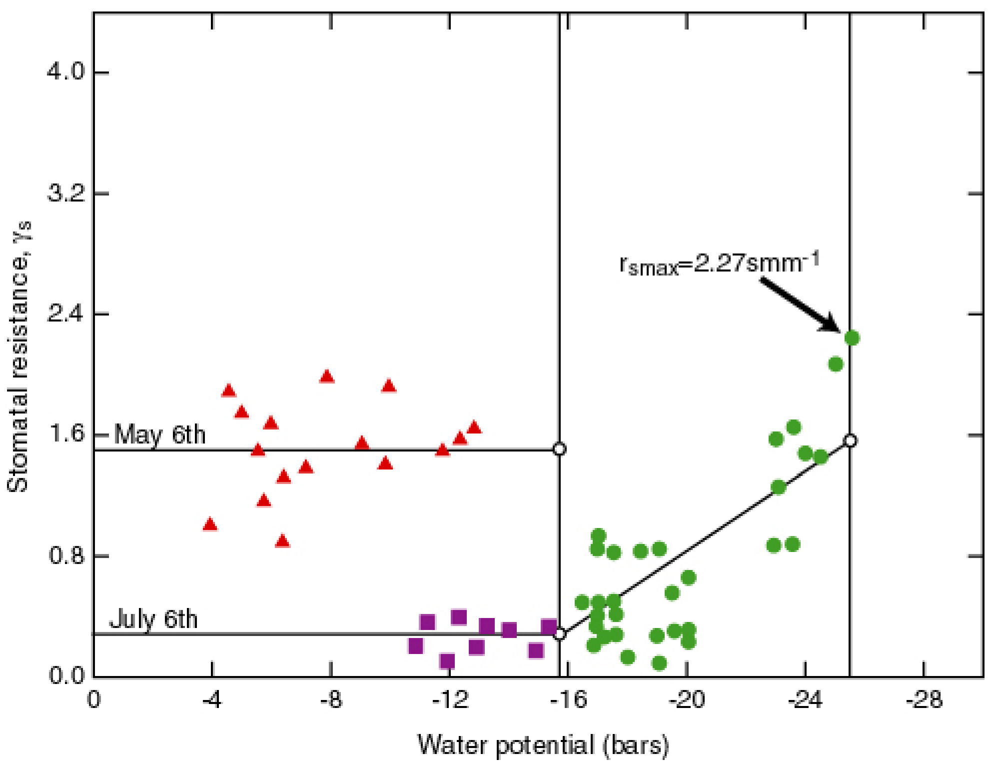

Figure 11.2. Variations in the mean leaf stomatal resistance for oak trees (r_s) with the mean mesophyll water potential in a canopy (Ψ_m). Field measurements on different days during the 1983 growing season; (Freely adapted from a figure from Nizinski and Saugier, 1989)

The transpiration plateau and the leaf water potential threshold

Measurements showing some sort of leaf water potential threshold frequent the literature[1]. Consider a typical example of this which is presented by Turner (1991; his figure 6). Many of Turner's earlier papers present dramatic evidence of a threshold leaf water potential. Yet Turner became a firm advocate of soil water control (a root partisan). The presence of a water potential threshold is markedly visible in Turner's data, which we show in Figure 11.3. Note that there is very little change in stomatal conductance above a leaf water potential of about minus 10 to minus 12 bar, below which there is a catastrophic downward plunge in stomatal conductance with decreasing leaf water potential. (Recall that stomatal conductance is simply the inverse of stomatal resistance.) Of course, this could be the result of an indirect relationship in which the roots signal the leaves to shut down. Turner presents both sides of the root-shoot issue in his paper but flatly declares that leaf water potential is not the controlling factor in leaf hydration. Some of this problem may be semantics. An underlying factor may be controlling both the leaf water potential and stomatal resistance. Turner suggests that the mechanism is related to abscisic acid, which comes from the roots. However, he leaves the door open for the possibility of transient leaf control by implying that there might be two sources of stomatal response. One signal is in the roots and it is carried to the leaves in response to slow changes in soil water content. Another is in the leaves, which are also known to produce some abscisic acid; this allows for fluctuations in the mesophyll water potential (Figure 11.3).

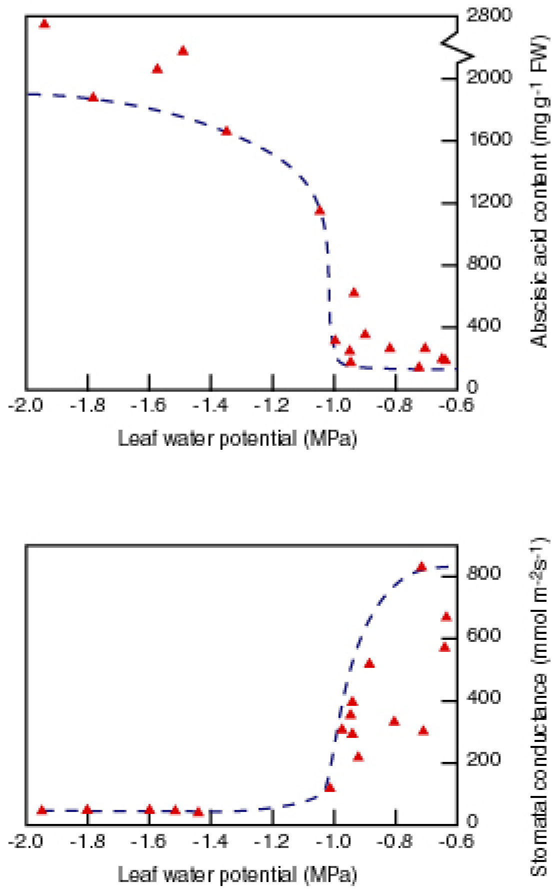

Figure 11.3. Relationship between stomatal conductance and leaf water potential (lower) and leaf abscisic acid content and leaf water potential (upper), in lupin in the field; (Freely adapted from a figure from Turner, 1991).

Perhaps a more cogent statement in favor of leaf control was made by two plant physiologists (Davies and Mansfield, 1987), whom we quote in the appendix. Note that they say that a water deficit affects the balance of regulators in the leaves. Even if leaf water potential is not in itself the mechanism for changing stomatal resistance, both it and the stomatal resistance are regulated in tandem by another mechanism, *e.g*. the leaf abscisic acid. In that case, our viewpoint is less than completely mechanistic, but it is still circumstantially correct to make stomatal resistance dependent on leaf water potential, since modelling the concentration of abscisic acid lies well outside the scope of our expertise or interests. You may draw your own conclusions.

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1. We have some similar measurements made at Rock Springs (Penn State) in a soybean crop. The threshold value of leaf water potential for this crop was also about -15 bar. More recent measurements by the distinguished plant experimentalist at USDA, James Bunce, brings new evidence to bear showing that the leaf indeed has a voice in controlling stomatal resistance.

Leaf turgor and wilting

According to Jones (1983), the onset of wilting, and so the threshold water potential, is associated with a vanishing of the difference in water potential between the inside and outside of the mesophyll cells. This difference, which is generally positive, is called the *turgor pressure*. (This term was introduced in the previous chapter.) You can think of turgor pressure of a balloon as that which is keeping it inflated; so also for the cell. For a given cell solute, turgor pressure decreases as the total water potential decreases, resulting in the cell becoming less and less turgid with a decrease in its water content. A shrinking of the mesophyll cells accompanies the loss of turgor, so that wilting (the reduction of turgor pressure to zero) manifests itself as a withering of the plant. Indeed, cell shrinkage is reflected by a slight shrinking of the stem diameter as the plant dries out!

A complete loss of turgor corresponds to a situation of distress in the plant. Although some plant scientists discount the vanishing of turgor pressure as a mechanism for inducing the increase of stomatal resistance, it is likely that the leaf water potential threshold occurs somewhere near that point. For many plants, zero turgor is reached when the water potential in the cells is close to -15 bar. Some scientists have suggested that, since the stem also shrinks with decreasing water content, a measure of stem diameter changes might constitute a measure of soil water content. Although many plants exhibit a stress threshold at values considerably less than (more negative than) or greater than (less negative) -15 bar, the latter figure has been adopted as the official wilting point. Choice of -15 bar arises because many soils exhibit a very rapid decrease in soil water potential with decreasing soil water content near -15 bar. That is the reason why the choice of soil type (and threshold water potential for the plant) is so important in this model. We can see that this value of -15 bar constitutes almost a magic number!

Loss of turgor and incipient wilting need not be permanent or even associated with a low soil water content. We see examples in which the fluxes exhibit a transient response to drying soil, suggesting that control of stomatal resistance also exists in the leaf or is manifested via the leaf. Consider Figure 11.4, which shows two evapotranspiration curves made over corn, one curve superposed on the other. The curves pertain to two adjacent fields, one irrigated and wet the other dry; (we are not sure how wet or how dry was the soil in each field, but you can certainly tell which is which). Note that the two curves are virtually identical in the morning and again later in the late afternoon; in between the one set of fluxes shows a distinct flattening. We refer to this flattening as the *transpiration plateau*.

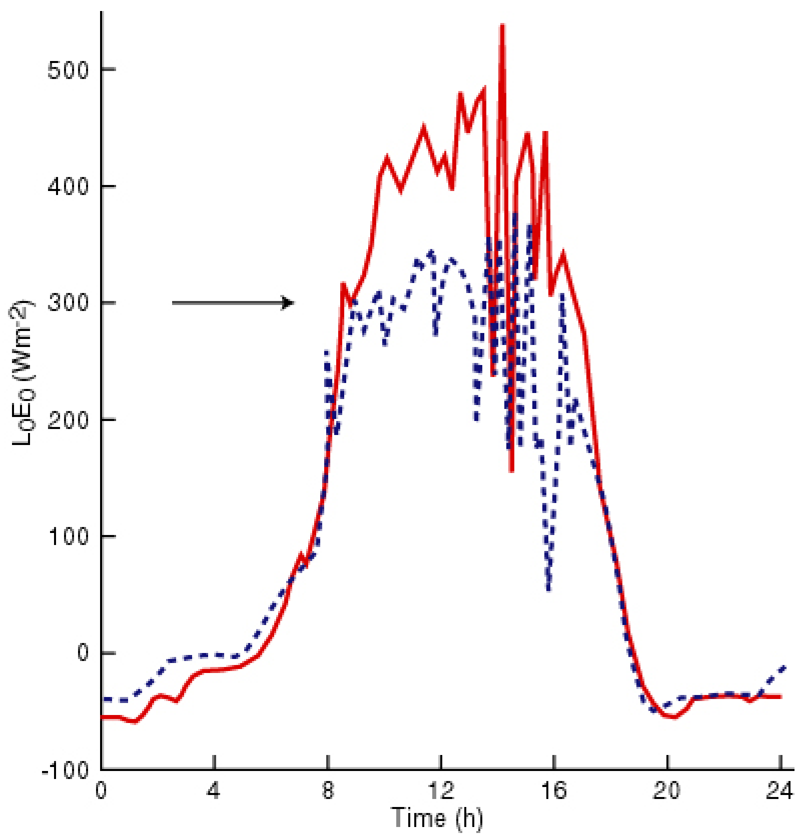

Figure 11.4. Measured evapotranspiration (Wm^{-2}) during the HAPEX experiment, Castelnau, France, June, 1986. Dashed line is for a non-irrigated field of corn while the solid line is for an adjacent irrigated field of corn. Horizontal arrow indicates an evapotranspiration plateau near 300 W m^{-2} for the non-irrigated field, suggesting that the leaf water potential is at threshold; (Freely adapted from a figure from Lynn and Carlson, 1990).

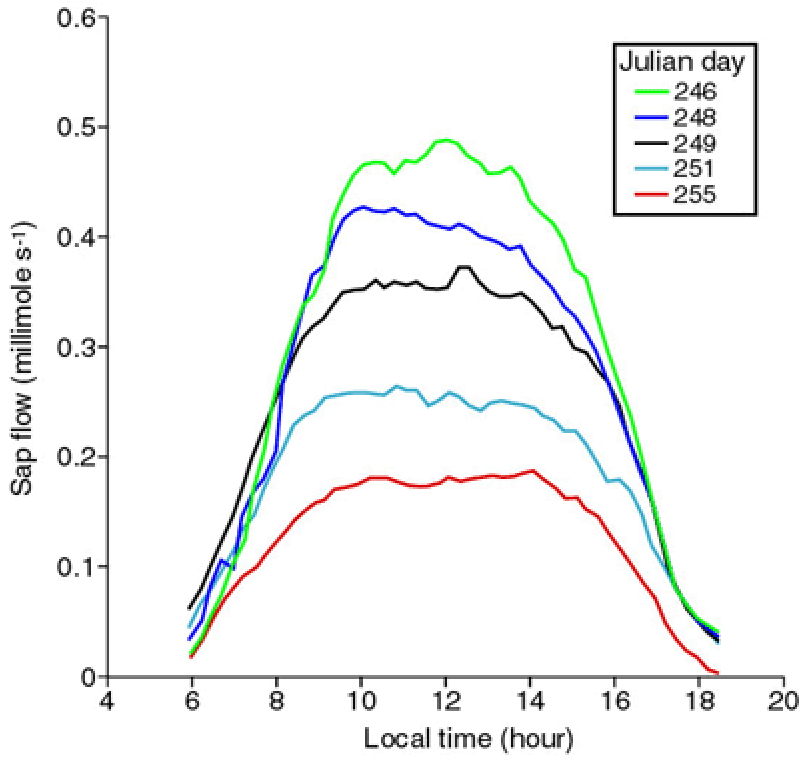

Figure 11.5. Sap flow vs. local time.

Figure 11.5 is another example of the emergence of a transpiration plateau over a period of several days during which the plant canopy dries out. Note that the transpiration plateau becomes larger each day, although almost no differences are observed before the plateau is reached and after the transpiration curve is no longer flat.

Modelling the transpiration plateau

How does this plateau come about and why is it transient? Recalling the previous scenario, we note that the leaf water potential decreases to a minimum during midday (in accord with the typical rise and fall of transpiration during the day). The leaf water potential may reach a threshold (*e.g.* -15 bar), below which the plant sounds the alarm to close its stomates. We refer to this threshold as the *critical leaf water potential*. Below the threshold, the stomatal resistance increases rapidly with a further decrease in leaf water potential. That increase in stomatal resistance will tend to retard transpiration and limit further decrease in leaf water potential.

We formulate stomatal resistance as a function of leaf water potential. This is done as follows: The stomatal resistance function (Eqn. 7.2) is expressed in terms of a component ($f(\Psi_e)$), where Ψ_e is an epidermal leaf water potential, which will be described shortly. When the leaf water potential lies above the imposed threshold leaf water potential, stomatal resistance depends weakly on leaf water potential (in the case of our corn plants, that sensitivity to leaf water potential is taken to be zero). Below the threshold, leaf water potential stomatal resistance increases very rapidly with decreasing leaf water content. We represent each regime, below and above the threshold, by straight lines. The pair of straight lines defining the sub-threshold and super-threshold regimes intersects at the point forming a sort of ramp function. The point of intersection defines the threshold leaf water potential. In Figure 11.2,the two straight lines, one horizontal and one sloping upward to the right, intersect to define the threshold water potential at -16 bar.

Because of the data scatter, it is really difficult to determine if stomatal resistance is at all sensitive to leaf water potential in the sub-threshold regime. Since the data is so limited, we arbitrarily chose the slope of the line to be zero in the sub-threshold regime, as in Figure 11.2 (no sensitivity of stomatal resistance to leaf water potential; you can change these slopes in the model in specifying a new vegetation type, but if you do, BEWARE). Beyond the threshold, the slope of the line is very steep. The most important number in this equation is, however, the value of the threshold leaf water potential. Note that -1.0 MPa is equal to -10 bar.

Leaf water potential and vapor pressure deficit

Most plant scientists will concede that *if* a leaf potential tripwire exists, (the alarm mechanism that signals the arrival of the threshold and zero turgor), it lies not in the mesophyll, but in the epidermis, specifically in the guard cells. Accordingly, we relate the stomatal resistance to the epidermal leaf water potential Ψ_e . Now, the mesophyll water potential Ψ_m governs the transpiration in Eqn. 10.3, but the epidermal water potential is the appropriate potential used in the stomatal resistance equation, Eqn. 7.2. Epidermal leaf water potential is similar in magnitude to that in the mesophyll, although it will always be equal to or lower than the mesophyll potential. We take the difference between the mesophyll and epidermal leaf water potential to be proportional to the vapor pressure deficit. The constant of proportionality between mesophyll water potential and epidermal water potential is called beta, the parameter you varied in the last scenario. (Don't worry too much about this refinement; it's a brand new approach to relating stomatal resistance to vapor pressure deficit; some plant scientists actually like the idea!) The two potentials can behave somewhat independently; the epidermal leaf water potential could decrease to the threshold, while the mesophyll leaf water potential increases, in accordance with a decrease in transpiration. (See if this helps you answer the question from Chapter 10 regarding Ψ_e and Ψ_m).

The more conventional way to parameterize the vapor pressure deficit effect is simply to increase the stomatal resistance by some function of the vapor pressure deficit. This is the basis for the so-called 'Jarvis' formulation, which is the form of the stomatal resistance equation used in our model, except, of course, for the VPD effect. Although the two methods for augmenting the stomatal resistance as a function of vapor pressure deficit represent fundamentally different processes, they tend to produce a similar effect when viewed overall. The problem with the vapor pressure deficit effect is that it tends to mask the water potential threshold effect by accomplishing the same thing: an increase in stomatal resistance during the middle part of the day, although somewhat skewed toward later afternoon.

Leaf water potential and transpiration

As formulated, the stomatal resistance should vary little with leaf water potential (or with soil water content) until the epidermal leaf water potential reaches the threshold value. Imposing a threshold for leaf water potential in the model leads to a midday increase in stomatal resistance when the transpiration becomes sufficiently large to require the leaf water potential to fall to the threshold. (A large vapor pressure deficit may also cause the leaf water potential to reach the threshold much sooner, according to our formulation.) When this happens, the transpiration starts to be arrested but, at the same time, the rise in sensible heat flux from the leaf and the concomitant rise in leaf temperature work to augment the transpiration flux (by increasing the vapor pressure deficit). The result of this tug of war is that the transpiration neither falls nor rises, but forms a plateau. The lower the soil water content, the more quickly in the morning the leaf water potential reaches the plateau and the longer the plateau lasts. Look at your results from Chapter 9. This process exemplifies the feedback mechanism previously discussed.

Note that the leaf water potential may not reach the threshold value if the soil is sufficiently wet, because it starts out from a relatively small negative value. In that case no soil water effect will be observed in the simulations. The same thing may happen if the sky is cloudy and the atmospheric demand for transpiration is weak, because then the leaf water potential will not fall very far in the absence of strong transpiration demand. Alternately, the leaf water potential may reach the threshold after starting out at a relatively small negative value because the demand for transpiration is large, the vapor pressure deficit large or for other reasons, *e.g.*, because the denominator of Eqn. 10.2 is large for some reason (*e.g.* Z_p is large).

You may have found in your previous simulations that stomatal resistance and transpiration do not vary greatly as a function of soil moisture content over a range of soil moisture values, although xylem resistance and other factors (such as ambient relative humidity) influence the behavior of stomatal resistance. When the water potential threshold is reached, however, large fluctuations in stomatal resistance occur. The earlier the onset of the plateau the lower the transpiration during the plateau and the higher the midday stomatal resistance. As suggested by Figure 10.2, the stomatal resistance and the transpiration may not change very much from one day to the next during soil drying until the leaf water potential reaches its threshold potential. After that, however, and until a general collapse of the plant occurs (permanent wilting), the changes are primarily during the period of the plateau.

A final word on the root-shoot controversy

As in many such controversies, the solution lies somewhere between the extremes. In all probability, both root and shoot influence the stomatal resistance. According to a French scientist named Tardieu, the real mechanism goes something like this: As the soil dries out, cytokinins (ABA) are sent to the leaves. When the leaf begins to perceive a water loss, stomatal resistance is increased and more ABA is produced, including in the leaves themselves. The effect of this hormone is to accentuate the response of the leaf to stress. The root secretions augment the effect of the threshold, although without affecting the stomatal resistance very much before reaching the threshold potential. As the root continues to dry out, the change in stomatal resistance with decreasing leaf water potential is enhanced, resulting in a more rapid increase in stomatal resistance with decreasing leaf water potential. Tardieu's solution incorporates elements of both arguments but preserves the threshold effect.

Values of threshold leaf water potential

1. Values of threshold leaf water potentials vary with species, age and previous stress episodes experienced by the plant. Many plants exhibit a threshold close to -15 bar (that magic number again!), although values vary between -9 and -22 bar depending on the species of plant. We set the threshold at -15 bar in our 'corn' simulations.

2. That the sensitivity is weak in corn when the leaf water potential has not yet reached the threshold is evidenced by the fact that field measurements tend to show no significant differences in fluxes from one day to the next before the threshold is reached but a considerable difference after the leaf water potential threshold is reached. An example of this effect is shown in Figure 11.4. Note that the two fluxes are essentially identical except during the period of the transpiration plateau for the dry field. This is a form of transient water stress.

3. The simplest possible function to represent an exponential consistent with a minimum of data is a pair of straight lines which intersect at the point where the exponential increase becomes large. This type of equation is called a 'linear discontinuous model'.

4. Evidence suggests that the threshold leaf water potential may decrease (become more negative) with repeated episodes of soil drying and water stress on the plant. This is a form of adaptation.

Interaction between various parameters and threshold leaf water potential

1) **Soil water content.** Stomatal resistance does not vary with soil water content until the threshold leaf water potential is reached. As the soil dries out (Figure 10.2), the soil water potential diminishes, requiring the leaf water potential to fall to lower and lower values each day in order to sustain the transpiration. After a while, the soil water potential has decreased sufficiently to allow the leaf water potential to reach the threshold, at least for a short time during the day; hence the plateau. Soils that dry out rapidly, such as sand, will experience a rapid decrease in soil water potential and therefore cause the vegetation to reach water stress more quickly. (What do you suppose would occur if you planted a vegetable such as potatoes with a threshold leaf water potential of -10 bar in sandy soil, which has a field capacity of about 10% by volume?)

2) **Xylem resistance.** Making the xylem resistance larger requires the leaf water potential to fall to lower values in order to maintain the same transpiration. Thus, the leaf water potential is more easily reached under moist soil conditions when the xylem resistance is large.

3) **Vapor pressure deficit**. The drier the atmospheric conditions, the greater the vapor pressure deficit, the larger (initially) the transpiration and so, the lower the leaf water potential; the lower

the leaf potential the more likely that it will reach threshold. Since the tripwire in our model resides in the epidermal leaf water potential, the larger the value of beta the more likely the leaf will reach this potential threshold. (Could this response be a self-protective behavior against drought?)

4) **Solar intensity**. Weak sunshine will require small transpiration and so, is not likely to cause the leaf water potential to reach threshold unless the soil is very dry. High solar intensity will tend to force a larger transpiration and so cause a lower leaf water potential to be reached, all other things being equal.

Simulations

Level 1

Here, we will allow you to make up your own mind on the 'root-shoot' controversy by performing a number of sensitivity simulations. Keep an open mind and try to determine as many mechanistic relationships as you can from the output.

Simulation series 1

Vary the soil water content

Run the standard summertime corn case and then lower the root zone soil water content (FSUB in the model input file) enough so that the plateau appears in the transpiration. Note the variation in leaf water potential and stomatal resistance during the period of the plateau, as well as the leaf temperature. Plot soil water content versus stomatal resistance (at midday) for three or four values of root zone soil water content and see if your graph resembles Figure 11.1. (Be careful, if you lower the root zone soil water content too far, you may burn up the plant and cause the model to behave badly.)

Simulation series 2

Vary the xylem resistance, Z_p

Induce the plateau by varying the xylem resistance, Z_p (10th number in the simsphere/data/veglut.dat file; for corn the default is 0.05). What change in Z_p will bring on the plateau earlier, an increase or a decrease? Plot Z_p versus stomatal resistance at midday for three or four different values of Z_p. How sensitive is the leaf water potential, transpiration, and stomatal resistance to this factor? What happens when you lower the xylem resistance?

It is interesting to note that the effect of ozone deposition on plants, ozone being a toxic gas for tissue, is to cause Z_p to increase (see Chapter 13). Given this fact and your results here, consider what effects atmospheric concentration of this toxic gas has on plant growth.

Simulation series 3

Vary the beta parameter

Increase beta, (as in a previous scenario) and pretend that the corn is somehow sensitive to changes in vapor pressure deficit. Report on the behavior of the transpiration plateau and the daily variation of stomatal resistance. Also, look closely at the changes in leaf and epidermal water potentials as beta is varied. Which of these potentials appears to be the 'controller' of stomatal resistance based on these simulations?

Questions

1. Suppose that the continuation of water stress would cause the critical leaf water potential to decrease. What effect would this change have on transpiration?
2. If the xylem resistance (Z_p) increases with plant age, what effect would this have on transpiration?
3. Why do you suppose that the leaf water potential does not decrease very far beyond the critical water potential threshold in any of these simulations?
4. Why do you suppose the transpiration is almost identical in these simulations prior to and after the transpiration plateau?

It should be obvious that there is no clear winner to the 'root-shoot' debate, and that many interactions and influential variables pertain when it comes to stomatal closure and transpiration. The fact that soil water content was not varied in simulations 2 and 3 should tell you that root control is not the sole answer, and that stomates respond to what the leaf status is as well. Just peer out of the closest window and you will see a multitude of variations of plant types and sizes. Thus far, we have considered a full, dense plant canopy with maximum solar radiation. Do you think a plant growing on its own, without the aid of a buffered canopy, might respond differently to the stresses examined in these simulations?

Try setting up a sparse corn canopy (lower the fractional vegetation cover and/or the LAI), submitting it to the changes in VPD, xylem resistance, and soil water content just performed. Does it appear that plants who are 'loners' have extra sensitivity to certain changes than those nestled in a large canopy do? Compare to your results from simulations 1-3 to see which type of plants are better off under certain conditions. Might you have any recommendations for the future evolution of plants who are out on their own to enable them to survive longer?

Slowly decrease soil water content in a series of steps from field capacity to something like 0.3 of field capacity, or at least until the stomatal resistance begins to rise markedly. Plot soil water content versus midday stomatal resistance. Imagine yourself fitting some polynomial function to the curve you just obtained. Argue whether this function would be easier to use in a land surface model to calculate stomatal resistance as a function of soil water content than having to formulate the entire plant hydraulics as in our model.

Definitions

PAR = photosynthetically active (solar) radiation
ABA = abscisic acid
Z_p = xylem resistance

Terms

Terms to look up and remember
transpiration plateau
cytokinins
plant water stress
turgor pressure

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Appendix

We have appreciated for some time how wide a range of environmental factors can elicit a response from a single pair of cells, but to date much research on plant water deficits has concentrated upon characterization of the effects of a deficit-induced increase in a single inhibitor of stomatal opening, namely, ABA. It now seems likely that water deficit may also reduce the concentration in the leaves of promoters of stomatal opening, and thus it is more realistic to expect that the balance of regulators in the leaves will help to determine the degree of stomatal opening. Stomatal behavior in different species under varying environmental conditions must be interpreted in terms of variation in one or more components of an interactive system linking external and endogenous factors. Changes in the levels of single components may or may not elicit a response, and the interactive system operates so that a change in a single component may promote or inhibit, depending on the prevailing conditions.

Simsphere Workbook: Chapter 12

Carbon Dioxide Fluxes in Simsphere

Introduction

Plants survive by taking in carbon dioxide and converting the carbon to its own substance (*assimilation*). The net carbon gain manifests itself as an increase in the biomass, which consists of roots, stem, leaves, flowers, etc. Carbon dioxide (CO_2) enters the plants through the stomates, and so, the rate of biomass increase is closely dependent on the stomatal resistance. Not surprisingly therefore, the economic value of a crop is closely tied to the level of transpiration, which also depends on the stomatal resistance. Since transpiration is not beneficial to the plant except to reduce the leaf temperature[1], we might expect plants to favor a maximization of carbon dioxide intake in relation to transpiration. Thus, plants benefit most by keeping the stomates open, regardless of the transpiration, as long as sufficient water reserves are available to the roots. By now you realize from the simulations that decreasing soil water content does not necessarily reduce transpiration until the plant perceives itself to be in danger of water stress, although the stress signal does not depend uniquely on soil water content.

In these days of the runaway greenhouse effect scare, some researchers take heart that an increase in carbon dioxide concentration in the atmosphere will lead to an enhanced carbon dioxide uptake by the plants and so, to an increased biomass production. Experiments done in the greenhouse and in the field suggest that an increase in carbon dioxide concentration also causes the stomatal resistance to increase, with the net effect being a gain in biomass and a decrease in transpiration, thus doubly benefiting the plant. At first glance, this dual effect of increasing assimilation and reducing water use would seem to be beneficial. We can use our simulation model to explore this finding.

Our main purpose here, however, is simply to examine the flux of carbon dioxide in a canopy (specifically the carbon dioxide assimilation rate $\boldsymbol{A}$). We can also test the idea that an increase in stomatal resistance associated with an increase in ambient carbon dioxide concentration leads to both an increase in the carbon dioxide assimilation rate and possibly to a decrease in transpiration.

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[1] Water transport from root to leaf is nevertheless critical in bringing nutrients and hormones to the plant factory. Transpiration is also necessary to maintain a reasonable leaf temperature. Although photochemical processes tend to be more efficient at higher temperatures, very high temperatures, however, will not only force the plant to lose more water through the leaves but will tend to destroy cellular function.

Calculating the carbon dioxide assimilation rate from the outside in

Let us return to the idea that a flux of a substance moves down a gradient of potential and across a resistance, the Ohm's law analog for diffusive fluxes. The source of carbon dioxide is in the atmosphere, let us say above the plant canopy, where the concentration of carbon dioxide gas ($\boldsymbol{C_a}$) has a mean value of about 350 parts per million (of CO_2) by volume of air (ppmv), which is numerically equivalent to 350 microbars or to 350 mol (CO_2) mol^{-1} (air) times 10^{-6} . We will refer to this ambient carbon dioxide concentration above the canopy as ($\boldsymbol{C}_a$). If the drop in carbon dioxide potential is $\boldsymbol{C}$ and the resistance across that potential drop is r, the flux of carbon dioxide (FCO_2) is given by Eqn. 12.1.

If the plant is to ingest carbon molecules there must be a diffusive flux of CO_2 downward through the surface layer along decreasing concentration to the leaf surface (see Figure 8.1). The appropriate resistances for this path are as follows: 1) Turbulent resistance in the surface layer, which we will call $r_{CO2.}$ 2) Once inside the canopy the molecules move through the interleaf airspaces and across the surface boundary layer of the leaf, where the resistance is $r_{ahc.}$ and 3) Ignoring the flux of carbon dioxide across the cuticular part of the leaf surface, the carbon dioxide molecules then penetrate into the leaf via the stomates where they encounter an internal (or intercellular) carbon dioxide concentration $\boldsymbol{C_i}$, and the stomatal resistance to carbon dioxide flux of r_{sc}. Ci is representative of the concentration of carbon dioxide in the mesophyll waiting to be photosynthesized - a sort of waiting room for carbon atoms in line to be processed. Accordingly, we can write a somewhat more elaborate version of Eqn. 12.1 in the form of Eqn. 12.2 (note that the density of carbon dioxide gas (kg m^{-3} of CO_2) is necessary to make the units agree with the left hand side of the equation, which has the units of kg (CO_2) m^{-2} s^{-1})). This formulation results in what is known as the CO_2 assimilation rate, $\boldsymbol{A}$.

Resistances for carbon dioxide flux are generally somewhat larger than those of water vapor because the molecular diffusivity of carbon dioxide in air is less than that of water vapor in air (possibly because the former is somewhat heavier (molecular weight 44) than water vapor (molecular weight 18)). However, the differences in resistances between carbon dioxide and water vapor in air are generally less than a factor of two (depending on what resistance one is talking about). Accordingly, let us agree for the sake of argument (since it alters no fundamental result) that the two sets of resistances, that for water vapor and that for carbon dioxide, are equal.

Imagine a flux of water vapor from the stomates into the surrounding interleaf airspaces, as in Eqn. 7.3, and thence into the surface layer above the canopy through resistance r_v . Ignoring the parallel water vapor flux from the ground below the canopy, the flux of water vapor between the leaf and the atmospheric surface layer (in kg m^{-2} s^{-1}) is given by Eqn. 12.3. Note that e_a here refers to the vapor pressure above the canopy, *i.e.* at some elevation where the carbon dioxide concentration is not immediately affected by transient perturbations in the canopy fluxes.

Now, if we equate the water vapor and carbon dioxide resistances, and take a ratio of the two fluxes (dividing Eqn. 12.2 by Eqn. 12.3 to yield Eqn. 12.4), we obtain a measure of the *water use efficiency (WUE)*. The latter is essentially the ratio of the carbon dioxide concentration gradient between the atmosphere above the plant canopy and that in the sub stomatal cavities to the gradient in vapor pressure between the inside of the leaf and that in the surface layer above the canopy. Note, however, that because of the 30-fold smaller concentration of carbon dioxide than water vapor in the atmosphere, the magnitude of the water fluxes will be much larger than the fluxes for CO_2. A typical value for the flux of carbon dioxide from air to plant (FCO_2) at noon on a sunny summer day is 1 x10^{-6} kg m^{-2} s^{-1}, or 20 micromoles per square meter per second.

The sub stomatal concentration $\boldsymbol{C}_{\mathrm{i}}$ is approximately constant under normal atmospheric and plant conditions. It is about 220 ppmv for C_3 plants, such as wheat, rice and potatoes, and 120 ppmv for C_4 plants, such as corn and sorghum. $\boldsymbol{C}_{\mathrm{i}}$ can vary slightly with ambient conditions. We will have more to say about this intriguing parameter later in these notes. (The physiological differences between $C_3$and C_4 plants is beyond the scope of this course, except to mention that the latter represents plants that have evolved to have four carbon pathways rather than three pathways in the photosynthesis process as in Δ C_3 plants.)

Eqn. 12.4 shows us that the primary control of water use efficiency is exerted by the vapor pressure deficit between that at the leaf surface and that above the canopy. Assuming that the latter is largely controlled by the atmosphere, the single most important variable in the WUE relationship is the vapor pressure in the leaf, which is to say that control rests with the leaf temperature. We might imagine that the plant is trying to maximize the WUE, but, at the same time, maximize its rate of carbon intake. Blum (1989) cites a formula relating plant yield (YE) to WUE, more specifically the product of WUE times the evapotranspiration. He also cites another formula relating biomass creation to the ratio of transpiration to potential evapotranspiration, which is a little bit like our parameter M, the moisture availability.

Calculating the carbon dioxide assimilation rate from the inside

Well beyond the scope of this course is the frightening terrain of pure plant physiology. Nevertheless, plant physiologists are also struggling with the modelling aspects of assimilation rates. One of the most well known of the current assimilation models is one constructed by Farquhar (1989); of feedforward fame. The Farquhar model, which deals primarily with C_3 plants, attempts to describe the curve shown in Figure 12.1, which emerges from numerous experiments in which $\boldsymbol{C}_{\mathrm{i}}$ is varied as a function of assimilation rate. We see that $\boldsymbol{A}$ increases first rapidly and almost linearly with increasing $\boldsymbol{C}_{\mathrm{i}}$ and then much more slowly beyond a bend in the curve which is actually not far from the characteristic present-day values of $\boldsymbol{C}_{\mathrm{i}}$ for the plant. Assumed in this figure is that $\boldsymbol{Ca}$ is also increased proportionally with $\boldsymbol{C}_{i}$. Typically, C_3 plants tend to have a more gradual transition from rapidly increasing assimilation rate to slowly increasing assimilation rate than C_4 plants, as shown in Figure 12.1. The bend in the two curves occurs close to the present-day normal values for internal carbon dioxide concentrations.

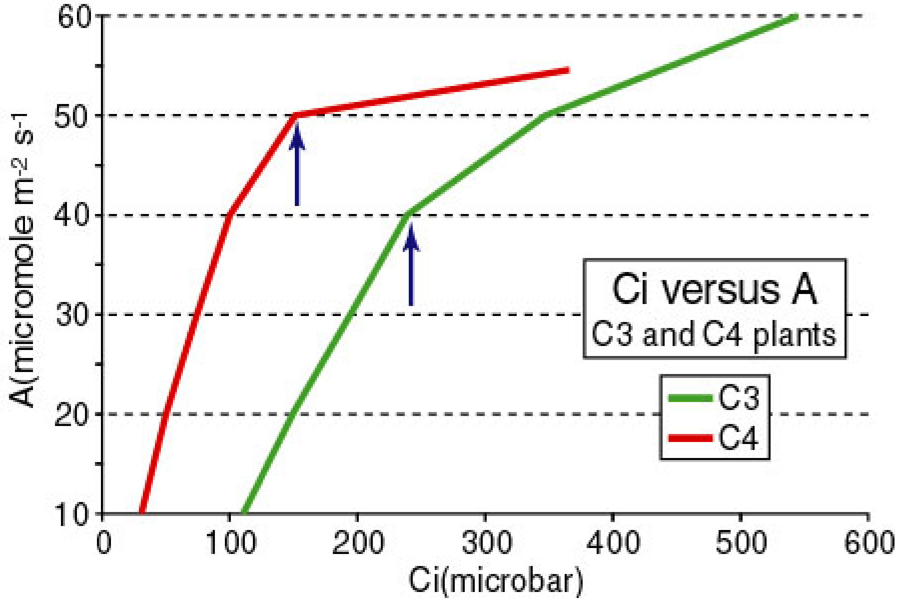

Figure 12.1. Schematic illustration of variation of assimilation rate versus internal carbon dioxide concentration for C3 and C4 plants. Arrows denote present day internal carbon dioxide concentrations.

The bend also represents a transition between two physiological states of the plant, one in which the photosynthesis is limited by the availability of an organic compound called *Rubisco*, which is involved in the reduction and oxidation in the C_3 pathway (low $\boldsymbol{C_i}$), and the other in which photosynthesis is limited by the availability of photon flux (high $\boldsymbol{C_i}$). Clearly, an increase in internal carbon dioxide concentration causes the assimilation rate to increase, although at a rapidly decreasing rate with increasing concentration. We will later touch on the importance of this decrease in assimilation rate with increasing carbon dioxide concentration.

Laboratory measurements show that an increase in carbon dioxide concentration at the surface of the leaf induces an increase in stomatal resistance. A glance at Figure 12.1 agrees with this in that the assimilation rate does not increase rapidly with an increase in internal carbon dioxide concentration beyond present-day concentrations. Experiments further show that while fluctuations in stomatal resistance and other local factors do not significantly affect internal carbon dioxide concentration, an increase in ambient carbon dioxide concentration moves the entire curves for both C_3 and C_4 plants, shown in Figure 12.1, toward the right. Despite this shift, the net effect of increasing C_i is one of an increase in assimilation rate even for C_4 plants. Note that the sharper transition at the bend in the curves followed by a nearly constant value of assimilation rate of the C_4 curve above the bend, thereby translating to a smaller gain in $\boldsymbol{A}$ for C_4 plants than for C_3 plants with an increase in $\boldsymbol{C}_a$ or $\boldsymbol{C}_i$.

Changes in assimilation rate and transpiration with increasing ambient carbon dioxide concentration

Worst-case scenarios (not one of ours) suggests a near doubling of ambient carbon dioxide concentration to 660 ppmv before the end of the 21^{st} century as the result of continued fossil fuel burning. This increase already comes on top of an increase from 280 ppmv during the middle of the 19^{th} century to about 360 ppmv during the 1990s, and 410 ppmv in the early 21^{st} century. A first guess based on Eqn. 12.2 is that the doubling in $\boldsymbol{C}_a$ would cause the assimilation rate $\boldsymbol{A}$ to increase by a factor of about 4 for C_3 plants and about 2.5 for C_4 plants (assuming no change in the values for $\boldsymbol{C}_i$). In fact, Cure and Acock (1986) examined all the published measurements they could find related to the response of plants to an increase of carbon dioxide. Their results show that the increase in assimilation rate is likely to be only about 40% for C_3 plants and about 25% for C_4 plants. Moreover, they show that plants grown under ambient concentrations of 660 ppmv, or allowed to come into equilibrium with their new enriched CO_2 environment, show an even lower increase in assimilation rate, about 30% for C_3 plants and less than 10% for C_4 plants. These increases in assimilation rates translate into increases in biomass production.

A further implication of Cure and Acock's (1986) data is that transpiration should decrease by about 20% as the result of carbon dioxide doubling! What Eqn. 12.1 shows clearly is that an increase of 20% in stomatal resistance is not sufficiently large to hold the assimilation rate to only a 40% increase; rather, both $\boldsymbol{C}_i$ and stomatal resistance must increase as ambient carbon dioxide concentration is increased in order for this to hold true. This increase in stomatal resistance seems to be associated with the leaf's ability to sense an increase in carbon dioxide concentration at its surface[2]. Recent studies have shown that an increase of carbon dioxide concentration by a factor of two may produce only a 5% decrease in transpiration and only a modest increase in biomass! This is tantamount to saying that a huge increase in ambient carbon concentration has little effect on the plants. How discouraging to learn that this beneficial side effect of fossil fuel burning will not have such a profound beneficial effect after all.

Alternately stated, if one doubles the amount of food on the table (ambient carbon dioxide concentration), people will stuff their mouths more (internal carbon dioxide concentration), but they will not ingest twice as much food. Moreover, given some time to equilibrate, people may get sufficiently fed up (literally) that they will become more resistant to temptation and not ingest much more food than before, although it is certainly true that the more food available the more one eats (up to a point). Thus, stomatal resistance increases in response to the plant's inability to assimilate all that is put on its plate, given the amount of available sunshine and nutrients for carrying out all its chemical reactions. Ultimately, the availability of sunlight will restrict the increase in biomass. (Speaking of resistance, anyone who has ever tried to feed an infant would know what happens when you try to increase the food intake rate by increasing the mass of goop on the end of a spoon! You do get more inside the infant, but a lot of resistance is put forth and a lot of goop ends up on the walls).

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[2] Measurements by Kell Wilson at the USDA field site in Beltsville, MD, of soybeans and corn indicate that the increase in stomatal resistance due to a doubling of carbon dioxide concentration is larger than reported by Cure and Acock, perhaps as much as 40 or 50%.

Simulations

Level 1

The model you have come to know and love calculates carbon dioxide flux and disgorges it in units of kg m^{-2} s^{-1}. Let us review a couple principles of scaling from a leaf to a canopy. As with water vapor fluxes, the calculations refer to flux per unit sunlit leaf area, but the output is in terms of flux per unit horizontal surface area. As in previous simulations, we calculate fluxes for each leaf or leaf strata and divide the leaf resistances by the leaf area index multiplied by a *shelter factor*. The reason why we divide by the leaf area index is that we must sum up all the individual leaf fluxes for one-sided transpiration. As pointed out previously, were one to divide resistances by LAI, (nearly equivalent to multiplying the transpiration fluxes by LAI) the resultant fluxes would generally be too large. The transpiring area would be overestimated in that case, since many leaves are shaded by other leaves and thus have a larger stomatal resistance. Accordingly, we use an equation that reduces the leaf resistances by an amount that varies between about 1.0 for a low leaf area index to about 2.0 for large leaf area indices. Both the carbon dioxide and water vapor fluxes are scaled in this way. You may have noticed that the minimum stomatal resistance assigned to our corn canopy is 50 sm^{-1}, yet much of the output of r_s looked at so far shows the stomatal resistance at values below 50. This is due to the shelter factor calculation.

Simulation series 1

Induce water stress on the corn canopy

Using a large leaf area index (LAI = 7), re-run the standard corn simulation, but this time examine the carbon dioxide fluxes and the water vapor fluxes together, as well as the Water Use Efficiency (WUE, which the model calculates for you as WATEFF in the output data). Then, run a simulation in which water stress manifests itself as a plateau in the evapotranspiration and note the changes in WUE, carbon dioxide fluxes and transpiration during the day from the unstressed run. Transpiration is approximately equal to the latent heat flux (LE in the model output) for a dense canopy (say, XLAI=7). Look back at the previous chapter for ways to simulate the transpiration plateau.

Questions

1. What is the most noticeable change in the fluxes from the unstressed to the stressed simulations?
2. What does this tell you about the behavior of the stomatal resistance and its impact on Eqn. 12.4 (although it is not explicitly included)?
3. Under these conditions, how well did the plant perform in maximizing its goal (increased biomass) with the onset of stress?
4. What are the differences in the character of the stress response of assimilation rate (which is the carbon dioxide flux, CO2F in the output) to transpiration?

Simulation series 2

Increase the xylem resistance

Xylem resistance does not seem to change with time very rapidly in response to stress, though it does exhibit a decrease and then an increase as the plant matures and then ages. Run a simulation after doubling the xylem resistance and examine the carbon dioxide fluxes, WUE and transpiration once again. Changes in the xylem resistance, called Z_p in the model, occur during the life of a plant, and we have seen some of its effects in previous chapters. It probably decreases in the early stages of plant life and then increases again with time as the plant matures and then senesces.

Questions

1. Given that we are stressing the plant once again, how do its responses with regard to WUE and carbon fluxes compare with those under the water stress imparted in simulation 1?
2. Are they more or less drastic? Once again, try and figure out why the WUE changes the way it does, i.e. which variables in Eqn. 12.4 are different and why?
3. How could one tell from examination of the assimilation and transpiration curves whether stress was coming from a lack of water or something that is causing the xylem resistance to increase?

Level 2

Simulation series 3

Pollute the ambient atmosphere; a vision of the future?

Check out the carbon dioxide doubling issue referred to by Cure and Acock (1986). First, double the ambient carbon dioxide concentration (CO in the model input file) to 660 ppmv, as many prognosticators claim it will be the end of the 21st century. Note the increase in the fluxes of carbon dioxide and the WUE from the base case and see if it is similar to the 30% increase indicated by the results of Cure and Acock. It isn't! How much more or less than 30% are the changes that you see?

So then, increase the minimum stomatal resistance by 40% (second column (first number) in ‘simsphere/data/veglut.dat’ file; for corn, increase it from 50 to 70) based on Wilson's field measurements for doubled carbon dioxide, and see if you reduce the carbon dioxide fluxes to the 30% level predicted. You can't, unless you also increase the substomatal concentration, $\boldsymbol{C}_i$, which you should try next. Once you have arrived at a new assimilation rate $\boldsymbol{A}$ (CO2 flux to the plant), increase the leaf area by the same fractional increase in $\boldsymbol{A}$ in order to simulate the increase in biomass resulting from this increase in carbon intake. Then, redo your simulations until you have iterated on a solution satisfying Wilson's stomatal resistance increases of 40% under carbon dioxide doubling. When you finish with these simulations you will discover a surprising fact: that the transpiration decrease is not all it’s cracked up to be!

Questions

1. At what values of r_{smin} and C_I did you finally come up with the 30% increase in CO_2 flux?
2. Are these values reasonable, physically speaking?
3. (At this point, look at the decrease in transpiration from the base case.) Does that value agree with the 20% decrease anticipated by the results of Cure and Acock? If not, is the model full of baloney, or are Cure and Acock out to lunch?
4. Field experiments and simulations with the Simsphere model, as opposed to laboratory measurements that make up most of Cure and Acock's data, suggest decreases in transpiration of only a few percent, rather than the 20% found by Cure and Acock, in response to a doubling of carbon dioxide. Suggest some additional feedback mechanisms that might exist when an entire canopy and an atmospheric boundary layer are considered.

Look at the equations presented in this chapter for more insight, as they are responsible for all the output and process we are considering here.

Regardless of the actual amount of decrease in transpiration, how do you think plants as a whole would change if the CO_2 concentration does double in the next 100 years, based on these simulations? Doesn't this seem like the best of both worlds for vegetation: increasing their WUE and biomass while at the same time retaining more water? Are there any downsides you can think of?

Equations

[Eqn 12.1]

$$FCO_2 = \frac{\Delta C}{r}$$

[Eqn 12.2]

$$FCO_2 = \frac{[C_a - C_i]\,\rho_{CO2}}{r_{CO2} + r_{ahc} + r_{sc}} = A$$

[Eqn 12.3]

$$FH_2O = \frac{[e_s(T_l) - e_a]}{(r_v + r_{ah} + r_s)\,\gamma L_e}\,\rho C_p = T_r$$

[Eqn 12.4]

$$\frac{A}{T_r} = WUE = \frac{\rho_{CO_2}}{\gamma L_e}\left[\frac{C_a - C_i}{e_s(T_l) - e_a}\right]$$

Definitions

A = Carbon dioxide assimilation rate
(C_a) = Concentration of carbon dioxide gas
(r_{CO2}) = Turbulent resistance in the surface layer
(C_i) = Internal (or intercellular) carbon dioxide concentration

(e_a) = Vapor pressure above the canopy
(WUE) = Water use efficiency
($\boldsymbol{C}_i$) = Sub stomatal concentration
Rubisco = Is involved in the reduction and oxidation in the C_3 pathway

Terms

Terms to look up and remember
Biomass
Rubisco
Water use efficiency

References

- Blum, A., 1989, Breeding methods for drought resistance, **Plants Under Stress**, Jones, Flowers and Jones, Eds., Cambridge University Press, 257pp.
- Farquhar, G. D., 1989, Models of integrated photosynthesis of cells and leaves, **Phil. Trans. Roy. Soc. London**, B., 323, 357-367.
- Cure, J. D. and B. Acock, 1986, Crop responses to carbon dioxide doubling: a literature survey, **Ag. and Forest Meteor.**, 38, 127-145.

Simsphere Workbook: Chapter 13

Ozone Fluxes in Simsphere

Introduction

Like carbon dioxide, plants ingest other gases, some of which are harmful. One of these harmful gases is ozone, which is produced by photon flux interaction with pollutants, mainly auto emissions. Although the mechanism by which ozone destroys plant tissue is conjectural, its presence has undeniable effects in damaging plant leaves. As such, ozone looms as an important economic factor in biomass production, and studies are underway to determine the effect of ozone on both corn and soybeans. The molecular weight of ozone (O_3) is 48 grams per mole. Not only is the molecular weight of ozone similar to that of carbon dioxide (44 g), but the molecular diffusivity in air and its density is also very similar. Thus, in principle, it is a simple matter to calculate ozone flux to the mesophyll using an analogous equation to 12.2, but with the ambient concentration of ozone replacing that of carbon dioxide, and the internal contact concentration of ozone as zero. Typical ambient ozone concentrations vary between about 0.04 ppmv for a relatively unpolluted atmosphere to about 0.12 ppmv for a polluted atmosphere (Los Angeles!). Thus, ozone concentrations are more than a thousand times smaller than that of carbon dioxide.

Unlike the carbon dioxide fluxes, however, ozone has a sink at the ground surface as well as in the leaves. Since the cuticular resistance is typically many times that of the stomatal resistance, most of the ozone ingested by the leaf is through the stomates. However, an approximately equal flux enters the ground surface in and around a plant canopy. Typical fluxes to the leaves are between 0.3 and 0.8 micrograms per square meter per second for a LAI of 3. Maximum fluxes tend to occur during the afternoon in association with the reduced stomatal and air resistances. Recent studies show that the effect of ozone on the leaves is to increase the plant stem/root hydraulic resistance, which results in an increase in stomatal resistance as well. Accordingly, not only does the ozone destroy leaf tissue, but also the plant restricts the intake of carbon dioxide, which is used to make plant material. Perhaps the plant is trying to protect itself by also restricting the intake of ozone. This might compensate somewhat for the increased intake of ozone and its attendant destruction of leaf material. Obviously, in view of the damage inflicted by ozone on crops, this buffering strategy by the plant is only partially successful.

Although ozone (O_3) is essential for filtering out harmful ultraviolet rays at 20 km above the earth's surface, where the gas is produced by the action of photons on oxygen molecules, it is a harmful gas if absorbed by living things. Ozone is also produced near the surface by the action of photons on the by-products of burnt fossil fuels. While this problem is most acute in places like California, ozone levels in State College do occasionally reach or exceed the maximum recommended level of safety recently specified by the EPA -- 80 parts per billion by volume (0.08 ppmv). Daily variations in ozone concentration depend on the solar insolation, turbulent mixing of the atmosphere, and the level of traffic. Figure 13.1 shows a typical example of the daytime variation in ozone concentration at the top of Walker Bldg at Penn State during a week in early summer in 1995.

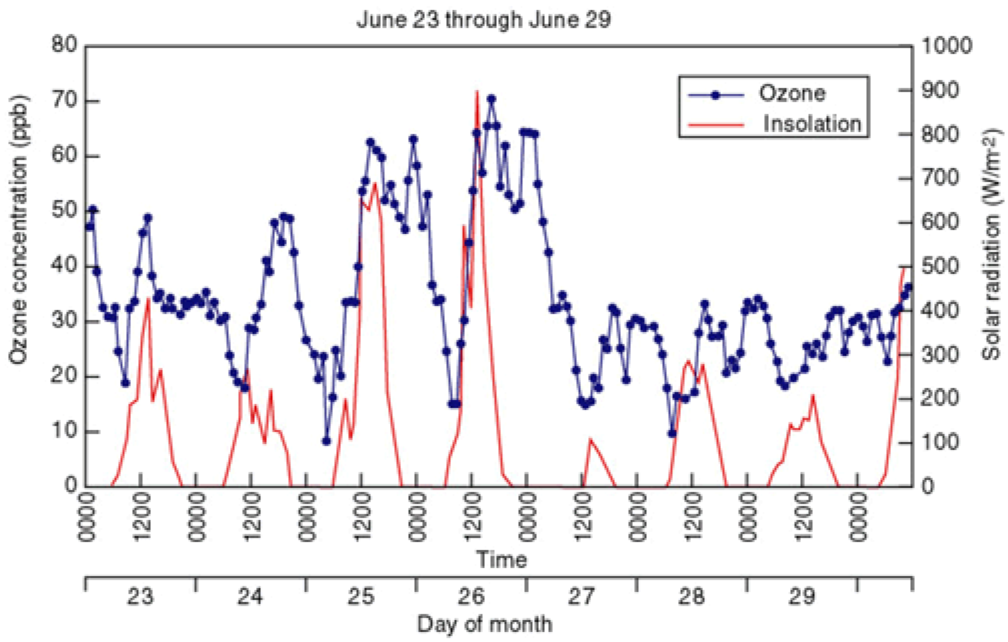

Figure 13.1. Ozone conentration and solar radiation versus time.

A large range of sensitivity to ozone exists in commercial plants and trees. Black Cherry is known to be sensitive to ozone, white ash is somewhat sensitive and red maple is tolerant. Studies on commercially grown plants in California, such as cotton and grapes, confirm that ozone has a deleterious effect on plant growth and biomass production. Evidence suggests that the adverse effects of ozone result from a reduction in root mass in relation to leaf area, accompanied by a reduction of stomatal conductance and intake of carbon. Yields of cotton in the San Joaquin Valley of California decrease by about 20% from exposure to ozone produced near the surface. Indeed, much larger declines in plant productivity have been observed during exposure to ozone by wheat and pea plants and in oak trees.

Exposure to ozone is known to increase stomatal resistance (decrease stomatal conductance), with response times for the stomates of a few hours or less. This response is especially rapid in older leaves, although younger ones may actually increase their stomatal conductance for a short time in response to the failure of the older leaves to photosynthesize. It follows that a greater ingestion of ozone will occur, all other things being equal, in dryer soils in which the plants may exhibit a higher stomatal resistance than in the wet soils. [1]

Intake of ozone into plants occurs primarily through the stomata and an overall increase in stomatal resistance occurs over the whole plant when exposed to ozone. This increase in stomatal resistance in response to the ozone intake is not sufficient to eliminate the destruction of leaf material, since leaves remain exposed to ingested ozone. Moreover, the effect of ozone on the more sensitive leaves remains for many hours or days after the exposure is removed. Reduced intake of both carbon dioxide and ozone and the destruction of leaf material constitute separate but equally detrimental effects of ozone exposure to plants. The effect of ozone exposure on the root growth is slower and at longer term.

The question has been asked: what are the mechanisms by which ozone affects plant well-being? Evidence suggests that the intake of ozone through the stomates is the primary mechanism for the limitation of carbon intake. Nevertheless, intake of ozone through the cuticle can result in ozone reaction with plant material. Indeed, it appears that the destruction of ozone (and its return to oxygen) occurs when the latter strikes reactive surfaces including water. Thus, ozone destruction can occur at the ground in and around plants, although a ground sink of ozone may depend on its wetness. There is evidence to suggest that ozone is blocked from entering the stomates when the leaves are wet, such as during the morning when dew lies on the plants.

1.) The reverse has been found in some experiments with Black Cherry, although this completely unexpected result has not yet been explained.

Ozone effects on hydraulic and stomatal resistance

A strong positive correlation has been demonstrated between canopy stomatal resistance (r_{st}) and hydraulic resistance (Z_p) of the root-stem complex in cotton plants. Normally, Z_p (also referred to as the xylem resistance) varies with the age and type of the plant, but it also varies with the general health of the plant. For ozone, laboratory experiments show that stomatal conductance and hydraulic conductance diminish almost linearly with ozone concentration for 12-hour exposures in cotton plants. These same studies also present the puzzling result that the mid day (minimum) leaf water potential remains almost constant (typically at about -14 bar) with varying ozone concentration. When plotted one against the other, hydraulic conductance varies almost linearly with stomatal conductance.

Stomatal conductance (or resistance) varies considerably over the course of a single day, and even from day to day depending on the amount of insolation. Yet hydraulic conductance is found to vary slowly with time; it may, for example, gradually diminish over the course of a week or two as the plant undergoes senescence or starts to suffer from water stress. Clearly, stomatal resistance depends on external and internal factors, among them hydraulic conductivity. Thus, if the latter is varied, stomatal resistance must adjust.

Recalling the equations governing the fluxes of water and water vapor, one can describe, by combining the transpiration equation (12.3) and the hydraulic conductivity equation (10.3), the *steady state* flow of water uptake and transpiration by a plant. In this case, it can be shown mathematically that stomatal resistance is dependent on the choice of the plant hydraulic resistance (Z_p) (or its inverse, the plant hydraulic conductance (G_p)). Using the aforementioned equation, which we have seen is contained within the model, the calculated stomatal resistance varies almost linearly with changing Z_p, fitting the observations almost exactly. Since Z_p increases in direct proportion to ozone concentration over a wide range of concentrations, one can deduce that stomatal resistance will increase in direct proportion to the ozone concentration.

As we have seen in the previous chapter, a doubling of r_{st} will not necessarily halve the carbon assimilation rate, ***A***, unless the other resistances in the denominator of the resistance equation were zero (which they are not!). Regardless of this mechanistic relationship, the close correspondence under many conditions between xylem hydraulic conductance and gas exchange performance suggests that alteration of root properties by ozone exposure are likely to exert harmful effects on shoot (leaf) and canopy gas exchange. By reducing the leaf area, an increase in ozone implicitly affects Z_p. In real plants, the reduction in leaf area due to ozone exposure is due to more leaves remaining on the plant than can be supported by the roots. These leaves suffer both the chemical poisoning by ozone but also a desiccation due to their inability to sustain transpiration. Thus, the leaves shrivel, die and fall off. One could argue that the reduction in root conductance, which accompanies ozone exposure as well, is a protective mechanism to reduce ozone intake by increasing stomatal resistance. The net result is a reduction in biomass not unlike that produced by water stress on the plant.

Changes in leaf water potential due to ozone

A curious observation made by Grantz (Grantz et al., 1999) is the lack of response by the leaf water potential to changing stomatal resistance induced by variations in ambient ozone concentration. A clue to this behavior is found in the values of leaf water potential, Ψ, which in the cotton experiments were typically about -14 bar. An increase in Z will require that, to maintain constant transpiration, Ψ must decrease still further. If the latter is very close to the critical water potential, however, a slight decrease in Ψ will result in a large increase in r_{st}. Thus, an increase in Z causes the stomatal resistance to increase by triggering a response in the leaves via the critical water potential mechanism discussed in previous chapters. Such a slight decrease in Ψ would not be readily measurable within the accuracy of current techniques (~˜0.5 - 1,0 bar), which is perhaps one reason why a few plant scientists claim that Ψ is non sensitive to water stress or ozone concentration. If this hypothesis that the mechanism operates through the critical water potential is correct, the effects of stomatal resistance increase due to increased ozone concentration should be much more apparent at mid day than in the early or mid morning.

Simulations

Level 1

We assume that the ozone concentration at the interior of the leaf is zero (unlike that of the internal carbon dioxide concentration, Ci, in plant leaves), and that all the ozone has broken down into oxygen and has bonded in unwanted fashion with the plant material. We set in the model the ambient ozone concentration at a level z_a (50 m) to C_{oz}; at which the ozone concentrations are assumed independent of the vegetation.

Values measured by the EPA would be at screen level (1.3 m). These will be somewhat lower than the ambient values at z_a. In reality, the near-surface concentrations will increase during the morning and decrease during the afternoon, in response to increased and then decreased photochemical reaction with fossil fuel products. We further assume that the flux of ozone from the air to the stomates occurs in parallel with that through the cuticle, and that the total leaf flux is in parallel with that through the cuticle and with that into the ground, as in the big leaf model described in chapter 7. Thus, we have three streams of ozone: one into the stomates, one into the cuticle, and the other into the ground. Diffusive resistances for the ground and the leaf surfaces are prescribed, respectively, by the physics of resistance for a flat surface and that for the stomates by a simple diffusive flux. We assume further that all the ozone reaching the ground is absorbed (probably too extreme an assumption), that the stomates are one-sided (although cuticles exist on both sides) and that a shelter factor P is necessary for the stomatal resistance (as explained previously).

Simulation series 1

- **Observe the diurnal behavior of near-surface ozone**

Set the value of ambient ozone concentration (COZ_AIR in the model input file) to 0.08 ppmv, the recommended maximum concentration by the EPA. Run the model with the standard set of values you have used in previous simulations for corn. Graph the values of ozone concentration at screen level (1.5 m) (canopy ozone concentration, O3CAN in the model output).

Questions

1. At what time of day is the maximum ozone concentration and why?
2. Why does the concentration of ozone (O3CAN in the model output) not reach 0.08 ppmv?
3. Compare the simulated results with the actual values for State College shown in Figure 13.1. What are some of the reasons for the much more irregular pattern of changes observed in the latter as compared with the simulations?

Simulation series 2

Vary the soil moisture, LAI, and wind speed

Vary the root zone soil moisture availability, the vegetation amount (LAI), and the wind speed as you did in the earlier chapters. Now observe the differences in ozone concentrations between these simulations.

Questions

1. Describe the differences in ozone concentration and flux between the high and low moisture availability, high and low vegetation amounts and high and low wind speeds and try to explain these differences. Recall from earlier chapters how the fluxes and resistances respond to these changes to help you explain the ozone response.

2. Under what atmospheric conditions is the plant/canopy better off as far as minimizing its exposure and intake of ozone? Do you think that auto humidification might help the plants in this situation as well?

Level 2

Simulation series 3

Below is a table obtained from measurements made in the field over two types of cotton plants.

g_{st} stomatal conductance mol m^{-2} s^{-1}	**G_p hydraulic conductance** mmol m^{-2} MPa^{-1} s^{-1}
0.4	4.0
0.39	4.15
0.53	4.8
0.525	4.8
0.39	5.6
0.58	6.0
0.59	6.0
0.65	6.9
0.8	7.1
0.83	7.0
0.75	8.4
0.8	8.9
0.92	9.2
0.86	9.0

Table 1. Stomatal conductance versus hydraulic conductance as measured in cotton at the University of California, Kearney Agricultural Center, Riverside California (Grantz et al., 1999).

Convert the values as given to resistance with units of bar $(Wm^{-2})^{-1}$ for hydraulic resistance and units of s m^{-1} for stomatal resistance. To do this first take the inverse of the values given in the table to get the resistances. Then, divide the hydraulic resistance in the inverse units given above by 4.45 and the stomatal resistance in the units given above by 0.025. Thus, a value of 10 units in hydraulic conductivity becomes 0.1 units in hydraulic resistance and 0.1/4.45 (or about 0.02) in units of bar $(Wm^{-2})^{-1}$. Similarly, a stomatal conductance of 0.5 units in the above notation becomes 2 units of resistance in the above notation and 2/0.025 (or 80) in units of s m^{-1}.

Now, see if you can reproduce these results using our default sounding for mid-latitude summer. To do this, vary the hydraulic (xylem) resistance between 0.025 and 0.055 bar $(Wm^{-2})^{-1}$ at intervals of 0.010 [i.e., .025 .035 .045 .055]. Assign a critical water potential of -30 bar (essentially eliminating this variable from play) and note the midday stomatal resistance for each. It should be obvious that such a low critical leaf water potential would not be reached and, therefore, stomatal resistance does not change. Now, lower the critical leaf water potential at your own discretion, and try to duplicate the results in the table above. Do not expect to match the values exactly, but try to get a similar linear agreement between the hydraulic and stomatal resistances. (Hint: you may begin to see agreement with the observations when the critical water potential is between -15 and -20 bar.) Graph the leaf water potential versus time for each when you get agreement with the table.

Questions

1. Can you reproduce the linear relationship between stomatal and hydraulic resistance?
2. What role does the critical water potential seem to play in producing this relationship? Can you explain the reason for this?
3. Noting the variation in leaf water potential when you obtain the linear agreement, why does the latter not seem to vary much? If you were an experimenter and your error bars were 0.5-1.0 bar in measuring the leaf water potential, what might you incorrectly conclude about the sensitivity of stomatal resistance to leaf water potential?

Simulation series 4

Return to your simulations in which you induced a midday 'stomatal closure' (a significant elevation of stomatal resistance at mid day due to reduced root zone soil water content). Rerun these simulations with two differing input values of Z (one high and one low) to simulate the effects of a high and low ozone concentration; (choose these two values as representing the extreme values in the above table). Choose a high and a low ozone concentration (say 0.12 and 0.04 ppmv) for these two cases. Now assess whether the plant will be better off in terms of water use efficiency, assimilation rate and ozone intake if it were shaded as if the sky were completely overcast. To assess the effect of shading, run another set of simulations with the same parameters but with a cloud factor representative of an overcast sky. The point here is that shading will increase stomatal resistance by reducing photon intake and therefore reduce photosynthesis, but, at the same time, it will reduce the transpiration and therefore reduce the water stress on the plant. Be sure to compare the stomatal resistance and leaf water potential values at mid day for the unshaded and shaded series of simulations.

Once you have done this, vary the root zone soil water content to find the conditions under which shading of the plant during dry periods of high ozone concentration may improve the productivity. If this works out you might want to try varying Z to simulate a variation in ozone concentration and find the range of soil water content an ozone concentrations for which shading will benefit the plant.

Definitions

Stomatal resistance for the canopy = r_{st}
Xylem resistance = Z_p
Plant Hydraulic Conductance = G_p
Internal Carbon Dioxide Concentration = C_i
Leaf Water Potential = Ψ

References

- Grantz, D. A, X Zhang and T. Carlson, 1999: Observations and model simulations link stomatal inhibitions to impaired hydraulic conductance following ozone exposure in cotton. *Plant, Cell and Environment.* (in press).